K-Ar dating, geochemical and Sr-Nd-Pb isotopic systematics of Paleocene mafic rocks in Central Jiangxi, SE China: Evidence for lithosphere replacement

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Geochemical and isotopic studies were carried out on Paleocene mafic rocks in the Taihe Basin, central Jiangxi, Southeast China, in order better to understand their magma sources and tectonic implications. K-Ar dating results show that these mafic rocks intruded during the Paleocene (50–65 Ma). These Paleocene mafic rocks are porphyritic diabase and have similar geochemical features, such as strong enrichment of large ion lithophile element (LILE, e.g., Rb, Ba, Sr), slight enrichment of light rare earth element (LREE) and high field strength element (HFSE, e.g., Nb, Ta, Ti, P). The mafic rocks are also characterized by high rare earth element (REE) and minor Eu anomalies ($\delta\text{Eu} = 0.88\text{--}1.2$) and have uniform initial $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7041–0.7064) and Pb isotopic composition ($^{206}\text{Pb}/^{204}\text{Pb} = 18.338\text{--}18.677$) as well as a relatively wide range of initial Nd isotopic ratios ($\epsilon_{\text{Nd}}(\text{T})$) varying from +0.8 to +6.2. These geochemical characteristics are different from those of subduction-related basalts, but similar to oceanic island basalts (OIB). Geochemical and isotopic evidence suggests that mafic rocks in the Taihe Basin were not significantly affected by crustal contamination and originated from asthenospheric mantle with a minor involvement of EM 2 mantle component. It is noted that Cretaceous basaltic rocks in Southeast China show enriched geochemical and isotopic compositions and were derived from a highly heterogeneous lithospheric mantle with minor asthenosphere components. On the other hand, Late Cenozoic basalts in Southeast China, which display OIB-like geochemical characteristics, were probably generated by melting of depleted asthenospheric mantle (DMM or MORB) with an involvement of EM 2 lithospheric mantle and a paucity of EM 1 lithospheric mantle. The Paleocene mafic rocks of the Taihe Basin exhibit transitional geochemical features between Cretaceous and Late Cenozoic basaltic magmatism, indicating that the involvement of asthenospheric mantle beneath Southeast China increased with time from Late Mesozoic to Cenozoic. It is suggested that the lithosphere replacement beneath Southeast China was associated with lithosphere extension and thinning, with decompression melting of the upwelling asthenosphere having taken place during the Paleocene period. Lithosphere replacement might be related to subduction roll-back or/and steepening of Paleo-Pacific plate due to injection of asthenospheric material into the mantle wedge beneath Southeast China.

Keywords: mafic rocks, Sr-Nd-Pb, lithosphere replacement, Paleocene, Central Jiangxi

INTRODUCTION

The Late Mesozoic and Cenozoic in Southeast China were characterized by intensive and widespread magmatism. Volumetrically dominant granitoids and rhyolites, with minor basaltic rocks were emplaced between the Jurassic and the Cretaceous (Zhou and Li, 2000), whereas predominantly basaltic rocks of Late Cenozoic ages were erupted in the coastal region of South-

east China (e.g., Zou *et al.*, 2000; Ho *et al.*, 2003) (Fig. 1a). Recent studies have indicated that magma sources responsible for the Late Mesozoic magmatism are different from those of the Late Cenozoic basalts in Southeast China (Xie *et al.*, 2001, 2005a). Studies of mantle-derived xenoliths suggest that sub-continental lithospheric mantle beneath Southeast China is of Phanerozoic age (Zheng *et al.*, 2004) and that ancient lithospheric mantle, up to 200 km thickness, had been removed and replaced by thinner, hotter and more fertile mantle material between the Late Mesozoic and the Cenozoic (Xu *et al.*, 2000, 2002). This replacement might not have been complete throughout the region (Xu *et al.*, 2002), because relicts of ancient lithospheric mantle (Middle Proterozoic,

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Fig. 1. (a) Spatial distribution of Late Mesozoic and Cenozoic basaltic magmatism in Southeast China (modified from Wang *et al.*, 2003; Xu and Xie, 2005; Xie *et al.*, 2006): ① - Jiangshan-Shaoxing fault zone; ② - Chenxian-Linwu fault zone. 1 - Xuanwudi basalt (110.3 Ma, Rb-Sr isochron age, Yu *et al.*, 1993); 2 - Xinchang basalt (4.9–9.4 Ma, ^{40}Ar - ^{39}Ar plateau ages, Ho *et al.*, 2003); 3 - Longyou basalt (Late Cenozoic, Xu and Xie, 2005); 4 - Jingling basalt (113 Ma, Rb-Sr isochron age, Xing *et al.*, 2004); 5 - Xiangshan basalt (120 Ma, Rb-Sr isochron age, Xu and Xie, 2005); 6 - Yushan basalt (98–102 Ma, K-Ar ages, Wang *et al.*, 2002); 7 - Guangtong Formation basalt (110 Ma, Rb-Sr isochron age, Zhou *et al.*, 1993); 8 - Wuyishan mafic dikes (105.7–118.5 Ma, K-Ar ages, Xie *et al.*, 2006); 9 - Danyan mafic dikes (110.4 Ma, K-Ar age, Xie *et al.*, 2006); 10 - Maopai mafic dikes (108.5 Ma, K-Ar age, Xie *et al.*, 2006); 11 - Mingxi basalt (2.2 Ma, ^{40}Ar - ^{39}Ar plateau age, Ho *et al.*, 2003); 12 - Yongtai basalt (103.0 Ma, zircon LA-ICPMS age, Xu and Xie, 2005) 13 - Shandai mafic dikes (100.2 Ma, K-Ar age, Xie *et al.*, 2006); 14 - Caotaobei mafic dikes (92.1 Ma, K-Ar age, Xie *et al.*, 2006); 15 - Yanbei mafic dikes (78.6–91.7 Ma, K-Ar ages, Xie *et al.*, 2006); 16 - Huichang basalt (84.6 Ma, Rb-Sr isochron age, Zhou *et al.*, 1998); 17 - Hengshan basalt (124.5–127.6 Ma, K-Ar ages, Wang *et al.*, 2003); 18 - Daijishan mafic dikes (139.9–147.5 Ma, K-Ar ages, Li and McCulloch, 1998; Xie *et al.*, 2006); 19 - North Guangdong mafic dikes (141.0–81.0 Ma, K-Ar ages, Li and McCulloch, 1998); 20 - Sanshui and Heyuan basalt (64–43 Ma, K-Ar ages, Chung *et al.*, 1997; Zhu and Wang, 1989; Zhu *et al.*, 2004); 21 - Niutoushan basalt (14.9–17.1 Ma, ^{40}Ar - ^{39}Ar plateau ages, Ho *et al.*, 2003); 22 - Chinmen and Liehyu Islands mafic dikes (97.8 Ma, ^{40}Ar - ^{39}Ar plateau age, Lan *et al.*, 1995). (b) Simplified geologic map of Paleocene mafic rocks (modified from Zheng and Zhang, 1996) in Central Jiangxi, Southeast China.

and possibly as early as Late Archean; Wang and O'Reilly, 2003) may have survived lithospheric thinning during the Late Mesozoic (Wang and O'Reilly, 2003). This is also suggested by seismic and geothermal data (Hu *et al.*, 2000; Yang, 2003; Yu *et al.*, 2003a), which show that old lithospheric remnants are present in the uppermost layers of the lithosphere (Xu *et al.*, 2002). Nevertheless, the timing and mechanisms for this event remain poorly understood, owing to the lack of knowledge about the nature of the mantle source during the Paleocene that links

Late Mesozoic to Late Cenozoic magmatism in Southeast China.

In contrast to Late Cenozoic basalts, only sparse outcrops of Paleocene mafic rocks occur in East China largely because of the thick Quaternary cover (Fan and Hooper, 1991). The available geochemical data on Paleocene mafic rocks were obtained from core samples from oil wells in the Subei and Liaohe basins and the North China Craton (Yang *et al.*, 1998; Cong *et al.*, 2001; Shen *et al.*, 2002). However, during recent geological surveys in Southeast

China, Paleocene mafic rocks have been found in the Taihe basin, Central Jiangxi (BGMJRJX, 1995a, b, c, d, e).

Paleocene mafic rocks together with Cretaceous and Late Cenozoic basaltic rocks provide a good opportunity to study the changing nature of the mantle source with time and investigate the implications for lithospheric replacement in Southeast China.

GEOLOGICAL SETTING

Southeast China comprises the Cathaysia Block in the southeast and the Yangtze Block in the northwest, separated by the Jianshan-Shaoxing fault (Zhang *et al.*, 2005), and the Chenzhou-Linwu fault, representing the Mesozoic lithospheric boundary between the Yangtze Block and Cathaysia Block (Wang *et al.*, 2003). These two tectonic units were amalgamated in the Neoproterozoic (e.g., Li, 1999). The Cathaysia Block has a complex tectono-magmatic history and was mainly developed on a Precambrian (Paleoproterozoic to Mesoproterozoic, and possibly late Archaean) metamorphic basement (Chen and Jahn, 1998). This ancient basement was unconformably overlain by Sinian marine clastic rocks and dolomite, Paleozoic marine carbonate and clastic rocks, and Early Mesozoic sandstone and shale. The Cathaysia Block was affected by Late Mesozoic magmatism, which is widespread in Southeastern China, and then followed by Late Cenozoic basalts erupted along the coastal region. Some of these basalts contain abundant mantle-derived xenoliths (Xu *et al.*, 2000, 2002, 2003; Yu *et al.*, 2003a; Wang and O'Reilly, 2003; Zheng *et al.*, 2004).

Paleozoic to Lower Triassic shallow-marine carbonates, clastic deposits, and flysch sequences were unconformably overlain by Mesozoic volcano-sedimentary rocks that accumulated in rift basins. These basins were formed since the Middle Jurassic, and contain volcanic rocks that are dominantly characterized by basalts and rhyolites with minor andesite-dacites in Central-Southern Jiangxi (Wang *et al.*, 2004, 2005). In addition, minor Jurassic gabbroic intrusions are also present in Southern Jiangxi, as exemplified by Middle Jurassic Chebu intrusion (Li *et al.*, 2003; Xie *et al.*, 2005b). Also present in Central-Southern Jiangxi are Late Mesozoic granitic rocks and minor Cretaceous basalts and associated mafic dikes (Li and McCulloch, 1998; Zhou and Li, 2000; Wang *et al.*, 2003; Peng *et al.*, 2004; Yu *et al.*, 2005; Xie *et al.*, 2006). Several Cretaceous to Tertiary rift basins in Central Jiangxi, contain red mudstone and sandstone with minor mafic intrusions (BGMJRJX, 1984; Zhou and Li, 2000). One of these is the Late Cretaceous Taihe basin, with an area of 4000 km² (Gilder *et al.*, 1991; Zheng and Zhang, 1996; Zhou and Li, 2000; Yu *et al.*, 2005) in which samples for this study were collected from five localities, namely: Luozishan, Hebu, Tianxinggang, Luokou

Fig. 2. Sketch map showing spatial relationships between mafic rocks and Late Cretaceous sedimentary rocks in Central Jiangxi. Southeast China (a) and (b) modified after 1:50 000 Futian and Taihe geological maps, respectively (BGMJRJX, 1995a, b).

and Xigangshan (Fig. 1b). The mafic intrusive rocks in the Taihe basin commonly have elliptical to lensoid shapes (Fig. 1b) and are emplaced in the Late Cretaceous Yulou Formation (siltstone and feldspathic sandstone) and Honggang Formation (mainly conglomerate) in the lower part of Guangxi Group (95–80 Ma) (Liu and Li, 2001), and Zhoutai Formation (siltstone and mudstone) of Gangzhou Group (100–95 Ma) (Liu and Li, 2001) (Fig. 2). The contact between the mafic intrusions and the host rocks is sharp, and chilled margins are present with widths ranging from several cm to tens of centimeters (BGMJRJX, 1995a, b, c, d, e). All samples in this study are generally fresh. The studied mafic rocks are porphyritic diabase, predominantly composed of plagioclase (40–50%), clinopyroxene (25–35%) and olivine (5–15%) with minor Ti-magnetite (5%). Olivine is an almost ubiquitous phenocryst phase, locally occurring together with clinopyroxene and plagioclase. The groundmass ranges from glassy to cryptocrystalline, and identifiable minerals in it are olivine, plagioclase, clinopyroxene and magnetite. These mafic rocks are generally unaltered, but in some altered samples olivine grains are replaced by iddingsite along minerals cracks; whereas clinopyroxene and plagioclase are replaced by minor chlorite.

ANALYTICAL METHODS

Major element analyses were carried out by a conventional wet chemical method at the Chemical Analytical Center, Institute of Geochemistry, Chinese Academy of Sciences (IGCAS). The analytical errors for major oxides are generally less than 2%. Trace elements and rare earth elements abundances were determined by solution ICP-MS performed at the ICP-MS Laboratory, Institute of Geochemistry, Chinese Academy of Sciences (IGCAS). The samples were digested by acid (HF + HClO₄) in bombs. Precision for most elements was typi-

Table 1. K-Ar dating results for mafic rocks in the Taihe Basin, Central Jiangxi, SE China

Sample No.	Weight/g	K (%)	$^{40}\text{Ar}^*$ (10^{-10} mol/g)	$^{40}\text{Ar}^*$ (%)	$^{40}\text{Ar}^*/^{40}\text{K}$ (10^{-3})	T (Ma)
LZS1	0.03207	1.08	0.9335	71.44	2.8959	49.2 ± 0.9
JCX7	0.03053	0.90	0.7941	59.20	2.9564	50.2 ± 1.2
XG1	0.02995	0.92	0.8060	55.77	2.9353	49.8 ± 1.6
XG2	0.0316	0.97	1.0699	65.10	3.6955	62.5 ± 1.9
LZX1	0.03164	1.45	1.3458	62.48	3.1098	52.7 ± 1.2
LZX3	0.02996	1.24	1.3999	64.73	3.7826	64.0 ± 1.0
TXG2	0.0279	1.81	1.9506	80.95	3.6108	61.1 ± 1.4
TXG3	0.03009	1.79	2.0637	82.42	3.8627	65.3 ± 1.1

Note: Sample Numbers with prefix letters for LZS, JCX, XG, LZX and TXG from Luozishan, Hebu, Xigangshan, Luokou and Tianxinggang areas, respectively. Sample Numbers and locations are same as Tables 2 and 3.

cally better than 5% RSD. The measured values for Zr, Hf, Nb and Ta were less than 10% in error compared to certified values. Sample preparations, instrument operating conditions and calibration procedures follow those established by Qi and Grégoire (2000).

For Sr-Nd isotopic analyses, sample powders (~100 mg) were dissolved in distilled HF-HNO₃ in Savillex Screwtop Teflon beakers at 150 C overnight. Sr and REE were separated on columns made of Sr and REE resins of the Eichrom Company using 0.1% HNO₃ as elutant. Separation of Nd from the REE fractions was carried out on HDEHP columns with a 0.18N HCl elutant. The isotopic analyses were performed using a Micromass Isoprobe multi-collector-ICPMS (MC-ICPMS) at Guangzhou Institute of Geochemistry, Chinese Academy of Sciences (GIGCAS). Measured Sr and Nd isotopic ratios were normalized using an $^{86}\text{Sr}/^{88}\text{Sr}$ value of 0.1194 and a $^{146}\text{Nd}/^{144}\text{Nd}$ value of 0.7219, respectively, $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios were calculated using the Rb, Sr, Sm and Nd abundances measured by ICP-MS. Analyses of standards during the period of analysis, are as follows: NBS987 gave $^{87}\text{Sr}/^{86}\text{Sr} = 0.710243 \pm 14$ (2 σ); Shin Etou JNDi-1 gave $^{143}\text{Nd}/^{144}\text{Nd} = 0.512124 \pm 11$ (2 σ), equivalent to a value of 0.511860 for the La Jolla international standard (Tanaka *et al.*, 2000). For Pb isotope analyses, sample powders were spiked and dissolved in concentrated HF at 800°C for 72h. Lead was separated and purified by conventional anion-exchange technique (AG1 $\times$ 8, 200–400 resin) with diluted HBr. Isotopic ratios were measured using the MAT-261 mass spectrometer at the Research Center for Geological Analysis, Institute of Uranium Geology (GAIUG). In-run analytical precision for 1 μg $^{208}\text{Pb}/^{206}\text{Pb}$ is better than 0.005%. Repeated analyses of NBS981 yielded $^{208}\text{Pb}/^{206}\text{Pb} = 2.1652465 \pm 0.000069$, $^{207}\text{Pb}/^{206}\text{Pb} = 0.9145100 \pm 0.000056$ and $^{204}\text{Pb}/^{206}\text{Pb} = 0.0591995 \pm 0.000013$. K-Ar dating was performed using the high sensitivity mass spectrometer (MM1200B) at the K-Ar and ^{40}Ar - ^{39}Ar Isotopic Laboratory of Institute of Geology, China Academy of Geological Society

(IGCAGS). The analytical result for Chinese standard ZBH-25 is 132.5 ± 1.6 Ma (recommended age 132.7 Ma). The age calculation parameters used are $K^{40} = 0.1167\%$, $K_e = 5.811 \times 10^{-11}/\text{year}$ and $K_b = 4.962 \times 10^{-10}/\text{year}$. K-Ar dating results, major and trace element composition and Nd-Sr-Pb isotopic data for mafic rocks in Central Jiangxi, SE China are listed in Tables 1 to 3, respectively.

RESULTS

Emplacement ages of Paleocene mafic rocks

Samples of mafic intrusions in the Taihe Basin yield K-Ar ages (Table 1) of 52.7–64.0 Ma (Luokou), 49.2 Ma (Luozishan), 50.2 Ma (Hebu), 61.1–65.3 Ma (Tianxinggang) and 49.8–62.5 Ma (Xigangshan). These results are consistent with previous results of K-Ar dating of samples from Luokou, Tianxinggang and Xigangshan which yielded 55.9 Ma, 62.6–63.1 Ma and 63.52 Ma, respectively (BGMJRJX, 1995a, b, e; Yu *et al.*, 2005) and Sm-Nd model ages of 62 ± 2 Ma (Liu and Li, 2001). The mafic rocks intruded the Late Cretaceous Yulou Formation and Honggang Formation in the lower part of Guangxi Group (95–80 Ma) and Zhoutai Formation of Gangzhou Group (100–95 Ma) (Liu and Li, 2001) in the Hebu, Tianxinggang and Luokou region (BGMJRJX, 1995a, b, d), implying that these mafic rocks are not earlier than 80 Ma in age. Although one sample from Luozishan gives a $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 90.2 ± 0.3 Ma (Peng *et al.*, 2004), both geological evidence and age dating in this study confirm that these mafic rocks in Central Jiangxi were intruded in the Paleocene.

Major and trace element data

The Taihe Basin Paleocene mafic rocks are characterized by a wide range of SiO₂ (45.01–51.67%) and K₂O + Na₂O (3.11–6.20%), low CaO (5.96–9.28%) and K₂O/Na₂O (0.24–0.59) abundances, but high Na₂O (2.36–4.36%) contents.

In the total alkali versus silica classification of Irvine

Fig. 3. Total alkalis ($K_2O + Na_2O$) versus SiO_2 diagram for Paleocene mafic rocks in the Taihe Basin, Central Jiangxi, Southeast China. The division between alkaline and sub-alkaline is after Irvine and Baragar (1971). The fields of Cretaceous mafic rocks and basalts (e.g., Lan *et al.*, 1995; Lapierre *et al.*, 1997; Li and McCulloch, 1998; Xiong *et al.*, 2003; Wang *et al.*, 2003; Xie *et al.*, 2006) and Late Cenozoic basalts (e.g., Flower *et al.*, 1992; Qi *et al.*, 1994; Liu *et al.*, 1994; Zou *et al.*, 2000) in Cathaysia Block are shown for comparison.

and Baragar (1971) these rocks plot in the alkaline to sub-alkaline fields (Fig. 3). Large variations in MgO (5.43–8.87%), Cr (68.0–212 ppm) and Ni (74.7–205 ppm), point to variable fractional crystallization during the magmatic evolution. This is also supported by correlations between SiO_2 , MgO, other major oxides and trace elements (Fig. 4). For example, all samples show negative correlations between SiO_2 and TiO_2 and P_2O_5 (Figs. 4a and c), which suggests fractionation of ilmenite, apatite and clinopyroxene (Wilson, 1989). Compatible elements such as Cr and Ni decrease with MgO in the mafic rocks (Figs. 4d and f), implying fractional crystallization of olivine and clinopyroxene (Wilson, 1989). Positive correlations between CaO/Al_2O_3 and CaO (Fig. 4e) are also indicative of fractional crystallization of clinopyroxene (Qi *et al.*, 1994).

The mafic rocks show high REE patterns (Fig. 5) that are characterized by strong LREE enrichment with minor Eu anomalies ($Eu/Eu^* = 0.88–1.2$; Table 2), variable La_N/Yb_N (4.8–14.6) and LREE/HREE ratios (5.0–10.0). The abundances of incompatible trace elements for these rocks are higher than those for Primitive Mantle (PM) and Mid-ocean ridge basalt (MORB) (Sun and McDonough, 1989). The mafic rocks exhibit strong enrichment of large ion lithophile elements (LILE) (Ba, Sr, K and Rb), LREE, slight enrichment of high strength field elements (HSFE) (Nb, Ta, Zr, Hf) and incompatible trace element patterns. These geochemical features resemble

Fig. 4. Harker-type diagrams of trace elements versus major elements for Paleocene mafic rocks in the Taihe Basin, Central Jiangxi, Southeast China.

those of ocean island basalt (OIB) (Sun and McDonough, 1989) (Fig. 6). The mafic rocks also have relatively high Nb/La ratios (1.1–1.6), distinguishing them from continental arc basalts (McCulloch and Gamble, 1991). It is noted that the MORB normalized trace element patterns for these samples (Fig. 6) are similar to those of Late Cenozoic basalts in Southeast China (Peng *et al.*, 1986; Flower *et al.*, 1992; Liu *et al.*, 1994; Qi *et al.*, 1994; Chung *et al.*, 1994, 1995; Zou *et al.*, 2000; Ho *et al.*, 2003). The MORB-normalized multielement diagrams display small differences, for example the Luokou mafic rocks exhibit weak Rb positive anomalies and minor Ba negative anomalies, while the Tianxinggang rocks show weak Ba positive anomalies and minor Rb negative anomalies.

Table 2. Major element (%) and trace element and rare earth element concentrations (ppm) of Paleocene mafic rocks in the Taihe Basin, Central Jiangxi, SE China

Sample	LZX1	LZX6	LZX3	963-1 ^a	643-1 ^a	212 ^a	XG1	XG2	XG5	XGM ^b	TXG2	TXG3	TXG6	469-1 ^c	JCX3	JCX6	JCX7	LZS1	LZS3	LZS7	JCXN ^d	LZSM ^e
SiO ₂	47.44	47.05	51.61	49.08	48.66	50.80	49.83	49.93	49.36	50.14	48.78	48.02	47.45	46.54	46.90	46.46	45.70	46.20	46.35	45.01	46.08	46.86
TiO ₂	0.31	1.12	0.87	1.50	1.50	1.35	1.50	1.75	1.85	2.15	1.62	1.90	2.77	2.75	2.83	2.81	2.82	1.55	2.40	2.90	2.35	2.20
Al ₂ O ₃	18.99	17.22	16.97	16.47	16.54	16.41	15.99	16.72	16.56	15.70	16.24	18.20	16.80	15.22	15.58	14.85	15.83	18.45	16.07	18.02	14.30	13.72
Fe ₂ O ₃	2.12	4.62	3.81	3.02	5.26	5.65	4.36	4.26	1.89	4.16	4.73	3.99	1.62	3.03	1.71	4.53	4.28	5.95	0.53	4.49	3.66	3.12
FeO	5.92	4.07	3.90	5.66	3.91	3.47	7.94	7.20	7.10	7.83	7.00	6.30	6.20	7.93	6.25	5.34	5.95	4.82	7.52	5.40	7.86	8.18
MnO	0.17	0.19	0.19	0.15	0.15	0.15	0.22	0.27	0.15	0.18	0.35	0.39	0.08	0.18	0.12	0.18	0.21	0.34	0.14	0.23	0.14	0.16
MgO	6.92	7.46	7.13	8.01	7.54	7.44	5.52	5.43	5.73	6.34	6.03	5.99	6.55	7.17	8.01	7.61	6.88	6.71	7.75	6.53	8.87	9.26
CaO	7.29	6.46	5.96	6.67	7.17	7.01	8.64	8.73	9.28	8.76	7.11	6.08	7.87	7.60	8.80	8.66	8.85	7.62	9.17	8.34	8.63	8.35
Na ₂ O	4.30	2.75	2.91	4.36	3.31	3.33	2.36	2.43	3.22	2.68	2.70	2.79	4.27	3.24	3.86	3.56	3.74	3.25	3.98	3.78	3.29	3.16
K ₂ O	1.41	1.13	1.05	1.67	1.92	1.86	0.75	0.77	1.00	0.93	1.58	1.57	1.93	2.05	0.94	1.06	1.02	0.99	1.20	0.98	1.10	1.30
P ₂ O ₅	0.65	0.77	0.97	0.15	0.62	0.52	0.49	0.53	0.45	0.40	0.73	0.80	0.62	0.68	0.55	0.70	0.70	0.83	0.57	1.00	0.57	0.05
LOI	2.97	2.21	2.29	2.65	3.37	2.87	1.70	1.51	1.96	—	2.91	3.72	2.38	2.82	3.63	2.67	2.67	2.93	2.57	2.30	2.98	—
CO ₂	1.05	4.89	1.70	—	—	—	—	—	1.10	—	—	—	0.90	—	0.70	0.95	0.95	—	1.25	0.55	—	—
Sum	99.42	99.94	99.36	99.39	99.95	100.9	99.30	99.53	99.65	99.27	99.78	99.75	99.44	99.21	99.88	99.38	99.60	99.64	99.50	99.53	99.83	96.36
Trace element																						
Sc	24.3	21.3	23.0	—	—	—	28.4	28.7	25.3	—	21.1	18.7	23.5	—	22.6	23.0	23.2	25.5	27.1	22.6	—	—
Cr	219	240	238	—	—	—	98.3	109	68	—	181	166	156	—	295	283	321	214	217	161	—	—
V	164	194	188	—	—	—	241	235	213	—	184	186	161	—	190	174	212	197	200	167	—	—
Co	47.2	53.1	51.7	—	—	—	86.6	86.6	77.9	—	67.4	67.9	59.6	—	56.6	54.1	68.1	70.4	65.1	45.9	—	—
Ni	137	130	127	—	—	—	74.4	82	82.7	—	118	113	119	—	205	200	173	127	154	128	—	—
Cs	9.39	5.05	5.89	—	—	—	0.531	0.481	0.603	—	5.17	7.21	33.2	—	2.87	5.59	1.95	0.448	0.652	0.82	—	—
Pb	6.94	7.52	7.26	—	—	—	4.91	4.97	6.06	—	4.53	4.43	3.92	—	2.61	2.05	2.88	3.27	2.58	2.5	—	—
U	1.69	1.47	1.53	—	—	—	0.758	0.78	1.43	—	1.22	1.18	2.96	—	1.27	0.954	1.09	1.61	1.51	1.48	—	—
Sr	767	677	650	—	—	—	428	427	409	—	708	706	674	—	967	615	686	982	834	824	—	—

Sample	LZX1	LZX6	LZX3	963-1 ^a	643-1 ^a	212 ^a	XG1	XG2	XG5	XGM ^b	TXG2	TXG3	TXG6	469-1 ^c	JCX3	JCX6	JCX7	LZS1	LZS3	LZS7	JCXM ^d	LZSM ^e
Rb	28.36	22.14	16.71	—	—	—	21.95	22.64	20.7	—	45.28	44.37	47.21	—	10.1	20.0	13	51.1	47.4	31.5	—	—
Ba	485	488	491	—	—	—	323	306	273	—	403	286	463	—	525	598	474	608	507	727	—	—
Th	6.37	5.84	5.73	—	—	—	3.19	3.24	3.04	—	5.06	5.12	5.68	—	4.76	3.95	4.25	5.7	5.4	5.96	—	—
Ta	1.96	1.89	1.84	—	—	—	1.5	1.42	1.45	—	3.32	3.28	3.29	—	3.04	2.60	2.82	3.77	3.29	3.88	—	—
Nb	39.3	37.8	38.5	—	—	—	26	25.2	25.8	—	58.8	57.5	58.5	—	59	46.1	56.7	74.7	60.4	68.0	—	—
Zr	172	167	167	—	—	—	200	204	206	—	290	286	283	—	204	192	201	243	210	263	—	—
Hf	4.4	4.14	4.02	—	—	—	6.2	5.79	6.03	—	7.2	7.15	7.16	—	5.29	4.74	5.28	6.35	5.52	5.77	—	—
Y	31.0	29.5	29.8	—	—	—	36.7	35.7	32.1	—	34.6	33.6	34.9	—	31.2	22.7	30.4	32.2	30.8	25.4	—	—
Rare earth element																						
La	36.7	35.3	36.0	—	36.08	—	23.7	23.0	24.1	—	45.2	42.7	42.5	37.83	37.5	31.11	37.6	49.8	38.8	46.59	—	37.83
Ce	72.3	70.7	69.6	—	64.97	—	52.5	52.0	55.4	—	89.0	87.0	88.4	85.0	74.8	60.67	74.3	94.6	76.4	89.38	—	85
Pr	8.05	7.65	8.07	—	8.06	—	6.28	6.11	6.58	—	10.54	10.39	10.3	10.26	8.39	7.243	8.43	10.7	8.7	10.76	—	10.26
Nd	31.5	29.9	31.6	—	23.84	—	30.1	27.5	28.4	—	44.5	41.9	41	41.64	34.8	31.71	36.2	43.6	34.9	43.8	—	41.64
Sm	5.93	6.13	6.33	—	6.56	—	6.32	7.12	6.77	—	8.80	8.47	8.10	7.98	6.99	7.07	7.65	8.43	7.04	8.283	—	7.98
Eu	1.96	1.72	1.84	—	1.94	—	2.03	1.98	2.17	—	2.47	2.41	2.73	2.67	2.57	2.384	2.33	2.41	2.64	2.945	—	2.67
Gd	5.79	5.66	5.81	—	5.32	—	6.94	6.64	6.90	—	8.11	7.55	7.54	7.95	7	6.295	6.95	7.58	7.02	7.168	—	7.95
Tb	0.89	0.94	0.88	—	0.96	—	1.13	1.12	1.11	—	1.20	1.23	1.18	0.95	1.09	0.997	1.1	1.17	1.13	1.034	—	0.95
Dy	5.35	5.45	5.29	—	5.17	—	6.96	6.87	6.12	—	7.00	6.86	6.02	6.62	6.04	4.841	5.83	6.7	5.91	5.589	—	6.62
Ho	1.01	1.01	1.09	—	0.96	—	1.36	1.30	1.17	—	1.28	1.27	1.23	0.97	1.05	0.916	1.08	1.12	1.04	0.973	—	0.97
Er	2.92	2.83	2.92	—	2.19	—	3.58	3.59	3.37	—	3.43	3.32	3.40	2.80	2.64	2.356	2.77	2.83	2.89	2.558	—	2.8
Tm	0.435	0.383	0.379	—	0.32	—	0.491	0.497	0.426	—	0.422	0.442	0.414	0.4	0.317	0.298	0.345	0.38	0.404	0.356	—	0.4
Yb	2.68	2.82	2.83	—	1.98	—	3.31	3.12	2.77	—	3.00	3.02	3.01	2.41	2.21	1.817	2.35	2.71	2.5	2.156	—	2.41
Lu	0.386	0.391	0.393	—	0.32	—	0.444	0.416	0.389	—	0.423	0.383	0.394	0.400	0.303	0.263	0.312	0.382	0.323	0.32	—	0.4
REE	145	141	146	—	176	—	171	173	159	—	225	217	216	208	186	158	187	232	190	222	—	208
δCe	1.0	1.1	1.1	—	1.0	—	1.0	1.0	0.90	—	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	0.96	—	1.0
δEu	0.94	0.88	1.0	—	1.0	—	0.89	0.93	1.0	—	0.90	0.92	1.1	1.0	1.1	1.1	1.0	0.92	1.1	1.2	—	1.0

Note: — not analysis.

^aBGMRIX (1995a), ^bBGMRIX (1995e), ^cBGMRIX (1995b), ^dBGMRIX (1995d), ^eBGMRIX (1995c).

Fig. 5. Chondrite-normalized REE patterns for Paleocene mafic rocks in the Taihe Basin, Central Jiangxi, Southeast China. Normalization values after Boynton (1984). Field of Late Cenozoic basalts (e.g., Flower et al., 1992; Qi et al., 1994; Liu et al., 1994; Zou et al., 2000) in Cathaysia Block are shown for comparison.

Fig. 6. MORB-normalized incompatible element plots for Paleocene mafic rocks in the Taihe Basin, Central Jiangxi, Southeast China. MORB, Primitive mantle (PM) and OIB values after Sun and McDonough (1989), upper crust and granulite in the Southeast China after Gao et al. (1998) and Yu et al. (2003b), respectively. The fields of Cretaceous basalts and mafic dikes (e.g., Lan et al., 1995; Lapiere et al., 1997; Xiong et al., 2003; Ge et al., 2003; Wang et al., 2003; Xie et al., 2006) and Late Cenozoic basalts (e.g., Flower et al., 1992; Qi et al., 1994; Liu et al., 1994; Zou et al., 2000) in Cathaysia Block are shown for comparison.

Sr-Nd-Pb isotope data

The Taihe Basin mafic rocks have relatively high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from 0.7041 to 0.7064, whereas Nd isotopes are heterogeneous, with $\epsilon_{\text{Nd}}(\text{T})$ ranging from +0.8 to +6.2. The $\epsilon_{\text{Nd}}(\text{T})$ values, as shown in Table 3, are +1.7–+1.9 (samples from Luokou), +0.8 (Luozishan), +6.2 (Hebu), +4.7–+4.9 (Tianxinggang) and +1.6–2.0 (Xigangshan). The Sr-Nd isotopic correlation diagrams (Fig. 7) show that they plot within the OIB field, similar

to those of Samoan and Society Island Basalts, which contain relatively high EM 2 mantle component (Zindler and Hart, 1986). The $^{208}\text{Pb}/^{204}\text{Pb}$ ratios for the Paleocene mafic rocks, as shown in Table 3, are 36.919 to 37.150 (Luokou), 36.824 (Luozishan), 36.787 (Hebu), 38.692 to 38.807 (Tianxinggang) and 38.714 to 38.898 (Xigangshan), whereas they have similar $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ ratios of 18.338–18.677 and 15.443–15.636, respectively (Table 3).

Table 3. Pb, Sr and Nd isotopic compositions of Paleocene mafic rocks in the Taihe Basin, Central Jiangxi, SE China

Sample No.	XG1	XG2	TXG2	TXG3	LZX1	LZX3	LZS1	JCX7
U (ppm)	0.758	0.78	1.22	1.18	1.69	1.53	1.61	1.09
Pb (ppm)	4.91	4.97	4.53	4.43	6.94	7.26	3.27	2.88
Th (ppm)	3.19	3.24	5.06	5.12	6.37	5.73	5.70	4.25
$^{208}\text{Pb}/^{204}\text{Pb}(2\sigma)$	39.021 ± 14	38.838 ± 10	39.02 ± 25	38.912 ± 39	37.321 ± 22	37.065 ± 19	37.147 ± 90	37.059 ± 25
$^{207}\text{Pb}/^{204}\text{Pb}(2\sigma)$	15.617 ± 6	15.599 ± 4	15.610 ± 10	15.538 ± 16	15.643 ± 9	15.543 ± 8	15.457 ± 37	15.474 ± 10
$^{206}\text{Pb}/^{204}\text{Pb}(2\sigma)$	18.767 ± 7	18.645 ± 5	18.762 ± 12	18.685 ± 19	18.643 ± 10	18.510 ± 90	18.684 ± 45	18.553 ± 12
$(^{208}\text{Pb}/^{204}\text{Pb})_i$	38.898	38.714	38.807	38.692	37.15	36.919	36.824	36.787
$(^{207}\text{Pb}/^{204}\text{Pb})_i$	15.613	15.594	15.603	15.531	15.636	15.538	15.443	15.464
$(^{206}\text{Pb}/^{204}\text{Pb})_i$	18.677	18.553	18.605	18.53	18.504	18.391	18.404	18.338
Rb (ppm)	21.9	22.6	44.4	45.3	28.4	16.7	51.1	13.0
Sr (ppm)	428	427	706	708	767	650	982	686
$^{87}\text{Rb}/^{86}\text{Sr}$	0.148	0.1531	0.1851	0.1819	0.1071	0.0743	0.1505	0.0548
$^{87}\text{Sr}/^{86}\text{Sr}(2\sigma)$	0.706273 ± 11	0.706275 ± 14	0.70447 ± 11	0.704236 ± 13	0.706509 ± 13	0.705469 ± 11	0.705918 ± 11	0.704237 ± 14
$(^{87}\text{Sr}/^{86}\text{Sr})_i$	0.7062	0.7061	0.7043	0.7041	0.7064	0.7054	0.7058	0.7042
Sm (ppm)	6.32	7.12	8.47	8.8	5.93	6.33	8.43	7.65
Nd (ppm)	30.1	27.5	41.9	44.5	31.5	31.6	43.6	36.2
$^{147}\text{Sm}/^{144}\text{Nd}$	0.1269	0.1565	0.1196	0.1222	0.1138	0.1211	0.1169	0.1278
$^{143}\text{Nd}/^{144}\text{Nd}(2\sigma)$	0.512718 ± 8	0.512702 ± 8	0.512846 ± 8	0.512862 ± 9	0.51270 ± 9	0.512695 ± 8	0.512653 ± 10	0.512933 ± 9
$(^{143}\text{Nd}/^{144}\text{Nd})_i$	0.5127	0.5126	0.5128	0.5128	0.5127	0.5126	0.5126	0.5129
$\epsilon_{\text{Nd}}(T)$	2.0	1.6	4.7	4.9	1.9	1.7	0.8	6.2

Note: Initial $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ and Pb isotope are calculated using K-Ar age. $\lambda_{\text{U}238} = 1.55125 \times 10^{-10}/\text{year}$, $\lambda_{\text{U}235} = 9.8485 \times 10^{-10}/\text{year}$, $\lambda_{\text{Th}232} = 4.9475 \times 10^{-11}/\text{year}$, $(^{87}\text{Sr}/^{86}\text{Sr})_{\text{UR}}^0 = 0.7045$, $(^{87}\text{Rb}/^{86}\text{Sr})_{\text{UR}}^0 = 0.0827$, $(^{143}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}}^0 = 0.51238$, $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}^0 = 0.1967$; $\lambda_{\text{Rb}} = 1.42 \times 10^{-11}/\text{year}$, $\lambda_{\text{Sm}} = 6.54 \times 10^{-12}/\text{year}$ (Rollinson, 1993).

DISCUSSION

Fig. 7. Nd and Sr isotopic variation diagram for Paleocene mafic rocks in the Taihe Basin, Central Jiangxi, Southeast China. DMM, HIMU, EM I and EM II are from Zindler and Hart (1986), MORB in the East Pacific Rise (EPR) is from Wilson (1989), OIB and East Taiwan ophiolite (ETO) N-type basalts field are from Ho et al. (2003). ● Paleocene mafic rock in Central Jiangxi (this study), ■ Paleocene basalts from Sanshui and Heyuan basin, Guangdong (Chung et al., 1997; Zhu and Wang, 1989; Zhu et al., 2004), The fields of Cretaceous basalts and mafic dikes (e.g., Lan et al., 1995; Lapierre et al., 1997; Li and McCulloch, 1998; Chen et al., 2000; Xiong et al., 2003; Ge et al., 2003; Wang et al., 2003; Xie et al., 2006) and Late Cenozoic basalts (e.g., Peng et al., 1986; Tu et al., 1991, 1992; Chung et al., 1995; Zou et al., 2000; Ho et al., 2003) in Cathaysia Block are shown for comparison.

Petrogenesis: crustal contamination or source enrichment?

It is important to assess whether continental basaltic magma is affected by en route crustal contamination. The following evidence suggests that the Taihe Basin mafic rocks were little or not affected by crustal contamination (Fig. 6). (1) Large ion lithophile elements (e.g., Ba, Th, K, Rb) in these mafic rocks have lower abundances than those in the upper crust, and high strength field element (HSFE) and heavy rare earth element (HREE) compositions are higher than those in upper crust (Gao et al., 1998) and granulite xenoliths (Yu et al., 2003b) in Southeast China. In addition, these rocks exhibit geochemical characteristics without Nb and Ta negative anomalies, similar to those of Late Cenozoic basalts in Southeast China (Ho et al., 2003) and OIB (Sun and McDonough, 1989). (2) Nb/U and Ce/Pb ratios are 18.0–52.0 and 9.1–35.8, respectively, higher than those for continental crust (Nb/U = 12.1 and Ce/Pb = 4.1) (Taylor and McLennan, 1995) and the crust in Eastern China (Nb/U = 9.6 and Ce/Pb = 4.0) (Gao et al., 1998). In addition, as shown in Fig. 8, La/Nb and Ba/Nb ratios are 0.64–0.94 and 5.0–13.0, respectively, similar to that of Dupal OIB (Jahn et al., 1999). (3) Pb is significantly enriched in continental crust relative to mantle-derived melts, such as MORB and OIB. Consequently, mantle-derived basalts contaminated from crustal material will significantly enhance the Pb isotopic

Fig. 8. Diagram of La/Nb versus Ba/Nb ratios for Paleocene mafic rocks in Taihe Basin, Central Jiangxi, Southeast China.

Fig. 9. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ or $^{143}\text{Nd}/^{144}\text{Nd}$ and SiO_2 or Th variation diagram for the Taihe Basin Paleocene mafic rocks, Southeast China.

compositions. The average $^{208}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of 37.849, 15.553 and 18.500, respectively, for the Taihe Basin mafic rocks are similar to those of least contaminated Late Cenozoic basalts ($^{208}\text{Pb}/^{204}\text{Pb}$ 37.61–39.26; $^{207}\text{Pb}/^{204}\text{Pb}$ 15.45–15.65 and $^{206}\text{Pb}/^{204}\text{Pb}$ 17.65–19.03) in Southeast China (Peng *et al.*, 1986; Tu *et al.*, 1991, 1992; Chung *et al.*, 1995; Zou *et al.*, 2000). (4) The SiO_2 and Th contents of continental crust are generally higher than those of mantle-derived melts, and these rocks if significantly influenced by crustal contamination en route, would exhibit positive correlations among Sr–Nd isotopic compositions and SiO_2 or Th (Piccirillo *et al.*, 1989). However, the studied mafic rocks do not ex-

Fig. 10. $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ variation diagram for the Taihe Basin Paleocene mafic rocks, Southeast China. DMM, MORB, EM I, EM II and BSE are from Zindler and Hart (1986). ● Paleocene mafic rocks in Central Jiangxi (this study), ■ Paleocene basalts from Sanshui basin, Guangdong (Zhu and Wang, 1989), The fields of Cretaceous basalts and mafic dikes (e.g., Xie *et al.*, 2001, 2006) and Late Cenozoic basalts (e.g., Peng *et al.*, 1986; Tu *et al.*, 1991, 1992; Chung *et al.*, 1995; Zou *et al.*, 2000) in Cathaysia Block are shown for comparison.

hibit clear correlations between $^{87}\text{Sr}/^{86}\text{Sr}$ or $^{143}\text{Nd}/^{144}\text{Nd}$ and crust-derived elements such as SiO_2 and Th (Fig. 9).

From this evidence, we conclude that the Taihe Basin magmas were not significantly affected by crustal contamination. The geochemical characteristics of Paleocene mafic rocks in Southeast China indicate that they inherited mantle characteristics. Consequently, these mafic rocks, with a marked enrichment in LILE and LREE, probably resulted from source enrichment beneath Southeast China.

Magma source

As discussed above, the Taihe Basin Paleocene mafic rocks exhibit EM 2-like isotopic affinities, having high $^{206}\text{Pb}/^{204}\text{Pb}$ (18.338–18.677), high $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7041–0.7064) and moderate $^{143}\text{Nd}/^{144}\text{Nd}$ ($\epsilon_{\text{Nd}}(\text{T}) = +0.8$ – $+6.2$) (Zindler and Hart, 1986). The transitional Pb isotopic compositions between MORB and EM 2 mantle (Fig. 10) suggest mixing between asthenospheric mantle and EM 2 mantle components. These isotopic characteristics can generally be explained as a mixture of asthenosphere and EM 2 mantle, as best illustrated by the $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ (Fig. 11) and $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams (Fig. 12). It has been proposed the EM 2 mantle component in Southeast China was derived from continental lithospheric mantle such as Tungchihsu Group II pyroxenite (Ho *et al.*, 2000), and related to the subduc-

Fig. 11. Initial $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{144}\text{Nd}/^{143}\text{Nd}$ variation diagram for Paleocene mafic rocks, Southeast China. ● Paleocene mafic rocks (this study), ■ Paleocene basalts from Sanshui basin, Guangdong (Zhu and Wang, 1989). Fields of Cretaceous basalts and mafic dikes (e.g., Xie *et al.*, 2001, 2006) and Late Cenozoic basalts (e.g., Peng *et al.*, 1986; Tu *et al.*, 1991, 1992; Chung *et al.*, 1995; Zou *et al.*, 2000) in Cathaysia Block are shown for comparison.

tion of the Paleo-Pacific plate beneath the Eurasian continent in the Late Mesozoic (Chung *et al.*, 1995). Alternatively, these isotopic characteristics may reflect an input from sediments associated with the subduction slab (Zhu and Wang, 1989; Ho *et al.*, 2003). In addition, continental lithospheric mantle reservoirs in extensional terrains may have diverse origins and distinct compositions in southeastern China (Chung *et al.*, 1995). The large variations of Nd and Pb isotopic compositions for the Taihe Basin mafic rocks also demonstrate a highly heterogeneous sub-lithospheric mantle beneath this region during the Paleocene. On the basis of the above-mentioned geochemical and isotopic features we argue that the original magma for Central Jiangxi Paleocene mafic rocks was derived from the asthenospheric mantle with a minor involvement of EM 2 lithospheric mantle.

Secular magma sources for basaltic rocks

Basaltic magmatism is generally the result of mantle partial melting, and magma sources probably change as lithospheric replacement/thinning progresses (e.g., Perry *et al.*, 1988; Daley and DePaolo, 1992). Therefore, it is important to place some temporal constraints on this process by tracking the changes in composition of basaltic magmas through time and relating these changes to the predominant tectonic regime in Southeast China.

In Southeast China Late Cenozoic basalts have OIB-like geochemical characteristics and were dominantly originated by partial melting of depleted asthenospheric mantle (DMM or MORB), with a minor EM 2 lithospheric

Fig. 12. Initial $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ variation diagram for Paleocene mafic rocks in Central Jiangxi, Southeast China. ● Paleocene mafic rocks (this study), ■ Paleocene basalts from Sanshui basin, Guangdong (Zhu and Wang, 1989). The fields of Cretaceous basalts and mafic dikes (e.g., Xie *et al.*, 2001, 2006) and Late Cenozoic basalts (e.g., Peng *et al.*, 1986; Tu *et al.*, 1991, 1992; Chung *et al.*, 1995; Zou *et al.*, 2000) in Cathaysia Block are shown for comparison.

mantle component (Peng *et al.*, 1986; Flower *et al.*, 1992; Tu *et al.*, 1991, 1992; Liu *et al.*, 1994; Qi *et al.*, 1994; Chung *et al.*, 1994, 1995; Zou *et al.*, 2000; Ho *et al.*, 2003). It is also generally accepted that Late Cenozoic volcanism in Southeast China was related to lithospheric extension due to asthenospheric upwelling (Chung *et al.*, 1994, 1995; Qi *et al.*, 1994; Zou *et al.*, 2000; Ho *et al.*, 2003). On the other hand, Cretaceous basaltic magmatism exhibits enriched geochemical and isotopic features, similar to those of continental arc basalts, indicating that they were derived from highly heterogeneous EM 2 lithospheric mantle with a minor involvement of asthenospheric mantle (Lan *et al.*, 1995; Li and McCulloch, 1998; Xie *et al.*, 2001; Xiong *et al.*, 2003; Ge *et al.*, 2003; Wang *et al.*, 2003; Xie *et al.*, 2006), suggesting contamination associated with the subduction of Paleo-Pacific plate (Jahn *et al.*, 1990; Northrup *et al.*, 1995; Lapierre *et al.*, 1997). By contrast, the Paleocene mafic rocks show transitional geochemical features between Cretaceous and Late Cenozoic basaltic rocks in Southeast China. This observation is based on the following evidence:

(1) Most of Cretaceous basaltic rocks belong to the sub-alkaline series (Li and McCulloch, 1998; Zhao *et al.*, 2004; Xie *et al.*, 2006), whereas Late Cenozoic basaltic rocks have compositions ranging from tholeiitic to a strongly alkali affinity (Ho *et al.*, 2003) (Fig. 3).

(2) MORB-normalized incompatible trace elements patterns for Cretaceous basaltic rocks are similar to those of continental arc basalts with Nb and Ta negative anoma-

lies (Lan *et al.*, 1995; Lapierre *et al.*, 1997; Xiong *et al.*, 2003; Ge *et al.*, 2003; Yu *et al.*, 2003b; Zhao *et al.*, 2004; Xie *et al.*, 2006), whereas Late Cenozoic basalts display OIB-like geochemical characteristics with well defined Nb and Ta positive anomalies (Peng *et al.*, 1986; Flower *et al.*, 1992; Tu *et al.*, 1992; Qi *et al.*, 1994; Liu *et al.*, 1994; Chung *et al.*, 1994, 1995; Zou *et al.*, 2000; Ho *et al.*, 2003). Paleocene mafic rocks in this study show transitional geochemical characteristics between Cretaceous and Late Cenozoic basaltic rocks, and are similar to Late Cenozoic basalts (Fig. 6).

(3) Although with similar Pb isotopic compositions (Fig. 10), Cretaceous, Paleocene and Late Cenozoic basaltic magmatism exhibit significantly different Sr-Nd isotopic data. Cretaceous basaltic rocks have high initial Sr ratios ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7051\text{--}0.7145$) and low initial Nd ratios ($^{143}\text{Nd}/^{144}\text{Nd}_i = 0.5119\text{--}0.5128$) (Lan *et al.*, 1995; Lapierre *et al.*, 1997; Li and McCulloch, 1998; Xie *et al.*, 2001, 2006; Xiong *et al.*, 2003; Ge *et al.*, 2003; Wang *et al.*, 2003). Late Cenozoic basalts have low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7032–0.7066) and high initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (0.5126–0.5129) (Peng *et al.*, 1986; Tu *et al.*, 1991, 1992; Chung *et al.*, 1994, 1995; Liu *et al.*, 1994; Zou *et al.*, 2000; Ho *et al.*, 2003). Paleocene mafic rocks in this study plot within fields that are transitional between Cretaceous and Late Cenozoic basaltic rocks on the $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $^{143}\text{Nd}/^{144}\text{Nd}$, $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams (Figs. 7, 11, 12).

From this evidence, we conclude that a secular magma source responsible for both Cretaceous and Cenozoic basaltic rocks in Southeast China changed from the subduction-modified lithospheric mantle in the earlier stages to asthenospheric mantle in the later stages, reflecting a change from a continental arc setting to an intra-plate extensional regime (Chung *et al.*, 1997; Yu *et al.*, 2003b, this study).

Implications for lithospheric replacement

La/Nb, $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ are powerful monitors to distinguish lithospheric mantle from asthenosphere sources for extension-related magmatism (Fitton *et al.*, 1988; DePaolo and Daley, 2000). For example, Fitton *et al.* (1988) postulated that basalts with La/Nb ratios <1.5 from the Basin and Range Province (USA) were originated from melting of an asthenospheric source, whereas lavas with La/Nb ratios of >1.5 from the Sierra Nevada were derived from enriched lithospheric mantle. DePaolo and Daley (2000) proposed that the depth of the lithosphere-asthenosphere boundary can be inferred from the ϵ_{Nd} , $^{87}\text{Sr}/^{86}\text{Sr}$ and La/Nb values of the lavas as a function of age, and an estimate of the depth of the magma source can be obtained on the basis of silica saturation. As shown in Fig. 13, Cretaceous basaltic rocks have higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ and La/Nb (0.04–

Fig. 13. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ or $^{143}\text{Nd}/^{144}\text{Nd}$ versus La/Nb variation diagram for Paleocene mafic rocks in Central Jiangxi, Southeast China. The ratios of Cretaceous and Late Cenozoic basaltic rocks are shown for comparison. ● Paleocene basaltic rocks in Southeast China (Chung *et al.*, 1997, this study), ○ Late Cenozoic basalts in Cathaysia Block (e.g., Chung *et al.*, 1995; Zou *et al.*, 2000; Ho *et al.*, 2003), □ Cretaceous basalts and mafic rocks in Cathaysia Block (e.g., Lan *et al.*, 1995; Lapierre *et al.*, 1997; Li and McCulloch, 1998; Xiong *et al.*, 2003; Ge *et al.*, 2003).

1.0) ratios than Late Cenozoic basalts with La/Nb ranging from (1.1–3.2), whereas the ϵ_{Nd} and La/Nb values for Cretaceous and Cenozoic basaltic rocks increase and the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios decrease with time. This trend suggests that secular magma source responsible for basaltic rocks in Southeast China had been gradually evolved from enriched lithospheric mantle in the Cretaceous to depleted asthenospheric mantle in the Late Cenozoic.

This temporal shift from a lithospheric to asthenospheric magma source beneath Southeast China is similar to that seen in the Basin and Range Province and Northeast Japan where the involvement of subcontinental lithosphere for magma source significantly decreases with time as lithospheric extension and thinning increase (Nohda *et al.*, 1988; Tatsumi *et al.*, 1988; Fitton *et al.*, 1991; DePaolo and Daley, 2000). Therefore, we believe that the involvement of asthenospheric mantle beneath Southeast China gradually increased with time, and the temporal shift from a lithospheric to asthenospheric magma source was probably related to progressive lithospheric extension and thinning, as well as asthenospheric upwelling in this region (Chung *et al.*, 1997; Xu *et al.*, 2000; Wang and O'Reilly, 2003). Many studies indicated that dry asthenosphere could melt only when the lithosphere is sufficiently thinned (<60 km) (Xu *et al.*, 2004). Consequently, the lithospheric mantle was probably significantly thinned and replaced at the beginning of the Paleocene in Southeast China; otherwise Paleocene mafic rocks in this study would not exhibit OIB-like geochemical affinities, distinct from Cretaceous basaltic rocks. Lithospheric thinning and replacement had finished and decompression melting of the upwelling asthenosphere had taken place during the Paleocene pe-

riod in Southeast China. Paleocene mafic rocks in this study represent the surface manifestation of this event (Xu *et al.*, 2004).

The integration of available geological, geochemical and geophysical data of Late Mesozoic igneous rocks indicates that the angle of Pacific plate subducting slab beneath Southeastern China increased from a very low angle to a middle angle from 180 to 80 Ma, and that mantle input may have played an important role in the formation of extensive Mesozoic magmatism (Zhou and Li, 2000). Perhaps, this significant change from a continental arc setting to an intra-plate extensional regime reflected the eastward migration of the subduction zone of the Pacific Plate and/or increase of the subduction angle beneath Southeast China (Yu *et al.*, 2003b). An analysis of the motion of the Pacific plate relative to the Eurasian continent suggests that the rate of the Pacific-Eurasia convergence was 120–140 mm/yr in the Late Cretaceous and declined substantially during the Paleocene period and reached a minimum in the Eocene time of about 30–40 mm/yr, which correlates with widespread extension along the eastern margin of Eurasia in the Cenozoic (Northrup *et al.*, 1995).

As discussed above, secular geochemical and isotopic variation for Cretaceous and Cenozoic basaltic rocks in Southeast China is similar to that seen in Northeast Japan where the Nd and Sr isotopic characteristics of Late Tertiary basalts show a trend from enriched signature in the pre-opening stage of Japan Sea to a depleted one in the post-opening (Nohda *et al.*, 1988; Tatsumi *et al.*, 1988). This temporal shift from a lithospheric to asthenospheric magma source may be associated with back-arc rifting and trench retreat and the steepening of the subducted oceanic lithosphere, probably induced by the asthenosphere injection into the mantle wedge (Nohda *et al.*, 1988; Tatsumi and Kimura, 1991). Consequently, we suggest that lithospheric replacement may have been related to lithosphere extension and thinning, and asthenospheric upwelling during the Paleocene, which resulted from the subduction roll-back and/or steepening of the Pacific plate due to injection of asthenospheric material into the mantle wedge beneath Southeast China.

CONCLUSIONS

On the basis of the age data integrated with Sr-Nd-Pb isotope systematics presented in this contribution, we conclude that:

(1) K-Ar dating and field evidence show that mafic rocks in the Taihe Basin in central Jiangxi, SE China were emplaced in the Paleocene.

(2) Geochemical and isotopic evidence suggests that these mafic rocks are dominantly originated from asthenospheric mantle (DMM or MORB) with a minor

involvement of EM 2 lithospheric mantle.

(3) Paleocene mafic rocks exhibit a transitional geochemical characteristics between Cretaceous and Late Cenozoic basaltic rocks, indicating that the involvement of asthenospheric mantle beneath Southeast China gradually increase with time.

(4) The lithospheric replacement was associated with lithospheric extension and thinning, with decompression melting of the upwelling asthenosphere having taken place in the Paleocene. This, in turn, was probably related to subduction roll-back and/or steepening of Paleo-Pacific plate due to injection of asthenospheric material into the mantle wedge beneath Southeast China.

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