Geochronology, geochemistry and tectonic significance of the late Mesozoic volcanic sequences in the northern Wuyi Mountain volcanic belt of South China

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ABSTRACT

The widespread occurrence of late Mesozoic volcanic rocks in the Gan-Hang Belt in South China is associated with similarly widespread mineralization, but many important questions surrounding these volcanic rocks have not been clearly answered. The Tianhuashan basin located in the northern Wuyi Mountain volcanic belt is one of the most important volcanic basins in the Gan-Hang Belt, and it is primarily composed of the Daguding and Ehuling Formations and their intrusive counterparts. LA-ICP-MS zircon U–Pb dating shows that the Daguding Formation erupted in the Late Jurassic (152–160 Ma), whereas the Ehuling Formation erupted in the Early Cretaceous (131–139 Ma) in the Tianhuashan basin. Volcanic rocks are rhyolite and share similar trace and rare earth element patterns with an enrichment of LREEs and a depletion in Sr, Ba, Nb, Ta, P, Eu and Ti. They are also characterized by negative whole rock $\epsilon_{\text{Nd}}(t)$ and zircon $\epsilon_{\text{Hf}}(t)$ values with Paleoproterozoic $t_{2\text{DM}}$ ages, suggesting that they were derived primarily from the remelting of ancient crustal materials. Daguding volcanic rocks are strongly peraluminous and show a higher Mg# than pure crustal melts, implying that they were likely derived from Paleoproterozoic metasedimentary basement materials. However, Ehuling volcanic rocks are weakly peraluminous and have a pronounced A₂-type geochemical signature. Detailed elemental and isotopic data suggest that they were formed by the partial melting of the Paleoproterozoic metamorphic basement (including metasedimentary and metaigneous rocks) at a high temperature (~840°C), followed by fractional crystallization. These results imply that during the Late Jurassic, South China on the Gan-Hang Belt was a continental arc coupled with the subduction of the Paleo-Pacific plate. Since the beginning of the Early Cretaceous, an intra-arc rift has formed along the Gan-Hang Belt as a consequence of slab rollback. These results also indicate that the extension in the Gan-Hang Belt began later than the southwestern part of the Shi-Hang Zone and lasted from

139 Ma to 122 Ma.

Keywords:

Late Mesozoic volcanism; Petrogenesis; Tectonic setting; Northern Wuyi Mountain volcanic belt; South China

1. Introduction

During the Late Mesozoic period, magmatic rocks were extensively developed in South China, most of which were granites and their volcanic counterparts (Zhou et al., 2006). Their ages, petrogenesis, tectonic settings and geodynamics have been the focus of many studies over the last few decades (Chen and Jahn, 1998; He et al., 2010; Hsü et al., 1988; Hsü et al., 1990; Jiang et al., 2006; Jiang et al., 2008; Jiang et al., 2009; Jiang et al., 2011; Lapierre et al., 1997; Li and Li, 2007; Li et al., 2014; Qi et al., 2013; Wang et al., 2013; Wong et al., 2009; Wong et al., 2011; Xu, 2008; Yang et al., 2012; Yu et al., 2006; Zhou and Li, 2000; Zhou et al., 2006). Although volcanic rocks are well developed and widely distributed in South China, as an important part of the coastal area within China and even the circum-Pacific volcanic belt, they have received much less attention than granites (e.g., Wang et al., 2013; Guo et al., 2012; Jiang et al., 2005; Su et al., 2014; Yang et al., 2011; Yang et al., 2013). However, these volcanic rocks contain important information about the chronologic framework and petrogenesis of multistage magmatism and could shed new light on the detailed geodynamic evolution and tectonic regimes in this region.

The northern Wuyi Mountain volcanic belt is an important volcanic rock belt in South China, which is controlled by basement uplift from the northeast towards the southwest. A series of volcanic basins has developed along this belt (such as Tianhuashan, Huanggangshan, and Tongboshan), accompanied by abundant magmatism-related polymetallic mineral deposits (Huang et al., 2012; Meng et al., 2012; Zhang et al., 2009; Zhang et al., 2012). As an important component of this belt, the Lengshuikeng ore field in the Tianhuashan volcanic basin is a super large Ag–Pb–Zn polymetallic mineral deposit, and it has attracted considerable attention from many geologists worldwide. It is generally believed that mineralization is closely related to late Mesozoic magmatic activity. However, previous studies have focused primarily on the geochronological aspects of the ore-related granitoid rock (Huang, 1993; Meng et al., 2009; Meng et al., 2012; Qiu et al., 2013; Su et al., 2014; Wang et al., 2013; Xu et al., 2002; Zuo et al., 2008; Zuo et al., 2010). Because the host rock is Mesozoic volcanic rock, the precise and accurate isotopic age determination of the rocks is crucial not only for the geologic time of Late Mesozoic volcanism and associated tectonic events but also for the origin of the Lengshuikeng Ag–Pb–Zn polymetallic mineral deposit. Furthermore, the origin of these volcanic rocks and the processes involved in their formation provide an opportunity to reveal the late Mesozoic tectonic evolution of the northern Wuyi Mountain metallogenic belt. It is thus apparent that the petrogenetic characteristics of volcanic rocks need to be studied in much greater detail to further understand the Late Mesozoic magmatic activity and the process of crust-mantle interaction in the northern Wuyi Mountain area.

In this study, two drill holes accompanied by several outcrops, which contain both of the volcanic formations (Daguding Fm. and Ehuling Fm.), were selected as the most suitable sites for a comprehensive study of the volcanism in the Tianhuashan volcanic basin. We present precise zircon U–Pb ages, along with major and trace elements and Sr–Nd–Hf isotopic compositions for volcanic rocks, in an attempt to better constrain the ages and petrogenetic processes involved and to shed new light on the geodynamic setting of Late Mesozoic tectonics and magmatism in the northern Wuyi Mountain area.

2. Geological background

The South China Block has a complex tectonic evolution history, and it is composed of two major Precambrian continental blocks: the Yangtze Block in the northwest and the Cathaysia Block in the southeast. The two blocks are divided by the Gan–Hang tectonic belt (Gilder et al., 1996; Yu et al., 2006), a tectonic suture zone between the Yangtze and Cathaysia blocks, which starts from Shaoxing, Zhejiang Province in the northeast and extends to Yongfeng, Jiangxi Province in the southwest (Fig. 1). The Late Mesozoic period is characterized by the multistage eruption of volcanic rocks and their intrusive counterparts in South China, resulting in the extensive emplacement of a series of Late Mesozoic volcanic-intrusive complexes in South China, which have been conventionally divided into three major stages from the early to late Yanshanian: 180–160 Ma, 160–135 Ma and 135–90 Ma. Among these, felsic rocks are volumetrically dominant, whereas mafic rocks (e.g., basalts, gabbros, andesites and syenites) are relatively rare (Li and Li, 2007; Li, 2000; Zhou and Li, 2000).

The Tianhuashan basin, which contains a super large Ag–Pb–Zn polymetallic mineral deposit (Lengshuikeng deposit) in the northwestern margin, is geographically located on the border of the Jiangxi and Fujian provinces. Tectonically, it is located in the northeast corner of the Cathaysia Block and is bounded by the Jiangshan–Shaoxing suture zone to the north and the Yingtan–Anyuan Fault to the west. It is a member of the northern Wuyi Mountain volcanic belt adjacent to the southern margin of the Gan–Hang tectonic belt. This belt is composed of a number of NNE-trending Mesozoic extensional volcanic basins, such as the Tianhuashan, Huanggangshan and Tongboshan basins (Fig. 1). The Tianhuashan basin shows an ellipsoidal shape and is approximately 50 km long and 40 km wide, covering an area of approximately 1500 km². It is one of the largest volcanic basins in the region and is composed of the Tiantaishan and Yuefengshan sub-basins.

The stratigraphic sequence in the basin is summarized in Table 1. The basement of the Tianhuashan basin consists of a metamorphic complex including biotite quartz schists, quartz-biotite schists and biotite plagioclase gneisses of the Neoproterozoic Sinian period. The Paleozoic strata are sporadically distributed at the margins of the basin with a carboniferous system that primarily consists of quartz conglomerate and sandstone of the Zishan Fm. at the bottom and limestone of the Huanglong Fm. at the top. The Huanglong Fm. is discordantly overlain by the late Mesozoic volcanic rocks, which can be divided into two magmatic cycles: the early stage Daguding Fm. and the late-stage Ehuling Fm. The Daguding Fm. volcanics are composed of two distinguishing members. The lower member consists of rhyolitic crystal tuff, rhyolitic ignimbrite and tuff, whereas most of the upper member consists of andesite, which is the marker bed of the Daguding Fm. The Ehuling Fm., which shows discordant contact with the

underlying Daguding Fm., is extensively exposed throughout the basin, and it can be divided into three members in light of their eruption cycles and specific rock types. The first member primarily consists of rhyolitic crystal tuff and ignimbrite with tuffaceous sandstone at the bottom. The second member is composed primarily of rhyolite and ignimbrite with tuffaceous siltstone at the bottom. The major rock types of the third member are rhyolitic ignimbrite with tuff at the bottom.

In the Lengshuikeng district, the Ehuling Fm. outcrops continuously in accordance with the northeast towards the nappe structure, whereas the Daguding Fm. is barely exposed (Fig. 2). Since the late 1960s, many geological investigations have been conducted, and a number of giant Ag–Pb–Zn polymetallic mineral deposits have been explored in this area. According to their mineralization characteristics, these deposits are generally classified into two types: the porphyry type, which exists within the granite porphyry and the periphery of the rock, and the stratabound type, which occurs within the layered volcanic rocks of the Daguding Fm. and the Ehuling Fm. beneath the porphyry type. The former includes Yinkluling, Yinzhushan, and Baojia deposits, whereas the latter includes Xiabao, Yinglin, and Yinkeng deposits. As an important component of the Northern Wuyi Mountain metallogenic belt, the Lengshuikeng ore district has attracted considerable attention from many geologists since the discovery of the polymetallic mineral deposits, but the exact timing of the volcanic units remains controversial, and further investigations are required to clarify the geochronological framework. The first published geochronological data of the volcanic units suggest an age of 101 Ma for the Ehuling Fm. and an age range of 112 Ma to 123 Ma for the Daguding Fm. on the basis of the Rb–Sr and K–Ar method, respectively (Xu et al., 2001). The age of the Ehuling Fm. was updated to 129.7 Ma in 2002 (Xu et al., 2002). However, in situ zircon U–Pb, the dating results demonstrate that the eruption time of the Daguding Fm. was approximately 157 to 161 Ma, but the Ehuling Fm. shows conflicting ages, between 146.6 Ma and 161.8 Ma (Meng et al., 2012; Qiu et al., 2013; Wang et al., 2013; Xu et al., 2013). Recently, Su et al. (2014) reported that both the Daguding Fm. and the Ehuling Fm. formed in the Early Cretaceous (~144–137 Ma) based on zircon U–Pb ages. Hence, correct sampling and precise and accurate isotopic age determination is urgent and crucial for the geologic time of Late Mesozoic volcanism and for understanding the relationship between the volcanic rocks and the mineralization in the region.

Multiple stages of intrusive activities ranging from the Caledonian to Yanshanian periods occurred in this area. Except for the vastly exposed Caledonian migmatitic granites in the northwest part of the district, the rest can be divided into two main stages. The early stage is represented by Late Jurassic granite porphyry, which is primarily distributed in the center part of the basin and is associated with mineralization; the latter is represented by the Early Cretaceous granitoids, which are composed of quartz syenite porphyry, rhyolite porphyry and K-feldspar granite porphyry and are extensively developed throughout the basin. There are distinct, intrusive relationships among the early granite porphyry, the late rhyolite porphyry, the K-feldspar granite porphyry, and all of these intrusions and volcanic units on the geological map (Fig. 2).

3. Samples and analytical methods

3.1. Samples and petrography

The samples for this study were primarily collected from two drill cores along with the representative sections of volcanic units in the Lengshuikeng district. The light gray-green crystal tuff from the lower segment of the Daguding Fm., which was collected from the drill core ZK15150 and ZK15701, with one from the 188-2 footrill, consists of broken crystal fragments (45%-60%), volcanic ash and vitroclasts (Fig. 3a, b, c). The crystals are quartz and plagioclase, and a few are biotite. These samples show slightly pervasive hydrothermal alteration, such as chloritization, sericitization, silicification and carbonization. The biotite was replaced by chlorite or broken down to muscovite and Fe–Ti oxides along their cleavage planes and crystal margins (Fig. 3b). The feldspar fragments changed into a complex of sericite, chlorite, and carbonate (Fig. 3a, b, c).

Samples from the Ehuling Fm., which were collected from the drill core ZK15701 and outcrops, consisting of lithoclasts (5%-10%), crystals (5%-15%), vitroclastic and volcanic ash (Fig. 3d, e, f). The lower segment is light gray-green, whereas the upper segment is flesh pink. The crystals are quartz, and a few are plagioclases. The quartz has undergone corrosion to varying degrees with an irregular harbor shape, and some were clearly broken into smaller pieces (Fig. 3e). Argillic alteration is notable in some feldspar grains, and several samples show obvious ignimbrite texture and pseudofluidal structure (Fig. 3f). More detailed information about the lithologies, pyroclast assemblages and GPS positions of the studied samples are given in Table 2.

3.2. Zircon U–Pb and Hf isotope analyses

Zircon grains were separated from the volcanic rock samples through crushing, conventional magnetic and heavy liquid separation methods followed by handpicking under a binocular microscope. The selected grains were then mounted on adhesive tape, enclosed in epoxy resin, and polished to approximately 1/3 to 1/2 their thickness for analysis. Cathodoluminescence (CL) images of the zircon grains were taken using a Mono CL3 detector (manufactured by Gatan, U.S.A.) attached to a scanning electron microscope (manufactured by JSM6510, JEOL, Japan) at the Beijing Geoanalysis Co., Ltd. These images helped select suitable spots for in-situ U–Pb dating and Lu–Hf analysis.

The zircon U–Pb analyses were conducted using the LA–ICP–MS method at the State Key Laboratory for Mineral Deposits Research at Nanjing University, using an Agilent 7500a ICP–MS equipped with a New Wave Research 213 nm laser ablation system. The ablated material was transported in a He carrier gas through 3 mm i.d. PVC tubing and then combined with Ar in a 30 cm³ mixing chamber prior to entering the ICP–MS for isotopic measurement.

Fractionation and instrumental mass biases were corrected by direct calibration against a homogeneous zircon standard, GEMOC/GJ-1 (608.5 Ma; Jackson et al., 2004), analyzed under carefully matched conditions. Samples were analyzed in ‘runs’ of ca. 15 analyses, which included 11 unknowns and were bracketed by two analyses of the standard. The unknowns included one analysis of an external zircon standard, Mud Tank (732Ma) (Black and Gulson, 1978), as an

independent control on reproducibility and instrument stability. The counting time for one analysis was typically 100 s (40 s measuring gas blank to establish the background and 60 s for the remainder of the analysis). The repetition rate of the laser was 5 Hz, with a fluence of approximately 20 J/cm². The spot size varied between 25 and 35 µm in diameter according to the sizes of zircon grains. Detailed analytical procedures were similar to those described by Jackson et al. (2004). The raw ICP-MS data were exported in ASCII format and processed using GLITTER (Van Achterbergh et al., 2001). Common lead contents were corrected using the method of Andersen (2002). The age calculations and plotting of Concordia diagrams were performed using Isoplot 3 (Ludwig, 2003).

Zircon Lu-Hf isotope analysis was conducted on the same spot or in the same domain of previous U–Pb analyses according to the CL images, using a New Wave UP193FX laser ablation microprobe attached to a Neptune multi-collector ICP-MS in the same lab at Nanjing University. The instrumental conditions and data acquisition were similar to those described by Wu et al. (2006) and Hou et al. (2007). Analyses were performed predominantly with a beam diameter of 35 µm and a repetition rate of 8 Hz. Both He and Ar carrier gases were used to transport the ablated sample from the laser-ablation cell via a mixing chamber to the ICP-MS torch. Atomic masses 172, 173, 175 to 180 and 182 were simultaneously measured in the static-collection mode. Corrections for isobaric interferences of ¹⁷⁶Yb and ¹⁷⁶Lu on ¹⁷⁶Hf referred to Iizuka and Hirata (2005), Wu et al. (2006) and Hou et al. (2007). The reference zircon Mud Tank (¹⁷⁶Hf/¹⁷⁷Hf = 0.282507 ± 6; Woodhead and Hergt, 2005) and the standard zircon 91500 (Wiedenbeck et al., 1995) were used to monitor performance conditions and analytical accuracy.

The initial Hf values (in term of ε_{Hf}) were calculated using a decay constant of 1.867 × 10⁻¹¹ per year (Söderlund et al., 2004) for ¹⁷⁶Lu and a chondritic model with ¹⁷⁶Hf/¹⁷⁷Hf = 0.282785 and ¹⁷⁶Lu/¹⁷⁷Hf = 0.0336 (Bouvier et al., 2008). Depleted mantle Hf model ages (T_{DM}) were calculated using the measured ¹⁷⁶Lu/¹⁷⁷Hf ratios of zircon, referring to a model of depleted mantle with a present-day ¹⁷⁶Hf/¹⁷⁷Hf ratio of 0.28325, similar to that of an average MORB (Nowell et al., 1998) and ¹⁷⁶Lu/¹⁷⁷Hf = 0.0384 (Griffin et al., 2000). “Crustal” model ages (T_{DM2}) were also calculated by projecting the initial ¹⁷⁶Hf/¹⁷⁷Hf ratios of the zircon to the depleted mantle model growth line using a mean ¹⁷⁶Lu/¹⁷⁷Hf value (0.015) for the average continental crust (Griffin et al., 2002).

3.3. Major and trace element analysis

Because surface-exposed volcanic rock is more prone to late surface weathering, fifteen unaltered or slightly altered samples from drill hole 15701 and drill hole 15150 were chosen for bulk-rock geochemical analysis. The locations of the two drill holes are shown in Fig. 2. The whole-rock samples were crushed and ground to 200-mesh using an agate ring mill.

All major and trace elements analyses for the samples were conducted at the State Key Laboratory for Mineral Deposits Research in Nanjing University. The major elements were determined by wavelength-dispersive X-ray fluorescence spectrometry (XRF) on fused glass, following a procedure described by Norrish and Hutton (1969) and with analytical uncertainties better than 0.5%. For trace element analyses, approximately 50 mg of the powered sample was digested using an HF + HNO₃ mixture in high-pressure Teflon bombs at 190 °C for 72 h. Rh was used as an internal standard to monitor signal drift during ICP-MS measurement. The trace

elements were measured using Finnigan Element II inductively coupled plasma mass spectrometry (ICP-MS). Standards and duplicated analyses indicated that the precisions of trace element analyses were generally better than 5%, and the accuracies were better than 10%. Detailed analytical procedures have been outlined previously in [Gao et al. \(2003\)](#).

3.4. Whole-rock Sr-Nd isotope analysis

The Sr and Nd isotopic compositions were analyzed using Finnigan Triton TI thermal ionization mass spectrometry (TIMS) at the State Key Laboratory for Mineral Deposits Research in Nanjing University. Samples of approximately 100 mg were dissolved in the same way as for trace element analyses. The complete separation of Sr was accomplished by a combination of cation-exchange chromatography in H^+ form and pyridinium form with the DCTA complex. Nd was then separated from the REE fractions by cation-exchange resin using HIBA as a cation exchange medium. After purification, the separated Sr was dissolved in 1 μ L of 1 N HCl and then loaded with TaF_5 solution onto pre-conditioned W filaments for $^{87}Sr/^{86}Sr$ analysis on TIMS. The separated Nd was dissolved in 1 μ L of 1N HCl and then loaded with phosphate acid onto Re double-filaments for Nd isotope analysis on TIMS. The Sr and Nd isotopic ratios were corrected for mass fractionation relative to the natural $^{86}Sr/^{88}Sr = 0.1194$ and $^{146}Nd/^{144}Nd = 0.7219$, respectively. During the course of this study, measurements of NIST SRM-987 Sr standard yielded an $^{87}Sr/^{86}Sr$ ratio of 0.710252 ± 0.000016 (2σ), and the JNdi-1Nd standard yielded a $^{143}Nd/^{144}Nd$ ratio of 0.512121 ± 0.000016 (2σ). More details on the analytical procedures are given in [Pu et al. \(2004, 2005\)](#). We used the following parameters to calculate the I_{Sr} , $\epsilon_{Nd}(t)$ values and Nd model ages: $\lambda_{Rb} = 1.42 \times 10^{-11} \text{ year}^{-1}$ ([Steiger and Jäger, 1977](#)); $\lambda_{Sm} = 6.54 \times 10^{-12} \text{ year}^{-1}$ ([Lugmair and Marti, 1978](#)); $(^{147}Sm/^{144}Nd)_{CHUR} = 0.1967$ ([Jacobsen and Wasserburg, 1980](#)); $(^{146}Nd/^{144}Nd)_{CHUR} = 0.512638$ ([Goldstein et al., 1984](#)); and $(^{146}Nd/^{144}Nd)_{DM} = 0.513151$, $(^{147}Sm/^{144}Nd)_{DM} = 0.2136$ ([Liew and Hofmann, 1988](#)). The $^{147}Sm/^{144}Nd$ value of 0.118 ([Jahn and Condie, 1995](#)) for average continental crust was used for the mantle extraction model age (T_{DM2}) for the source rocks of the magmas.

4. Results

4.1. Zircon U–Pb ages

CL images of representative zircon grains are shown in [Fig. 4](#). The Concordia diagrams and weighted averages of zircon U–Pb dating are presented in [Fig. 5](#). The detailed dating results are listed in [Appendix A](#). Because of errors in counting statistics during the analysis, the $^{207}Pb/^{206}Pb$ ages are more precise for older (>1 Ga) zircons, whereas the $^{206}Pb/^{238}U$ ages are more precise for younger zircons ([Griffin et al., 2004](#)). Therefore, we use the $^{207}Pb/^{206}Pb$ ages for older zircons (detrital) and $^{206}Pb/^{238}U$ ages for younger zircons in the following discussion.

4.1.1. Daguding Fm.

All zircons of the Daguding Fm are euhedral and typically 100-250 μm long. They exhibit well-developed growth zoning with corresponding high Th/U ratios (>0.4), which are typical characteristics of igneous zircons (Belousova et al., 2002). Most of them are relatively transparent and colorless.

Three samples from different depths of drill hole 15701 yield concordant and consistent age results, with 159.4 ± 2.4 Ma (2σ ; $n=14$; MSWD=1.5; Fig. 5a) for 701-840, 158.9 ± 1.5 Ma (2σ ; $n=21$; MSWD=0.59; Fig. 5b) for 701-728, and 158.3 ± 1.7 Ma (2σ ; $n=27$; MSWD=1.4; Fig. 5c) for 701-700, respectively. Two samples from drill hole 15150 give similar results, with 155.9 ± 1.8 Ma (2σ ; $n=22$; MSWD=1.4; Fig. 5d) for 150-380 and 155.0 ± 2.2 Ma (2σ ; $n=29$; MSWD=2.4; Fig. 5e) for 150-245. Sixteen analyses for sample LS45 from the outcrop yield a slightly younger weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 151.6 ± 4.1 Ma (2σ ; $n=16$; MSWD=1.06; Fig. 5f). All six samples from the lower, middle and upper parts of the Daguding Fm. suggest an eruption age of the Daguding Fm. of approximately 160 Ma.

4.1.2. Ehuling Fm.

All zircon grains from the Ehuling Fm. samples are euhedral and 100-200 μm long. They all show clear oscillatory zonations with high Th/U ratios of magmatic origin. Three samples from different depths of drill hole 15701 show concordant ages with weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 137.5 ± 1.3 Ma (701-637; 2σ ; $n=19$; MSWD=1.5; Fig. 5g), 133.3 ± 1.2 Ma (701-254; 2σ ; $n=24$; MSWD=1.4; Fig. 5h) and 130.4 ± 1.1 Ma (701-246; 2σ ; $n=18$; MSWD=0.99; Fig. 5i) for different segments of the Ehuling Fm. Two samples from the outcrop show concordant and consistent ages, with weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 138.4 ± 1.3 Ma (LS79; 2σ ; $n=17$; MSWD=1.5; Fig. 5j) and 137.5 ± 1.2 Ma (LS68; 2σ ; $n=22$; MSWD=1.8; Fig. 5k) for the middle and upper parts of the Ehuling Fm., respectively. All of these results demonstrate that the age of the Ehuling Fm. is approximately 130 Ma.

4.2. Zircon Hf isotopic compositions

A total of eight volcanic rocks were chosen for zircon Lu–Hf analyses, and approximately 30 zircon grains were used for each sample. The zircon Hf analyses were performed on the same domains of the same grain as those used for U–Pb dating (Fig. 4). The analytical results of the Lu–Hf isotopic compositions are given in Appendix B, and the frequency distribution diagrams of $\epsilon_{\text{Hf}}(t)$ and T_{DM2} values are presented in Fig. 6 and Fig. 7.

4.2.1. Daguding Fm.

The zircon Hf isotopic compositions of two samples, 701-728 and 701-700, from drill hole 15701 are nearly identical. Thirty analyses for 701-728 from the lower part of the Daguding Fm. yield initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios from 0.282243 to 0.282409 and $\epsilon_{\text{Hf}}(t)$ values from -15.7 to -9.8, with a weighted mean of -12.3 ± 0.4 (2SD), corresponding to two-stage model ages (T_{DM2}) of 1.80-2.16

Ga. Thirty-four zircon grains from 701-700 were used for Lu–Hf isotope analyses. Thirty-three of them yield similar initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios (0.282216-0.282412) and $\varepsilon_{\text{Hf}}(t)$ values (-16.7 to -9.8), with a weighted mean of -12.6 ± 0.4 (2SD), corresponding to a T_{DM2} of 1.80-2.22 Ga. Zircons from samples 150-380 and 150-245 from drill hole 15150 show similar $(^{176}\text{Hf}/^{177}\text{Hf})_i$ ratios of 0.282150-0.282395 and 0.282207-0.282483, respectively. The $\varepsilon_{\text{Hf}}(t)$ values are -19.1 to -10.4 and -17.1 to -7.3, with weighted means of -11.90 ± 0.44 (2 SD; n=32) and -12.20 ± 0.58 (2 SD; n=34) and T_{DM2} model ages of 1.83-2.37Ga and 1.64-2.25 Ga, respectively. Twenty-nine grains selected from the crystal tuff (LS45) of the upper part of the Daguding Fm. give initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of 0.282324-0.282434 and $\varepsilon_{\text{Hf}}(t)$ values of -13.0 to -9.2, with a weighted mean of -11.41 ± 0.31 (2 SD), corresponding to T_{DM2} model ages of 1.75-1.99 Ga.

All five samples have relatively homogenous zircon Hf isotope compositions. A total of 159 zircon grains yield initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of 0.282017-0.282483 and $\varepsilon_{\text{Hf}}(t)$ values of -21.9 to -7.3 (Fig. 6b) with a weighted mean of -12.15 ± 0.20 , corresponding to a T_{DM2} of 1.64-2.55 Ga (Fig. 7b).

4.2.2. Ehuling Fm.

Three samples from different depths of drill hole 15701 were selected for the Lu–Hf analyses. Zircon Hf isotope analyses were conducted on thirty-five grains for 701-637 from the lower part of the Ehuling Fm. These analyses yielded initial $^{176}\text{Hf}/^{177}\text{Hf}$ values between 0.282233 and 0.282491 and $\varepsilon_{\text{Hf}}(t)$ values between -16.8 and -7.4, with a weighted mean of -10.63 ± 0.77 (2 SD) and T_{DM2} ages from 1.63 to 2.21 Ga. Concordant and similar zircon Hf isotopic compositions were obtained for samples 701-254 and 701-246 from the upper parts of the formation. Their $(^{176}\text{Hf}/^{177}\text{Hf})_i$ ratios are 0.28280-0.282446 and 0.282159-0.282424, corresponding to $\varepsilon_{\text{Hf}}(t)$ values of -15.0 to -9.2 and -19.4 to -9.9, with weighted means of -11.68 ± 0.50 (2 SD; n=29) and -12.03 ± 0.56 (2 SD; n=35). The T_{DM2} model ages are 1.73-2.10 Ga and 1.78-2.37 Ga, respectively.

Overall, a total of 99 zircon grains yield initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of 0.282159-0.282491 and $\varepsilon_{\text{Hf}}(t)$ values of -19.4 to -7.4 (Fig. 6a), with a weighted mean of -11.6 ± 0.4 , corresponding to a T_{DM2} of 1.63-2.37 Ga (Fig. 7a).

4.3. Major and trace elements

4.3.1. Daguding Fm.

Daguding volcanic rocks have SiO_2 contents of 70.9-76.3 wt%. They all fall in the rhyolite field in TAS diagram (Fig. 8). These rocks are strongly peraluminous with values of the alumina saturation index ASI [=molar $\text{Al}_2\text{O}_3/(\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O})$] >1.1 (Table 3; Fig. 9a). These rocks also have high K_2O contents, with all data plotting in the high-K calc-alkaline and shoshonite fields in a K_2O vs SiO_2 classification diagram (Fig. 9b). These rocks show higher Mg# [=Mg/(Mg+Fe_T)] than pure crustal melts (Fig. 9c). In a plot of AR (alkalinity ratio) vs SiO_2 , all samples plot in the alkaline field, except for a few that straddle the boundary with the calc-alkaline field (Fig. 10).

Daguding volcanic rocks have total rare earth element (ΣREE) contents of 111-227 ppm and exhibit an enrichment of light REE (LREE) relative to heavy REE (HREE) with $(\text{La}_N/\text{Yb}_N)$ ratios ranging from 7.9 to 28.5. They also show notable negative Eu anomalies ($\text{Eu}/\text{Eu}^*=0.32\text{-}0.65$; Fig.

11a). Primitive mantle normalized trace element patterns of the samples show significantly negative Ba, Sr, Nb, Ta, Ti and P anomalies (Fig. 11b). The calculated zircon saturation temperatures are between 700 and 824°C (mean =796°C; Table 3).

4.3.2. Ehuling Fm.

Ehuling volcanic rocks have high SiO₂ contents in the range of 74.3 to 76.1 wt%. They also fall in the rhyolite field in the TAS diagram (Fig. 8). Most samples are weakly peraluminous (Table 3; Fig. 9a). These volcanic rocks are all alkali-rich with high K₂O contents (Fig. 9b), and the data are plotted in the alkaline field (Fig. 10). They show higher FeO*/MgO ratios and higher alkali contents than the Daguding volcanic rocks (Table 3; Fig. 12b). Most Ehuling volcanic rocks have a low Mg#, lower than pure crustal melts (Fig. 9c).

Ehuling volcanic rocks have relatively high REE contents (238-515 ppm) with LREE enrichment relative to HREE in all rocks (La_N/Yb_N =9.9 to 31.2). The rocks show significantly negative Eu anomalies (Eu/Eu* =0.04-0.09; Fig. 11a). In the primitive-mantle normalized trace element diagrams, they show significantly negative Ba and Sr anomalies (Fig. 11b). These volcanic rocks have high Ga/Al ratios (Fig. 12a) and high HFSE contents (Fig. 12b). The calculated zircon saturation temperatures for these rocks are relatively high (800-867 °C) with a mean temperature of 837°C (Table 3).

4.4. Sr–Nd isotopic compositions

4.4.1. Daguding Fm.

Four samples from the Daguding Fm. show a range of ⁸⁷Rb/⁸⁶Sr ratios from 5.13 to 7.66 and ⁸⁷Sr/⁸⁶Sr ratios from 0.720365 to 0.725540. The ¹⁴⁷Sm/¹⁴⁴Nd ratios range from 0.0910 to 0.1386, and ¹⁴³Nd/¹⁴⁴Nd ratios vary between 0.512025 and 0.512072. The initial ⁸⁷Sr/⁸⁶Sr ratios (I_{Sr}) and ε_{Nd}(t) values are calculated based on the respective zircon U–Pb ages determined in this study. The initial ⁸⁷Sr/⁸⁶Sr ratios cluster between 0.706174 and 0.710594, and the ε_{Nd}(t) values range from -10.2 to -9.5 (Table 3; Fig. 13), with the corresponding T_{DM2} model ages between 1.70 and 1.77 Ga (Table 3).

4.4.2. Ehuling Fm.

Ehuling volcanic rocks show similar Nd isotopic compositions, with ε_{Nd}(t) values of -9.5 to -8.8, whereas their Sr isotopic compositions are more variable, with initial ⁸⁷Sr/⁸⁶Sr ratios from 0.71258 to 0.71919 (Table 3; Fig. 13). The T_{DM2} model ages range from 1.641 to 1.702 Ga (Table 3).

5. Discussion

5.1 Timing of volcanism in northeastern Jiangxi

The Late Mesozoic volcanic rocks are well developed in the volcanic basins (such as Xiangshan, Shenyuan, and Tianhuashan) in the northeastern Jiangxi Province. According to their eruption cycles, stratigraphic sequences and contact relationships, they can be divided into the Daguding Fm. on the bottom and the Ehuling Fm. on the top. Since the late 1980s, many geochronological studies have been conducted, and a great number of ages have been reported through various methods (Table 4; Fig. 14b). Unfortunately, some of the published results are either of poor quality or are inconsistent with each other, as a result of defective experimental conditions or the violation of the principle of radioactive dating. On the whole, it is obvious that the K–Ar and Ar–Ar ages are usually much younger than ages obtained by the Rb–Sr or U–Pb methods. Moreover, some isotopic ages of the Ehuling Fm. are older than those of the underlying Daguding Fm., which is inconsistent with the geology. Because the closure temperatures of K–Ar and Rb–Sr isotopic systems are low in both single minerals and whole rocks (Dodson, 1973; Ganguly and Tirone, 1999; Harrison and McDougall, 1980; Villa, 1998), they can be easily disturbed by subsequent extensive hydrothermal alteration or other thermo-tectonic events. In contrast, the lowest closure temperature of the U–Pb isotope system in zircon can reach 800 °C (Lee et al., 1997; Cherniak and Watson, 2001), which is higher than the temperature of felsic magma (Carmichael et al., 1974). Therefore, the U–Pb isotope system in zircon could stay stable and closed in the geothermal event since the magmatic crystallization. In addition, the zircon U–Pb dating method has much smaller systematic error than the K–Ar and Rb–Sr methods. Therefore, the zircon U–Pb dating method would be the most appropriate dating method for multi-stage magmatic activities, and it has been widely used (e.g., Condie et al., 2009; Li et al., 2008; Li et al., 2011; Liu et al., 2012; Wan et al., 2006; Ye et al., 2007; Zhu et al., 2013). To mitigate the effect of dating methods, the following discussions based primarily on the zircon U–Pb data. However, except for the results based on Rb–Sr and K–Ar methods, controversies about the eruption ages of these volcanic rocks remain.

First, different researchers have obtained different ages on the same stratum in the same basin, when even using the same methods. Previous geologists have suggested that the volcanic-intrusive complex in the Xiangshan area was formed from the middle Jurassic to the early Cretaceous (Fang et al., 1982; Li et al., 1989; Liu, 1985). However, recent precise zircon U–Pb dating results have shown that both the Daguding and Ehuling Fms. at the Xiangshan basin were formed in the early Cretaceous (Fan et al., 2005; Yang et al., 2011; Yang and Jiang, 2013; Fig. 14b). In the Tianhuashan basin, Meng et al. (2012) reported that the crystal tuff from the Ehuling Fm. formed at ca. 157.2–158.2 Ma based on the U–Pb dating of zircon by the SHRIMP method, and they suggested that both the Ehuling and Daguding Fms. were formed in the Late Jurassic. This viewpoint received support from later published data (Qiu et al., 2013; Xu et al., 2013). Nevertheless, a recent geochronological investigation indicated that both the Ehuling and Daguding Fms.' volcanic rocks likely formed in 144–137 Ma, the Early Cretaceous (Su et al., 2014). Considering the continuous debate on the ages of these volcanic rocks, we preferred to collect the samples from well-logged drill cores and then systematically measure the ages of the

volcanic sequences in the Tianhuashan basin by the LA-ICP-MS U–Pb zircon dating method. Our results suggest that the Daguding Fm. was formed in the Late Jurassic (150–160 Ma), whereas the Ehuling Fm. was formed in the Early Cretaceous (130–140 Ma; Fig. 14a). The confirmation of two-stage eruptions is not only consistent with most of the published ages and the distributions of their respective stratum (Fig. 14b) but also accords with the two corresponding stages of intrusive rocks (Meng et al., 2012; Qiu et al., 2013; Su et al., 2014; Wang et al., 2013; Zuo et al., 2010). Furthermore, the field contact relations that the Late Jurassic granite porphyry intruded into the volcanic strata demonstrate that the Daguding Fm. formed in the Late Jurassic. Those published Late Jurassic ages of the Ehuling Fm. (Meng et al., 2012; Xu et al., 2013) may be due to an improper sampling; i.e., the measured crystal tuff samples were actually from the Daguding Fm. rather than the Ehuling Fm.

Second, the geological framework of the volcanic sequences among the basins was inconsistent. As discussed above, the volcanic sequences in the Tianhuashan basin formed in the Late Jurassic and Early Cretaceous, respectively. By comparison, all geochronological results consistently show that both the Ehuling Fm. and the Daguding Fm. in the Xiangshan and Shengyuan basins were formed in the Early Cretaceous (Fig. 14b; Wu et al., 2011; Yang et al., 2011; Yang et al., 2013; Zhang, 2000; Zhang and Li, 2007). Two situations may be responsible for the different dating results of the same named volcanic strata between different volcanic basins: (1) The Late Jurassic volcanic rocks in the Xiangshan and Shengyuan basins have not yet been found, and (2) the volcanic sequences between the Gan-Hang tectonic belt and the northern Wuyi Mountain volcanic belt are different, as are their corresponding ages. Therefore, more specific and systematic geochronological work is crucial to establish the chronology framework of magmatism and volcanism in the Xiangshan basin and the other volcanic basins in northeastern Jiangxi Province. The Daguding Fm. and Ehuling Fm. should be reexamined in the area.

5.2. Petrogenesis of the late Jurassic Daguding volcanic rocks

5.2.1. Genetic type

The Daguding volcanic rocks are strongly peraluminous with an ASI > 1.1. They fall into the cordierite-muscovite field in the ACF diagram (Fig. 15, Chappell and White, 1992) and have relatively low Na₂O and CaO contents, which indicate they belong to S-type volcanics (Chappell and White, 1992; Chappell and White, 2001). Moreover, Daguding volcanic rocks also show a slight decrease in Y and Th as Rb increases, typical of the S-type granite evolution trend. Most peraluminous granitoids are divided into two distinct groups: muscovite-bearing peraluminous granitoids (MPGs) and cordierite-bearing peraluminous granitoids (CPGs), based on the mineral and rock associations and the variation of their peraluminosity during differentiation (Barbarin, 1996). As suggested by Barbarin (1996), MPGs, which are relatively poor in biotite and contain large primary muscovite, are produced primarily in orogenic zones affected by major crustal shear or overthrust structures, through the “wet” anatexis of crustal rocks and the crystal fractionation of magmas. Nevertheless, the CPGs, which contain cordierite and accessory aluminum silicates, are suggested to be produced through the ‘dry’ anatexis of crustal rocks enhanced by underplating or the injection of hot mantle-derived magmas (Barbarin, 1996). The model emphasizes the

importance of the condition of crustal partial melting in the genesis of the two types of peraluminous granitoids.

Daguding volcanic rocks contain biotite (Fig. 3b) and show relatively high magmatic temperatures (770-837 °C; Table 3), which are similar to Tongshanling granite in the southern Hunan province, which Jiang et al. (2009) attributed to CPGs. Thus, these volcanic rocks should resemble the counterparts of CPGs.

5.2.2. Petrogenetic model

As mentioned above, Daguding volcanic rocks, which resemble the counterparts of CPGs, were likely derived from the anatexis of crustal rocks induced by the underplating or injection of hot mantle-derived magmas (Barbarin, 1996). These volcanics are strongly peraluminous, suggesting that primarily metasedimentary rocks were involved in partial melting. Their radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ (0.70617-0.71059), unradiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ ratios ($\epsilon_{\text{Nd}}(t) = -10.2$ to -9.2) and negative initial $\epsilon_{\text{Hf}}(t)$ values (-21.9 to -7.3) also indicate that the source rocks are primarily metasedimentary rocks rather than metaigneous rocks. The Paleoproterozoic T_{DM2} ages of both Nd and Hf isotopes require a Paleoproterozoic source rock (Fig. 13). The Paleoproterozoic basement in this region, the Mayuan Group, is composed of biotite plagioclase granulite and biotite gneiss. Experiments have demonstrated that the partial melting of biotite gneiss and quartz amphibolite from 3 to 15 kbar could generate high Mg# melts such as Daguding (Patino Douce and Beard, 1995). These volcanic rocks share the same Nd isotopic compositions as the Mayuan Group (Chen and Jahn, 1998). Their Hf isotopic compositions also fall roughly in the Hf isotope evolution trends of the known Proterozoic rocks (Fig. 16). Thus, these rocks were likely produced by the remelting of these ancient crustal rocks.

Daguding volcanic rocks have a relatively low HREE (Yb=1.0-2.2 ppm) and Y (15.6-23.7 ppm) contents and exhibit relatively steep HREE patterns ($\text{Gd}_\text{N}/\text{Yb}_\text{N}=1.7\text{-}3.0$; Fig. 11a), which suggests the garnet rather than amphibole is the main residual assemblage in the source. They also have a relatively low Sr (44-129 ppm) and show obviously negative Eu anomalies (Fig. 11a), which require that the source region be within the stability field of plagioclase and/or K-feldspar. In addition, volcanic rocks have very high Ba concentrations (363-904 ppm, Table 4), indicating that plagioclase rather than K-feldspar is the main residual assemblage in the source. Experiments indicate that during the dehydration-melting of metasedimentary (biotite gneiss), plagioclase keeps stable when the pressure is <15 kbar, whereas the garnet remains in residual phases at pressures >12.5 kbar (Patino Douce and Beard, 1995). The fact that both garnet and plagioclase remain as a residual assemblage implies that the source region of the Daguding volcanic rocks is relatively deep (~40-50 km, corresponding to 12.5-15 kbar).

5.3. Petrogenesis of the early Cretaceous Ehuling volcanic rocks

5.3.1. Genetic type

The geochemical characteristics of the Ehuling volcanic rocks suggest an A-type affinity. These volcanic rocks are enriched in alkalis (Figs. 10), REE (except for Eu, Fig. 11a) and HFSE

but are depleted in Ba and Sr (Fig. 11b). They also exhibit high Ga/Al ratios (Fig. 12a). These features of A-type granites have been recognized by many researchers (Collins et al., 1982; Eby, 1992; Whalen et al., 1987). Chappell and White (1992) noted that highly fractionated felsic I- and S-type granites can have major and trace element compositions overlapping those of typical A-type granites. However, A-type granites have much higher amounts of HFSE than I- and S-type granites. The fractionated I- and S-type granites evolve towards the A-type in terms of Nb and Ga, relative to the unfractionated granites, but the concomitant changes lead to higher Rb and U and lower Ba and Sr, which are distinct from A-type granites. Compared with A-type granites, fractionated S-type granites contain more Al_2O_3 and P_2O_5 . Furthermore, fractionated felsic granites tend to have HFSE trends that display a negative correlation with increasing crystal fractionation (King et al., 1997). In the discrimination diagram of FeO^*/MgO vs. $(\text{Zr}+\text{Nb}+\text{Y}+\text{Ce})$; (Fig. 12b; Whalen et al., 1987), they are plotted in the A-type granite field rather than fractionated granites. They also fall in the field of A-type granites in the Zr vs. $10000 \times \text{Ga}/\text{Al}$ diagram (Fig. 12a). The High Rb/Nb and Y/Nb ratios of these rocks further indicate that the Ehuling volcanic rocks belong to the A_2 group (Fig. 12c and d), which indicate these volcanic rocks might have derived from continental crust or underplated crust that has been through a cycle of continent-continent collision or island-arc magmatism (Eby, 1992). The calculated zircon saturation temperatures (T_{Zr} ; Watson and Harrison, 1983) of the Ehuling volcanic rocks are 826-867 °C (average 838°C, Table 3), which suggest that these volcanic rocks were formed at a high temperature.

5.3.2. Petrogenetic model

Various petrogenetic schemes have been proposed for the origin of A-type granitic and volcanic rocks, including extensive fractional crystallization from mantle-derived basaltic magmas (e.g., Anderson et al., 2003; Smith et al., 1999; Turner et al., 1992) and partial melting at elevated temperatures of specific continental crustal protoliths that have been depleted in water by the extraction of hydrous felsic magma (e.g., Collins et al., 1982; Creaser et al., 1991; Eby, 1992; Whalen et al., 1987). Additionally, Patino Douce (1997) emphasized the important role of the physical parameters of partial melting over the nature of the sources in engendering A-type melts.

It is worth noting that a number of Early Cretaceous A-type granites have been identified along the Gan-Hang Belt, including the Xiangshan volcanic-intrusive complex (135-137 Ma, Yang et al., 2011), Yangmeiwan granite (135 Ma, Yang et al., 2012), Daqiaowu granitic porphyry (133-136 Ma, Yang et al., 2012), Tongshan granite (129 Ma, Jiang et al., 2011), Baijhuajian granite (126 Ma, Wong et al., 2009) and Damaoshan granite (122-126 Ma, Jiang et al., 2011). There are two distinct views on the petrogenesis of these A-type granitic rocks. Wong et al. (2009) suggested that the Baijhuajian magma was derived originally from asthenospheric mantle and experienced extensive crystal fractionation and minor mixing with old crustal components. Others have preferred a formation model by the partial melting of ancient crustal protoliths followed by fractional crystallization (Jiang et al., 2011; Yang et al., 2011; Yang et al., 2012).

All Ehuling volcanic rocks are characterized by clearly negative and uniform $\epsilon_{\text{Nd}}(t)$ values of -9.5 to -8.8 and high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.71258-0.71919, which exclude the possibility that these A-type volcanic rocks were formed by extensive fractional crystallization from coeval mafic magmas. Instead, they show a close relationship to Proterozoic metamorphic basement rocks in

terms of Sr–Nd isotopic compositions (Fig. 13), indicating that they were most likely derived from these ancient crustal rocks. This finding is consistent with the petrogenesis of the A₂ group proposed by former geologists (Eby, 1992). These volcanic rocks are primarily slightly peraluminous (Fig. 9a), suggesting that metaigneous rocks and metasedimentary rocks melted to form these magmas. This view is further supported by the rocks' Sr–Nd isotopic compositions. Compared to metasedimentary rocks at the age of eruption, Ehuling volcanic rocks have much lower initial ⁸⁷Sr/⁸⁶Sr ratios and slightly higher ε_{Nd}(t) values (Fig. 13), which indicate that a small proportion of metaigneous rocks was likely involved in partial melting. Furthermore, these volcanic rocks had Paleoproterozoic T_{DM2} ages of both Nd and Hf isotopes, which suggests a major Paleoproterozoic crustal source. Their Hf isotopic compositions are also roughly consistent with the Hf isotope evolution trends of the Cathaysia crustal basement (Fig. 16), attesting to their crustal sources.

Ehuling volcanic rocks show relatively steep HREE patterns (Fig. 11a), a depletion of Sr and Ba (Fig. 11b) and significantly negative Eu anomalies (Fig. 11a), therefore requiring the melting of a source rock within the stability field of plagioclase and K-feldspar. Most of these volcanic rocks have a very low Mg#, lower than that of pure crustal melts, except for four samples (701-251, 701-258, 701-286 and 701-662), which show an Mg# consistent with pure crustal partial melts (Fig. 9c). Samples 701-258, which also have relatively high Eu, Sr and Ba contents, were regarded as the initial crustal partial melts. Other samples show a lower Mg# than pure crustal melts (Fig. 9c), suggesting that the magmas might have undergone the subsequent fractionation of mafic minerals, such as biotite, after partial melting. Element variations indicate that plagioclase and K-feldspar fractionation is also required (Fig. 17).

5.4. Tectonic evolution of Southeast China during the late Mesozoic

5.4.1. Late Jurassic (approximately 160 Ma)

The extensive magmatism and associated numerous mineral deposits in South China have attracted considerable attention from geologists for decades. However, there is an ongoing debate concerning the tectonic settings and geodynamics responsible for the extensive late Mesozoic magmatism (Charvet et al., 1994; Chen et al., 2008; Gilder et al., 1996; Jiang et al., 2009; Jiang et al., 2011; John et al., 1990; Lapierre et al., 1997; Li et al., 2003; Li et al., 2007; Li and Li, 2007; Li et al., 2014).

Li et al. (2004) proposed an intraplate rifting and extension environment for the genesis of Jurassic alkaline basalts and “intraplate-type” syenites in the interior of South China, and the secular variations in rock types and geochemical and isotopic characteristics were interpreted as a response to the lithosphere's extension and thinning. The occurrence of coeval A-type felsic and mafic magmatism in southern Jiangxi also provides support for an extension-to-rifting regime prevalent in much of the southeast China interior during the Middle to Late Jurassic times (He et al., 2010; Li et al., 2003).

Chen et al. (2008) investigated two Mesozoic basalt successions in South China and found that the east-west trending basalts extending from the Cathaysia Interior to the Cathaysia Folded Belt, which decreased in age from 175 to 98 Ma in an eastward direction, were derived from the

lithosphere-modified OIB-like asthenosphere. It was distinctly different from the Southeast Coast Magmatic Belt, which had been modified by the paleo-Pacific subduction system. These results indicated that the Jurassic mantle beneath the Cathaysia Interior and the Cathaysia Folded Belt was rather homogeneous and undisturbed by the subduction system before the Late Cretaceous, which is in accordance with the occurrence of two Early Cretaceous gabbroic intrusions in the coastal area of Fujian Province (Zhao et al., 2007). Hence, Jurassic magmatism in South China is explained as being due primarily to post-Indosinian continental extension and post-orogenic extension accompanied by mafic underplating, which may have been the cause of extensive Jurassic granitic magmatism in South China (Chen et al., 2008). However, the unconformity between strata does not support the above views. For example, the Early Cretaceous Dongkeng volcanic basins in southern Jiangxi unconformably overlie the pre-Jurassic strata (Chen et al., 2007) and the Early Cretaceous Nanyuan formation unconformably rests on the Early Jurassic Lishan Formation (Xing et al., 2008). Meanwhile, angular unconformity also widely existed in the northeast of the Gan-Hang Belt, with most of the unconformity caused by folding (Li et al., 2013). Moreover, the two Jurassic NE-trending A-type granite belts are inconsistent with the east-west linear array expected from the post-collisional model (Jiang et al., 2015).

Overall, most researchers have accepted that South China was influenced by the subduction of the Paleo-Pacific plate during the late Mesozoic. Unfortunately, the style of subduction is still debatable. Lapierre et al. (1997) investigated the high-K calc-alkaline volcanic and plutonic rocks in the coastal region and suggested that a normal subduction continued from the Late Jurassic to the Early Cretaceous, but this subduction-related model fails to explain the >1000-km-wide magmatic belt in South China. Zhou and Li (2000) demonstrated that the slab dip angle of Paleo-Pacific plate subduction underneath Southeastern China increased from a very low angle to a medium angle from the Early Jurassic to the Late Cretaceous, which they interpreted as accounting for the exceptionally wide magmatic arc and the migration of magmatic activity oceanward to the southeast. Unfortunately, this modified subduction model does not explain the intraplate feature of the Jurassic A-type granites and mafic intrusive and volcanic rocks in the interior of South China (Chen et al., 2008; Jiang et al., 2009; Jiang et al., 2015; Li et al., 2003; Li et al., 2004; Li et al., 2007). Li and Li (2007) proposed a model of flat-slab subduction on the basis of structural, geochronological and sedimentary facies observations. This model suggests that the subducted Paleo-Pacific plate reached the Yangtze Block by the Late Triassic (ca. 210 Ma), and the Jurassic magmatism was ascribed to the break-off and foundering event of the subducting/subducted flat slab in the Early Jurassic (Li et al., 2007; Li and Li, 2007). Geochronological and geochemical results from Mesozoic basalts and Jiufeng pluton in the South China Block also support the flat-slab subduction model (Huang et al., 2015; Meng et al., 2012). However, this model is inconsistent with the observations of He et al. (2010). First, slab break-off should generate contemporaneous mafic magmas with arc and non-arc geochemical signatures (Cvetković et al., 2004; Whalen et al., 2006), but all of the Jurassic and Early Cretaceous gabbros and syenites and A-type granites are of within-plate origin without subduction-related magmatism (He et al., 2010; Jiang et al., 2015; Li et al., 2003; Li et al., 2013). Second, there are no known Jurassic adakitic rocks within South China that would be expected as a result of broken slab melting (Chen et al., 2008; He et al., 2010). Finally, Wang et al. (2007) argued that the Indosinian magmatism was related to collision between Indochina and South China rather than the

Paleo-Pacific subduction system, and the tectonic elements also indicated an N–S compression as a result of the northward collision of the Indochina Block in the Triassic (Zhang et al., 2009), which is inconsistent with the westward subduction of the Pacific plate initiated at ~250 Ma. More recently, a large amount of Early Jurassic A-type and mafic magmatism has been recognized in the southern Jiangxi and western Fujian provinces (He et al., 2010; Jiang et al., 2015; Li et al., 2003); their time-space distribution and geochemical features suggest a continental arc setting (Jiang et al., 2015). This finding is also supported by coeval shoshonitic rocks (Jiang et al., 2009), which typically occur in destructive plate margin settings and extensional environments (Eklund et al., 1998; Morrison, 1980; Müller et al., 1992; Rogers et al., 1998; Turner et al., 1996). Integrating new and previous data, Jiang et al. (2015) proposed a repeated slab advance-retreat model for the late Mesozoic magmatic evolution of southeast China.

The consensus of all models is that the South China Block has been under extensional environments since the Jurassic. The long-time extension led to the decompressional melting of underlying mantle sources to form abundant Jurassic basaltic, syenitic and A₂-type granitic rocks in the Nanling region (He et al., 2010; Jiang et al., 2006; Jiang et al., 2009; Jiang et al., 2015; Li et al., 2003; Li et al., 2004). In particular, the Late Jurassic Daoxian basalts show significant arc-like geochemical features, a finding that is compatible with the continental arc-rifting environment, as the A₂-type granites suggested (Jiang et al., 2009; Jiang et al., 2015). As discussed earlier, two stages of the Late Mesozoic volcanism in the Tianhuashan basin have been successfully identified, Daguding Fm. in the Late Jurassic and Ehuling Fm. in the Early Cretaceous, respectively. The temporal and spatial distribution of these volcanic rocks may provide an opportunity to verify the late Mesozoic tectonic evolution of South China. Daguding potassium-rich calc-alkali rhyolitic volcanic rocks are relatively enriched in LILE and depleted in HFSE with negative Nb, Ta and Ti anomalies, which shows the affinity of the volcanic rocks related to subduction (Thirlwall et al., 1994). Moreover, in the discrimination diagrams (Fig. 18, Pearce et al., 1984), Daguding volcanic rocks fall in the VAG field, similarly to the contemporaneous granitoids in the WZG region (Zhejiang, Jiangxi, and Anhui provinces), which have been interpreted as having formed in a volcanic arc related to the subduction of the Izanagi plate (Li et al., 2013). Thus, the Late Jurassic magmatism in the northeastern Jiangxi Province likely happened in a continental arc environment related to the subduction of the paleo-Pacific plate.

5.4.2. Early Cretaceous (approximately 130 Ma)

Although the tectonic regime accounting for the Jurassic (early Yanshanian) magmatism inland is enigmatic and controversial, there is a general consensus that the Cretaceous (late Yanshanian) magmatism along the coastal areas occurred in an active continental margin due to the subduction of the Paleo-Pacific oceanic plate.

As mentioned earlier, in addition to Ehuling volcanic rocks, a number of Early Cretaceous A₂-type granites have been identified along the Gan-Hang Belt and in the WZG region (Jiang et al., 2011; Li et al., 2013; Wong et al., 2009; Yang et al., 2012; Zhou et al., 2013). It has long been recognized that A-type magmas are mostly emplaced in extensional tectonic settings, including continental arc, back-arc extension, post-collisional extension and within-plate settings (Eby, 1992; Förster et al., 1997; Jiang et al., 2009; Jiang et al., 2011; Smith et al., 1999; Whalen et al., 1996). Eby (1992) suggested that the A₁ group seemed to be restricted to hotspots, plumes or continental

rift zones located in anorogenic settings, whereas the A₂ group granitoids might be associated with the waning stages of arc magmatism. In addition, all A₂-type granitoids show a WPG affinity in the tectonic environment discrimination diagram (Fig. 18, Pearce et al., 1984), which suggests an extensional geological setting. In consideration of the regional coeval intermediate-mafic rocks and the geological characteristics (Deng et al., 2014; Jiang et al., 2006; Jiang et al., 2009; Jiang et al., 2011; Li et al., 2013; Yang et al., 2011; Yang et al., 2012; Zhao et al., 2012), we favor the viewpoint that Ehuling volcanic rocks might have erupted at the terminal stage of continental arc, most likely at the beginning of the extension of back-arc or of the rifting of continental arc. It is noteworthy that the timing of the extension along the Gan-Hang Belt in the Cretaceous was as early as ca. 139 Ma in the Tianhuashan basin, which was later than the one that occurred along the south part of the Shi-Hang Zone (He et al., 2010; Jiang et al., 2009; Li et al., 2003). Integrating the previously published data, it can also be concluded that the extension along the Gan-Hang Belt lasted from 139 Ma to 122 Ma in the Early Cretaceous (Jiang et al., 2011; Yang et al., 2011; Yang et al., 2012).

In summary, the Late Jurassic Daguding volcanic rocks in the northeastern Jiangxi Province formed in a continental arc environment related to the subduction of the Paleo-Pacific plate during the Late Jurassic (~160 Ma; Fig. 19a). Since the Early Cretaceous (~130 Ma), an intra-arc rift has formed or back-arc extension has developed along the Gan-Hang Belt (or Gan-Hang rift) as a consequence of slab rollback. This finding is further supported by recent geochronological and geochemical studies that have indicated that the regional extension gradually migrated from the inland area to the coastal area due to the progressive slab rollback of the Paleo-Pacific plate (Deng et al., 2014; Jiang et al., 2009; Jiang et al., 2011; Su et al., 2014; Wong et al., 2009; Yang et al., 2012). With ongoing extension during slab rollback, the crust and lithospheric mantle became progressively thinner. The upwelling of the asthenosphere and the underplating of mantle-derived magmas triggered crustal rocks to partially melt and generate hot magmas. The partial melting of granulitized Proterozoic metamorphic basement rocks (including both parametamorphic and orthometamorphic rocks) in the thinned crust close to the Gan-Hang Belt formed Ehuling A-type magmas (Fig. 19b).

6. Conclusions

- 1) LA-ICP-MS zircon U–Pb dating indicates that the Late Mesozoic volcanism in the Tianhuashan basin of the northeastern Jiangxi Province is episodic. The early volcanism took place in the Late Jurassic (~160 Ma), forming the Daguding Fm., and the late volcanism erupted in the Early Cretaceous (~130 Ma), forming the Ehuling Fm.
- 2) Daguding volcanic rocks show an association of rhyolite and have high K₂O contents. They are also strongly peraluminous and belong to the S-type. All rocks are characterized by negative εNd(t) and εHf(t) values and Paleoproterozoic T_{DM2} ages, implying that they were most likely derived from the Paleoproterozoic metasedimentary basement.
- 3) Ehuling volcanic rocks are primarily weakly peraluminous and show an A₂ subtype affinity. Geochemical and isotopic data suggest that they were formed by the partial melting of the granulitized Paleoproterozoic metamorphic basement (including parametamorphic and orthometamorphic rocks) at a high temperature and followed by the fractional crystallization of biotite, plagioclase and K-feldspar.

4) Detailed geochronological and geochemical data for the Late Jurassic to Early Cretaceous volcanic rocks in the northeastern Jiangxi Province imply that during the Late Jurassic, SE China on the Gan-Hang Belt was a continental arc related to the subduction of the Paleo-Pacific plate. Since the beginning of Early Cretaceous, an intra-arc rift has formed along the Gan-Hang Belt as a consequence of slab rollback. It can also be concluded that the extension in the Gan-Hang Belt began later than that in the southwestern part of the Shi-Hang Zone in South China and lasted from 139 Ma to 122 Ma.

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Figure captions:

Fig. 1. Geological sketch map showing the Gan-Hang Belt in Southeast China (modified from [Yu et al., 2006](#)).

Fig. 2. Geological map of the Lengshuikeng ore district (modified from the 912 Geological Team, Bureau of Geology and Mineral Resources of Jiangxi Province).

Fig. 3. Photomicrographs of representative volcanic rocks from the Lengshuikeng district (crossed nicols). (a) crystal tuff from the Daguding Fm., 150-245; (b) crystal tuff from the Daguding Fm., 701-702; (c) crystal tuff from the Daguding Fm., 701-840; (d) rhyolitic crystal ignimbrite from the Ehuling Fm., LS69; (e) lithic crystal ignimbrite from the Ehuling Fm., LS79; (f) rhyolitic crystal ignimbrite from the Ehuling Fm., 701-246; Qz—quartz.

Fig. 4. CL images of representative zircons from volcanic rocks of the Lengshuikeng district. Smaller solid circles indicate the spots of U–Pb dating, whereas larger dashed circles indicate the spots of Hf isotope analyses. Age and $\epsilon_{\text{Hf}}(t)$ values are also shown for each spot. All scale bars represent 100 μm .

Fig. 5. Zircon U–Pb Concordia diagrams and weighted average $^{206}\text{Pb}/^{238}\text{U}$ ages for volcanic rocks of the Daguding Fm. (a, b, c, d, e, f) and the Ehuling Fm. (g, h, i, j, k).

Fig. 6. Histograms of $\epsilon_{\text{Hf}}(t)$ values of zircons from volcanic rocks of the Daguding Fm. and the Ehuling Fm.

Fig. 7. T_{DM2} distribution of zircons from volcanic rocks of northern Wuyi Mountain Belt and U–Pb age spectra of detrital zircon populations ([Xu et al., 2007](#)) from the eastern Cathaysia Block. The shaded region indicates the ages of the Paleoproterozoic granitoids and coeval metamorphic rocks ([Chen and Jahn, 1998](#); [Li et al., 1996](#); [Li, 1997](#); [Li et al., 2010](#); [Xu et al., 2005](#); [Xu et al., 2007](#); [Yu et al., 2009](#); [Yuan et al., 1991](#)).

Fig. 8. Total alkali-silica (TAS) diagram of volcanic rocks from the Daguding and Ehuling Fms. ([Le Bas et al., 1986](#)).

Fig. 9. SiO_2 vs. ASI [=molar $\text{Al}_2\text{O}_3/(\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O})$] (a), K_2O (b) and $\text{Mg\#}$ [=Mg/(Mg+FeT)] (c) diagrams for the Daguding and Ehuling volcanic rocks. In (c), also shown are the fields of pure crustal partial melts obtained in experimental studies by the dehydration melting of low-K basaltic rocks at 8-16 kbar and 1000-1050 °C ([Rapp and Watson, 1995](#)), of moderately hydrous (1.7-2.3 wt% H_2O) medium- to high-K basaltic rocks at 7 kbar and 825-950 °C ([Sisson et al., 2005](#)), of pelitic rocks at 7-13 kbar and 825-950 °C ([Patiño Douce and Johnston, 1991](#)) and of biotite gneiss and quartz amphibolite at 7-15 kbar and 900-1000°C ([Patino Douce and Beard, 1995](#)). The symbols are the same as those in Fig. 8.

Fig. 10. AR vs. SiO₂ diagram (Wright, 1969) for volcanic rocks of the Daguding and Ehuling Fms., where AR (alkalinity ratio) = $[\text{Al}_2\text{O}_3 + \text{CaO} + (\text{Na}_2\text{O} + \text{K}_2\text{O})] / [\text{Al}_2\text{O}_3 + \text{CaO} - (\text{Na}_2\text{O} + \text{K}_2\text{O})]$ (wt.%). The symbols are the same as those in Fig. 8.

Fig. 11. (a) Chondrite-normalized REE patterns (normalization values are from Boynton, 1984) and (b) primitive-mantle-normalized multi-element spidergrams (normalization values are from McDonough and Sun, 1995) for volcanic rocks of the Daguding Fm and the Ehuling Fm.

Fig. 12. (a) Zr vs. 10000×Ga/Al, (b) FeO*/MgO vs. (Zr+Nb+Ce+Y) discrimination diagrams (after Whalen et al., 1987), showing the A-type nature of the Ehuling volcanic rock, and (c) the A-type granite subdivision diagrams of Nb–Y–3×Ga and (d) Rb/Nb vs. Y/Nb (after Eby, 1992), showing that the Ehuling volcanic rocks belong to the A₂ group. FG: fractionated felsic granites; OGT: unfractionated I-, S- and M-type granites. The symbols are the same as those in Fig. 8.

Fig. 13. Initial ⁸⁷Sr/⁸⁶Sr (I_{Sr}) vs. ε_{Nd}(t) diagram for the volcanic-intrusive complex and mafic microgranular enclaves from the Xiangshan. The Sr and Nd isotope data for orthometamorphic and parametamorphic rocks are from Yuan et al. (1991) and Hu et al. (1999). The initial ⁸⁷Sr/⁸⁶Sr ratios (I_{Sr}) and ε_{Nd}(t) values were calculated based on respective zircon U–Pb ages determined in this study. The symbols are the same as those in Fig. 8.

Fig. 14. (a) Columnar section of ZK15701 showing the sampling horizons and corresponding ages of this work and (b) geochronological results of previous work and this work and their distributions. Previous dating results of the volcanic rocks are shown in Table 4.

Fig. 15. ACF diagrams showing the relationship between the chemistry and mineralogy of Daguding volcanic rocks. The hatched area of the diagram corresponds to peraluminous rock compositions (Chappell and White, 1992).

Fig. 16. Diagram of ε_{Hf}(t) vs. U–Pb ages of zircons from the volcanic rocks in the Tianhuashan basin with the shaded region indicating Hf isotope evolution for the crustal basement of the eastern Cathaysia Block after Xu et al. (2007).

Fig. 17. Eu vs. Rb (a), Ba (b), Sr (c) and Yb (d) diagrams for Ehuling A-type volcanic rocks. Trends illustrate crystal fractionation of biotite (Bt), plagioclase (Pl) and K-feldspar (Kfs) starting with initial concentration (701–258). Tic marks along these arrows indicate the weight fraction of fractionation melt. The partition coefficient data for minerals in equilibrium with silicic liquids are from Nash and Crecraft (1985).

Fig 18. (a) Y–Nb and (b) Yb–Ta diagram (Pearce et al., 1984) for volcanic rocks, VAG: volcanic arc granites. The contemporaneous granitoids in the WZG region are also plotted for comparison (Li et al., 2013). WPG: within plate granites, ORG: ocean ridge granites. The symbols are the same as those in Fig. 8.

Fig. 19. Schematic diagram showing the Late Mesozoic tectonic evolution and magma generation processes in the Tianhuashan basin. The Late Jurassic Daguding volcanic rocks were formed in a

continental arc environment associated with the subduction of the Paleo-Pacific plate (a). The Early Cretaceous Ehuling volcanic rocks were formed in a continental arc-rifting setting as a consequence of slab rollback (b). See text for more discussion. Revised after [Zhou and Li \(2000\)](#) and [Jiang et al. \(2015\)](#)

Table captions:

Table 1 Strata in the Lengshuikeng district (modified from the 912 Geological Team, Bureau of Geology and Mineral Resources of Jiangxi Province).

Table 2 Lithologies, pyroclast assemblages and sample locations of the Late Mesozoic volcanic rocks from the Lengshuikeng district.

Table 3 Major (wt.%), trace elements (ppm) and Sr-Nd isotope compositions for volcanic rocks of the Daguding and Ehuling Fms.

Table 4 Compilation of isotope age for the Late Mesozoic volcanism in the northeastern Jiangxi Province.

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Table 1

Table 1 Strata in the Lengshuikeng district (modified from the 912 Geological Team, Bureau of Geology and Mineral Resources of Jiangxi Province).

System	Formation	Stage	Geological code	Thickness (m)	Major rock types
Quaternary			Q	0-20	Alluviation, slope wash
Jurassic	Ehuling	Third member	J_3e^3	160-1430	Rhyolitic ignimbrite with tuff at the bottom
		Second member	J_3e^2	140-550	Rhyolite, ignimbrite, with tuffaceous siltstone at the bottom
		First member	J_3e^1	180-420	Rhyolitic crystal tuff, ignimbrite with tuffaceous sandstone at the bottom
	Daguding	Upper member	J_3d^2	30-170	Andesite, tuff breccia
		Lower member	J_3d^1	100-310	Rhyolitic crystal tuff, rhyolitic ignimbrite, tuff
Carboniferous	Huanglong		C_2h	>170	Limestone
	Zishan		C_{1z}	30-150	Quartz conglomerate, sandstone
Sinian	Laohutang	Third member	Z_2l^3	>480	Biotite schist, quartz biotite schist
		Second member	Z_2l^2	410	K-feldspar gneiss, striped migmatite
		First member	Z_2l^1	>370	Quartzite, quartz schist
	Xiafang				
			Z_2x	>500	Biotite plagioclase gneiss, biotite quartz schist, quartzite

Table 2

Table 2 Lithologies, pyroclast assemblages and sample locations of the Late Mesozoic volcanic rocks from Lengshuikeng district					
Formation	Sample no.	Colour	Lithology	Pyroclast assemblages	Location
Ehuling	LS68-LS76	Flesh pink	Crystal ignimbrite	Crystals (15%,Qz and few Pl)+lithoclasts (10%)+volcanic ash	N27°54'7.86";E117°12'19.93"
	LS77-LS83	Gray-green	Lithic crystal ignimbrite	Crystals (15%, Qz and few pl)+lithoclasts (10%)+volcanic ash	N27°54'13.28";E117°12'21.94"
	701-246	Gray-green	Rhyolitic crystal ignimbrite	Crystals (10%,Qz and few Pl)+vitroclastic+volcanic ash	ZK15701 246m
	701-254	Gray-green	Rhyolitic crystal ignimbrite	Crystals (10%,Qz and few Pl)+vitroclastic+volcanic ash	ZK15701 254m
	701-637	Gray-green	Crystal ignimbrite	Crystals (5%,Qz and few Pl)+vitroclastic+volcanic ash	ZK15701 637m
Daguding	LS45	Gray-green	Crystal tuff	Crystals (60%, Qz, Pl and few Bt)+volcanic ash	N27°54' 43.8 " ; E117°12' 24.8 "
	150-245	Gray-green	Crystal tuff	Crystals (50%, Qz, Pl and few Bt)+volcanic ash+vitroclasts	ZK15150 245m
	150-380	Gray-green	Crystal tuff	Crystals (45%, Qz, Pl and few Bt)+volcanic ash+vitroclasts	ZK15150 380m
	701-700	Gray-green	Crystal tuff	Crystals (60%, Qz, Pl and Bt)+volcanic ash	ZK15701 700m
	701-728	Gray-green	Crystal tuff	Crystals (55%, Qz, Pl and few Bt)+volcanic ash	ZK15701 728m
	701-840	Gray-green	Crystal tuff	Crystals (50%, Qz, Pl and few Bt)+volcanic ash+vitroclasts	ZK15701 840m

Table 3

Table 3 Major (wt.%), trace elements (ppm) and Sr-Nd isotope compositions for volcanic rocks of the Daguding and Ehuling Fm.

Formation	Ehuling Fm.												
Sample	701-246	701-251	701-254	701-258	701-270	701-286	701-293	701-637	701-645	701-656	701-662	701-672	701-684
SiO ₂	74.34	74.27	76.06	75.29	76.03	74.68	74.97	75.57	75.07	75.23	75.44	75.83	74.86
TiO ₂	0.12	0.11	0.13	0.14	0.13	0.15	0.14	0.12	0.1	0.12	0.09	0.09	0.09
Al ₂ O ₃	12.38	14.1	12.19	12.34	12.46	12.16	12.65	12.24	12.38	12.79	12.57	12.38	12.38
Fe ₂ O ₃ T	2.5	1.6	1.93	1.75	1.69	2.41	2.03	2.08	1.88	1.87	1.7	1.67	1.96
MnO	0.1	0.03	0.04	0.06	0.07	0.11	0.05	0.1	0.07	0.06	0.04	0.08	0.1
MgO	0.2	0.25	0.15	0.18	0.11	0.23	0.12	0.1	0.12	0.15	0.2	0.12	0.16
CaO	1.21	0.4	0.41	0.95	0.88	0.95	0.61	0.71	0.59	0.57	0.35	0.99	0.82
Na ₂ O	2.17	2.06	1.91	2.13	3.11	2.41	2.65	3.2	3.09	3.09	3.17	3.31	2.88
K ₂ O	5.2	5.63	5.17	5.63	4.07	5.25	5.46	5.25	5.28	5.27	5.32	5.22	5.27
P ₂ O ₅	0.01	0.02	0.03	0	0.02	0.02	0.01	0	0.03	0.01	0.01	0.02	0.02
LOI	1.73	1.13	1.16	1.52	1.47	1.34	1.02	0.55	0.64	0.77	0.73	1.04	0.97
Total	99.96	99.61	99.19	99.99	100.03	99.72	99.71	99.91	99.25	99.93	99.63	100.75	99.51
A.R	3.37	3.26	3.57	3.81	3.33	3.81	4.15	4.76	4.64	4.34	4.83	4.52	4.23
ASI	1.09	1.38	1.29	1.09	1.12	1.07	1.11	1	1.04	1.08	1.08	0.96	1.04
Mg#	0.14	0.24	0.13	0.17	0.11	0.16	0.1	0.09	0.11	0.14	0.19	0.12	0.14
V	1.52	1.11	1.17	0.61	1.3	0.99	0.98	1.36	0.89	0.54	0.41	0.43	0.3
Cr	25.49	12.07	8.75	4.78	12.42	9.13	11.03	23.46	10.57	14.73	21.2	20.62	14.67
Ni	4.9	1.66	1.46	1.09	2.22	1.96	2.22	4.74	1.84	2.59	3.53	3.67	2.53
Ga	18.84	17.75	17.44	18.78	19.23	18.8	17.19	18.74	18.28	19.1	18.1	17.9	18.01
Rb	315.6	342.5	358.4	383.3	259.9	336.0	358.8	337.8	341.4	336.1	364.7	357.8	343.3
Sr	64.62	40.02	48.54	63.35	44.99	47.46	20.45	35.4	34.9	32.79	30.78	43.61	53.63
Y	43.68	45.13	43.36	44.57	39.29	45.76	25.15	44.14	44.48	42.81	43	48.75	49.16

Zr	269	275	276	260	267	259	271	248	252	239	230	197	238
Nb	31.65	31.86	31.98	33.27	30.44	30.24	31.09	32.39	31.3	31.25	31.67	32.61	30.33
Ba	111.1	98.3	87.76	169	42.55	61.2	15.87	39.17	52.22	14.76	25.92	9.92	19.31
La	114.8	129.6	129.35	141.77	114.8	125.2	82.4	94.77	83.81	62.13	57.15	53.24	57.83
Ce	159.1	170.1	190.2	205.8	203.3	181.2	114.1	159.8	146.3	125.1	113.2	81.79	96.83
Pr	22.59	24.43	23.88	27.12	22.16	24.26	14.52	20.72	19.3	15.62	14.79	13.46	15.42
Nd	77.95	83.15	81.57	93.83	76.66	83.88	47.93	73.36	69.01	57.32	54.62	51.12	57.61
Sm	11.44	11.78	11.47	13.69	11.04	12.15	6.38	11.5	11.12	9.97	9.71	10.01	10.37
Eu	0.31	0.29	0.27	0.34	0.27	0.29	0.15	0.2	0.2	0.13	0.12	0.11	0.13
Gd	10.16	10.6	10.35	12.28	9.75	10.83	5.77	9.97	9.61	8.66	8.42	8.78	8.93
Tb	1.34	1.35	1.28	1.58	1.21	1.39	0.71	1.33	1.3	1.22	1.2	1.34	1.29
Dy	6.75	6.78	6.13	7.77	5.9	6.87	3.47	6.75	6.62	6.42	6.31	7.34	6.86
Ho	1.27	1.25	1.13	1.42	1.09	1.27	0.65	1.25	1.24	1.2	1.19	1.39	1.29
Er	3.91	3.79	3.42	4.32	3.34	3.84	2.02	3.74	3.69	3.57	3.51	4.08	3.83
Tm	0.56	0.52	0.47	0.6	0.46	0.54	0.29	0.52	0.52	0.51	0.5	0.59	0.54
Yb	3.66	3.43	3.09	3.92	3.04	3.52	1.9	3.4	3.38	3.34	3.21	3.86	3.57
Lu	0.57	0.53	0.47	0.6	0.47	0.54	0.3	0.51	0.51	0.51	0.49	0.58	0.54
Hf	8.58	8.77	8.89	8.56	8.49	8.36	8.49	8.28	8.48	8.28	8.22	7.48	8.34
Ta	1.6	1.63	1.64	1.75	1.58	1.57	1.61	1.66	1.65	1.65	1.71	1.78	1.68
Pb	32.58	26.55	16.1	36.36	19.33	35.5	24.4	44.72	50.39	49.3	45.75	46.41	41.46
Th	27.43	28.8	29.88	35.76	30.52	29.48	20.31	30.39	29.21	29.99	27.39	29.31	28.45
U	3.16	2.85	2.58	3.25	6.7	3.14	1.89	3.8	3.58	4.16	3.49	3.61	3.4
ΣREE	414.45	447.59	463.06	515.03	453.57	455.77	280.54	387.81	356.61	295.67	274.4	237.69	265.04
(La/Yb) _N	22.48	27.09	29.98	25.92	27.08	25.51	31.18	20	17.78	13.35	12.77	9.9	11.6
Eu/Eu*	0.09	0.08	0.08	0.08	0.08	0.08	0.07	0.06	0.06	0.04	0.04	0.04	0.04
⁸⁷ Rb/ ⁸⁶ Sr	14.131	24.765	21.366	17.507	16.718	20.488	50.766	27.608	28.309	29.664	34.280	23.737	18.525

$^{87}\text{Sr}/^{86}\text{Sr}\pm 2\sigma$	0.744755±10			0.747668±8	0.744256±12								
$(^{87}\text{Sr}/^{86}\text{Sr})_i$	0.719189			0.714498	0.71258								
$^{147}\text{Sm}/^{144}\text{Nd}$	0.0887	0.0856	0.085	0.0882	0.087	0.0875	0.0804	0.0947	0.0974	0.1051	0.1074	0.1183	0.1088
$^{143}\text{Nd}/^{144}\text{Nd}\pm 2\sigma$	0.512065±3	0.512070±4	0.512090±4	0.512060±4	0.512060±6	0.512079±6	0.512058±6	0.512088±6	0.512066±4	0.512105±4	0.512092±4	0.512081±4	0.512076±3
$\epsilon_{\text{Nd}}(\text{t})$	-9.4	-9.2	-8.8	-9.4	-9.4	-9	-9.3	-8.9	-9.4	-8.8	-9.1	-9.5	-9.4
$T_{\text{DM2}} (\text{Ma})$	1689	1675	1641	1694	1693	1663	1687	1658	1697	1646	1670	1702	1697
M	1.28	1.02	1.05	1.27	1.23	1.3	1.25	1.4	1.34	1.3	1.29	1.46	1.35
$T_{\text{Zr}}(^{\circ}\text{C})$	843	867	865	841	847	838	846	826	832	830	827	800	826

Formation	Daguding Fm.									
Sample	150-245	150-380	701-700	701-728	701-770	701-796	701-817	701-831	701-840	701-852
SiO ₂	74.78	70.88	74.94	73.59	72.61	73.72	72.5	72.79	74.15	76.25
TiO ₂	0.28	0.31	0.3	0.3	0.25	0.26	0.31	0.29	0.18	0.12
Al ₂ O ₃	12.87	12.51	13.42	13.11	12.94	12.81	12.96	12.98	12.9	11.86
Fe ₂ O ₃ T	1.79	5.04	2.12	2.56	1.84	1.7	2.13	1.72	1.52	1.05
MnO	0.2	0.97	0.09	0.25	0.15	0.09	0.16	0.14	0.14	0.36
MgO	0.43	0.4	0.53	0.45	0.46	0.67	0.59	0.53	0.52	0.45
CaO	0.65	0.15	0.06	0.13	1.56	1.23	1.32	1.83	1.42	0.66
Na ₂ O	1.13	0.16	0.52	0.77	1.52	2.37	2.06	1.96	1.01	0.31
K ₂ O	5.24	6.95	5.86	6.68	5.54	4.87	4.71	4.75	5.33	6.03
P ₂ O ₅	0.06	0.08	0.07	0.07	0.06	0.06	0.08	0.07	0.03	0.02
LOI	2.11	2.42	1.58	1.87	2.23	1.89	2.24	2.56	2.4	1.95
Total	99.53	99.87	99.49	99.78	99.16	99.67	99.06	99.61	99.61	99.05
A.R	2.78	3.56	2.8	3.57	2.9	3.13	2.8	2.66	2.59	3.05

ASI	1.48	1.55	1.84	1.5	1.14	1.12	1.19	1.11	1.29	1.44
Mg#	0.32	0.14	0.33	0.26	0.33	0.44	0.35	0.38	0.4	0.46
V	19.2	31.96	23.96	32.83	27.97	27.17	28.7	26.2	20.73	13.22
Cr	8.99	12.64	11.21	34.77	11.17	9.99	26.4	20.56	32.86	8.26
Ni	2.0	5.78	2.53	6.48	2.8	2.36	6.04	4.24	6.95	1.42
Ga	16.59	19.97	14.99	19.95	20.26	18.36	20.3	20.52	19.2	16.93
Rb	284.4	402.3	395.6	406.4	254.0	220.4	229.2	245.7	335.3	373.8
Sr	69.2	72.51	67.82	46.72	111.1	124.3	86.54	128.9	70.51	43.86
Y	23.66	16.08	20.67	15.58	17.9	21.84	20.04	26.36	23.08	21.77
Zr	134	161	159	151	138	116	150	154	132	93.68
Nb	24.71	20.1	22.37	19.4	20.84	20.54	23.91	24.82	28.43	27.77
Ba	619	904	400	539	548	636	599	537	363	366
La	54.1	55.44	52.91	39.04	36.53	37.47	45.43	53.4	29.98	22.92
Ce	96.29	89.92	92.1	69.49	74.45	65.49	82.69	92.88	60.1	44.14
Pr	10.27	10.23	10.28	8.88	9.53	9.45	10.86	12.79	7.19	5.55
Nd	36.04	35.27	36.15	25.55	28.82	29.35	33.75	41.49	24.11	18.87
Sm	6.23	5.31	5.71	4.04	5.15	5.31	5.92	7.38	5.11	4.33
Eu	0.96	1.08	0.8	0.67	0.77	0.88	0.9	1.02	0.6	0.45
Gd	5.54	4.81	5.11	3.56	4.58	4.83	5.22	6.58	4.75	4.15
Tb	0.75	0.64	0.67	0.45	0.66	0.68	0.73	0.92	0.77	0.7
Dy	3.74	3.15	3.31	2.18	3.39	3.57	3.62	4.74	4.32	4.08
Ho	0.68	0.56	0.59	0.38	0.62	0.66	0.65	0.85	0.81	0.77
Er	1.99	1.61	1.74	1.12	1.76	1.87	1.84	2.42	2.35	2.22
Tm	0.28	0.21	0.23	0.15	0.24	0.25	0.25	0.32	0.34	0.32
Yb	1.83	1.39	1.52	0.99	1.58	1.65	1.62	2.15	2.18	2.07
Lu	0.28	0.21	0.23	0.15	0.23	0.24	0.24	0.32	0.33	0.31

Hf	4.64	5.24	5.38	4.04	3.89	3.37	4.29	4.59	4.37	3.46
Ta	1.49	1.15	1.4	1.13	1.34	1.27	1.41	1.56	1.97	2.06
Pb	17.67	70.68	218	106	50.62	21.49	196	98.89	33.68	39.97
Th	22.33	15.71	21.5	16.36	21.77	23.74	21.38	29.83	25.5	24.56
U	2.48	4.26	3.18	2.12	1.97	1.7	1.69	2.95	2.84	5.08
ΣREE	218.99	209.82	211.35	156.65	168.31	161.7	193.72	227.26	142.94	110.88
(La/Yb) _N	21.2	28.52	25.01	28.29	16.58	16.29	20.12	17.82	9.86	7.94
Eu/Eu*	0.5	0.65	0.45	0.54	0.48	0.53	0.49	0.45	0.37	0.32
⁸⁷ Rb/ ⁸⁶ Sr	11.893	16.052	16.879	25.168	6.616	5.131	7.663	5.515	13.761	24.661
⁸⁷ Sr/ ⁸⁶ Sr±2σ					0.725540±6	0.720365±11	0.723483±8	0.720365±11		
(⁸⁷ Sr/ ⁸⁶ Sr) _i					0.710594	0.708774	0.706174	0.707908		
¹⁴⁷ Sm/ ¹⁴⁴ Nd	0.1044	0.091	0.0954	0.0955	0.108	0.1093	0.106	0.1075	0.1281	0.1386
¹⁴³ Nd/ ¹⁴⁴ Nd±2σ	0.512033±5	0.512025±4	0.512046±5	0.512028±4	0.512027±4	0.512037±4	0.512025±4	0.512072±6	0.512046±5	0.512063±4
ε _{Nd} (t)	-10	-9.9	-9.5	-9.9	-10.1	-10	-10.1	-9.2	-10.2	-10
T _{DM2} (Ma)	1757	1747	1721	1749	1770	1757	1770	1699	1774	1763
M	0.92	0.91	0.74	0.93	1.23	1.25	1.18	1.27	1.07	0.92
T _{Zr} (°C)	807	824	837	817	786	770	797	792	794	775

Fe₂O₃^T, total Fe; Eu/Eu* is a measure of the Eu anomaly when compare to Sm and Gd. Eu/Eu*= Eu_N/(Sm_N*Gd_N)^{1/2}; N = Chondrite normalized. T_{Zr}=12,900/(2.95+0.85M+lnD^{Zr, zircon/melt})(Watson and Harrison, 1983), where D^{Zr, zircon/melt} is the ratio of Zr concentrations (ppm) in zircon to that in the saturated melt, M=cation ratio (Na+K+2·Ca)/(Al · Si). The geothermometer is calibrated for M=0.9 to 1.7.

Table 4

Table 4 Compilation of isotope age for the Late Mesozoic volcanism in the northeastern Jiangxi.

Province	Location	Formation	Method	Age(Ma)	Reference
Jiangxi	Tianhuashan	Ehuling	Rb-Sr	101.3±4.1	Xu et al., 2001
Jiangxi	Tianhuashan	Ehuling	K-Ar	102	Xu et al., 2001
Jiangxi	Tianhuashan	Ehuling	Rb-Sr	129.7±8.9	Xu et al., 2002
Jiangxi	Tianhuashan	Ehuling	LA-ICP-MS	161.8±1.4	Xu et al., 2013
Jiangxi	Tianhuashan	Ehuling	LA-ICP-MS	146.6±2.2	Qiu et al., 2013
Jiangxi	Tianhuashan	Ehuling	LA-ICP-MS	140±1	Su et al., 2014
Jiangxi	Tianhuashan	Ehuling	LA-ICP-MS	137±1	Su et al., 2014
Jiangxi	Tianhuashan	Ehuling	SHRIMP U-Pb	157.8±1.6	Meng et al., 2012
Jiangxi	Tianhuashan	Ehuling	SHRIMP U-Pb	157.2±1.5	Meng et al., 2012
Jiangxi	Tianhuashan	Ehuling	SHRIMP U-Pb	158.2±1.8	Meng et al., 2012
Jiangxi	Huanggangshan	Ehuling	SHRIMP U-Pb	131±2	Qiu et al., 2011
Jiangxi	Huanggangshan	Ehuling	SHRIMP U-Pb	138±1	Qiu et al., 2011
Jiangxi	Shengyuan	Ehuling	Ar-Ar	122.6±1.0	Zhang, 2000
Jiangxi	Shengyuan	Ehuling	K-Ar	127±6	Zhang, 2000
Jiangxi	Xiangshan	Ehuling	Single-zircon U-Pb	129.5±7.9	Zhang and Li, 2007
Jiangxi	Xiangshan	Ehuling	Rb-Sr	136±3	Zhang, 1996
Jiangxi	Xiangshan	Ehuling	Rb-Sr	140.7	Zhang, 1996
Jiangxi	Xiangshan	Ehuling	Ar-Ar	141.2±1.5	Zhang, 1996
Jiangxi	Xiangshan	Ehuling	LA-ICP-MS	134.1±1.6	Yang et al., 2013
Jiangxi	Xiangshan	Ehuling	LA-ICP-MS	135.0±2.0	Yang et al., 2013
Jiangxi	Shixi	Ehuling	Ar-Ar	141.8	Zhang, 1996
ZheJiang	Xujiadun	Ehuling	SHRIMP U-Pb	138.0±2.4	Zhang et al., 2009
Jiangxi	Tianhuashan	Ehuling	LA-ICP-MS	130.4±1.1	This study

Jiangxi	Tianhuashan	Ehuling	LA-ICP-MS	133.3±1.2	This study
Jiangxi	Tianhuashan	Ehuling	LA-ICP-MS	137.5±1.2	This study
Jiangxi	Tianhuashan	Ehuling	LA-ICP-MS	137.5±1.3	This study
Jiangxi	Tianhuashan	Ehuling	LA-ICP-MS	138.4±1.3	This study
Jiangxi	Tianhuashan	Daguding	K-Ar	112-123	Xu et al., 2001
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	160.8±1.9	Qiu et al., 2013
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	161.0±1.0	Wang et al., 2013
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	144±1	Su et al., 2014
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	142±1	Su et al., 2014
Jiangxi	Shengyuan	Daguding	Rb-Sr	141.2±9.9	Zhang, 2000
Jiangxi	Shengyuan	Daguding	Rb-Sr	138±8.8	Wu et al., 2011
Jiangxi	Xiangshan	Daguding	LA-ICP-MS	136.4±1.5	Yang et al., 2013
Jiangxi	Xiangshan	Daguding	LA-ICP-MS	136.8±2.5	Yang et al., 2013
Jiangxi	Xiangshan	Daguding	LA-ICP-MS	137.3±0.9	Yang et al., 2013
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	151.6±4.1	This study
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	155.0±2.2	This study
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	155.9±1.8	This study
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	158.3±1.7	This study
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	158.9±1.5	This study
Jiangxi	Tianhuashan	Daguding	LA-ICP-MS	159.4±2.4	This study