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**Geochemical and Sr–Nd–Pb–Hf–O isotopic compositions of the
Tiezhai complex: implications for lithosphere destruction of the
North China Craton**

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Abstract:

It is generally accepted that in the Late Mesozoic, more than 100 km of ancient refractory lithospheric mantle beneath the eastern part of the North China Craton (NCC) was removed and replaced by young and fertile mantle material. However, the mechanisms and a tectonic model for this lithospheric destruction remain subjects of debate. To elucidate this lithospheric destruction, an investigation that included whole-rock geochemical analyses, zircon U–Pb age determinations, and in situ Hf–O isotope analyses of samples from the Tiezhai syenite porphyry and mafic rocks from western Shandong Province (the Luxi terrane) was conducted. Zircon U–Pb dates reveal that the Tiezhai syenite porphyry, gabbro, and monzonite were emplaced during the Early Cretaceous (weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 130–129 Ma). Abundant Mesoarchean–Paleoproterozoic inherited zircons/cores were identified in both the syenite porphyry and the gabbro, yielding $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2.95–2.34 Ga. Both the syenite porphyry and mafic rocks are metaluminous, high-K calc-alkaline rocks enriched in large ion lithophile elements (e.g., Rb, Sr, and Ba) and light rare earth elements, and depleted in high field strength elements (e.g., Nb, Ta, and Ti) and heavy rare earth elements. The samples have no obviously Eu anomalies. The syenite porphyry has high Sr and La contents, high Sr/Y (196–292) and $(\text{La}/\text{Yb})_{\text{N}}$ (24.8–36.2) ratios, and low Y and Yb contents, similar to adakitic rocks. The Tiezhai syenite porphyry has a relatively narrow range of $(^{87}\text{Sr}/^{86}\text{Sr})_{\text{i}}$ ratios (0.7057–0.7059), $\epsilon_{\text{Nd}}(t)$ values (–8.9 to –6.8), and relatively high radiogenic Pb isotopic compositions [$(^{206}\text{Pb}/^{204}\text{Pb})_{\text{i}} = 18.36\text{--}18.37$, $(^{207}\text{Pb}/^{204}\text{Pb})_{\text{i}} = 15.54\text{--}15.55$, and $(^{208}\text{Pb}/^{204}\text{Pb})_{\text{i}} = 38.25\text{--}38.31$]. The Tiezhai mafic rocks can be divided into two groups (1 and 2) based on their whole-rock Sr, Nd, and Pb isotopic compositions. Group 1 has lower $(^{87}\text{Sr}/^{86}\text{Sr})_{\text{i}}$ ratios than group 2 (0.7046–0.7047 vs. 0.7072–0.7096), lower radiogenic Pb isotopic

compositions [$(^{206}\text{Pb}/^{204}\text{Pb})_i = 17.68\text{--}17.71$ vs. $18.08\text{--}18.12$; $(^{207}\text{Pb}/^{204}\text{Pb})_i = 15.44\text{--}15.45$ vs. $15.52\text{--}15.53$; $(^{208}\text{Pb}/^{204}\text{Pb})_i = 37.72\text{--}37.98$ vs. $40.17\text{--}40.60$], and slightly higher $\varepsilon_{\text{Nd}}(t)$ values (-8.0 to -7.1 vs. -10.8 to -8.2). The syenite porphyry and gabbro from group 2 have similar narrow ranges of zircon rim $\varepsilon_{\text{Hf}}(t)$ values (mostly between -4 and $+2$), and high zircon rim $\delta^{18}\text{O}$ values (syenite porphyry = $6.9\text{‰}\text{--}9.0\text{‰}$; gabbro = $7.7\text{‰}\text{--}9.1\text{‰}$), and low inherited zircon/core $\delta^{18}\text{O}$ values (syenite porphyry = $5.8\text{‰}\text{--}6.9\text{‰}$; gabbro = $5.7\text{‰}\text{--}6.9\text{‰}$). The Tiezhai mafic rocks were probably derived from partial melting of the Luxi EM1-asthenospheric mantle previously metasomatized by fluids derived from subducted oceanic crust and sediments. The melts probably assimilated ancient NCC upper crust material ($\sim 5\%$ assimilated by group 1 mafic rocks, $20\text{--}30\%$ assimilated by group 2 mafic rocks) during magma ascent. The Tiezhai syenite porphyry was most likely derived from partial melting of the newly underplated thick lower crust and assimilated $\sim 10\%$ of the ancient NCC upper crust material during ascent. The addition of fluids derived from oceanic crust and sediments to the overlying subcontinental lithospheric mantle during subduction was an important factor contributing to lithosphere destruction in the eastern NCC. Slab rollback, together with the ridge subduction of the paleo-Pacific slab beneath the eastern NCC, is the most reasonable tectonic model for NCC destruction and formation of the Early Cretaceous large-scale magmatic rocks in the eastern NCC.

Keywords: Geochronology; Sr–Nd–Pb isotopes; Zircon Hf–O isotopes; Tiezhai complex; North China Craton

1. Introduction

The North China Craton (NCC) is an ideal terrane in which to study extensive lithospheric thinning, because the lithospheric thickness was reduced from ~200 km during the Ordovician to 60–80 km in the Cenozoic (Menzies et al., 1993; Griffin et al., 1998; Menzies and Xu, 1998; Fan et al., 2000; Gao et al., 2002). There is no doubt that the lithospheric mantle of the NCC was thinned considerably during the Late Mesozoic (Fan et al., 2000; Xu, 2001; Wilde et al., 2003; Wu et al., 2003; Yang et al., 2003; Zhang et al., 2003, 2017; Li et al., 2017a, 2017b), however, mechanisms and tectonic models for this lithospheric destruction are still debated.

Three main mechanisms have been proposed for NCC lithosphere destruction: thermo–mechanical–chemical erosion (Griffin et al., 1998; Xu, 2001; Xu et al., 2004b, 2009; Zheng et al., 2007), rapid delamination (Gao et al., 2004; Jiang et al., 2010; Ma et al., 2014a; He et al., 2016; Sun et al., 2017), and addition of water to the overlying subcontinental lithospheric mantle (SCLM) during subduction of the paleo-Pacific slab (Niu, 2005; Windley et al., 2010; Zhu et al., 2012). According to these hypotheses, the modification and thinning of the NCC SCLM resulted from recycling of crustal material in the mantle sources, for example melts from subducted Yangtze Block (YB) continental crust, melts from delaminated NCC lower crust, or fluids from subducted paleo-Pacific oceanic crust. A study of mantle xenoliths from the NCC SCLM would be helpful for constraining the nature of the recycled crustal material and aid in evaluating these hypotheses. Unfortunately, mantle xenoliths in the NCC are uncommon so mantle-derived mafic rocks must be studied instead.

Tectonic models concerning lithospheric destruction and the processes that generated Early Cretaceous magmatic activity in the eastern NCC are also still debated. Zhao and Zheng (2009)

and Zhang et al. (2010a) proposed that post-collision extension of the YB and the NCC during the Triassic was the main geodynamic mechanism. However, other studies have suggested that the Early Cretaceous magmatism cannot be fully explained by processes following continental collision between the NCC and the YB in the Triassic because of the time lag of more than 100 Myr between collision and magmatism (Ma et al., 2006, 2014a; Jiang et al., 2010; Lan et al., 2011). Rather, these studies proposed that slab roll-back, together with ridge subduction of the paleo-Pacific slab, was the most likely geodynamic mechanism that caused Early Cretaceous magmatism (Chen and Zhou, 2005; Wu et al., 2005; Sun et al., 2007; Jiang et al., 2010; Ling et al., 2013; Li et al., 2014a, 2017b; Ma et al., 2014a; Wang et al., 2015, 2016; Cai et al., 2018). Whether slab subduction would generate an extensional environment (Lan et al., 2011, 2013) or even provide material (Chen and Zhou, 2005; Ma et al., 2014b; Wang et al., 2015, 2016) for large-scale Early Cretaceous magmatism in the eastern NCC remains uncertain. Systematic geochronologic, geochemical, and isotopic investigations of mantle-derived magmatic rocks from the NCC could provide evidence about if and how the subducted paleo-Pacific slab affected the mantle.

The Tiezhai complex is an Early Cretaceous pluton in western Shandong Province (the Luxi terrane), and until now, has not been studied systematically. Xu et al. (2004a) dated the Tiezhai syenite porphyry at 124.9 ± 0.2 Ma using by the ^{40}Ar – ^{39}Ar method on biotite and hornblende. Yang (2007) proposed that the Tiezhai complex originated from partial melting of enriched lithospheric mantle based on whole-rock geochemistry and the Sr–Nd isotopic compositions of two samples. Zhong and Huang (2012) used laser ablation-inductively coupled plasma-mass spectrometry (LA–ICP–MS) U–Pb radiometric dating on zircon to date the Tiezhai gabbrodiorite at 132 ± 1 Ma. It should be noted, however, that the petrogenesis and geodynamic setting of this

pluton are poorly constrained. In this study, we present new zircon U–Pb geochronology, in situ Hf–O isotopic data, whole-rock geochemical data, and Sr–Nd–Pb isotopic compositions for the Tiezhai complex. The aims of the study are to: (1) identify the source and clarify the petrogenesis of the magmas that formed the complex, and (2) constrain the mechanisms and tectonic model for the destruction of the NCC lithosphere.

2. Geological background

The NCC has a 3.8 Ga Archean core (Liu et al., 1992) and is the largest craton in China, bounded to the north by the Central Asian Orogenic Belt (Sengör et al., 1993; Davis et al., 2001) and to the south by the Qinling–Tongbai–Dabie–Sulu Orogenic Belt (Li et al., 1993; Meng and Zhang, 2000). Based on lithological, geochemical, and metamorphic pressure/temperature/time (P/T/t) pathways of basement rocks, the craton can be divided into three sub-blocks (Fig. 1a; Zhao et al., 2005): the Eastern North China Craton (ENCC), the Western North China Craton (WNCC), and the intervening Trans-North China Orogen (TNCO) that formed during the collision between the ENCC and WNCC at either ~1.8 Ga (Zhao et al., 2001, 2005) or ~2.5 Ga (Zhai et al., 2003). Recently proposed models suggest that the core of the NCC is composed of seven ‘microblocks’, namely the Jiaoliao, Qianhuai, Ordos, Ji’ning, Alashan, Xuhuai and Xuchang blocks (Li et al., 2016; Santosh et al., 2016; Yang et al., 2016a, 2016b; Li and Santosh, 2017b; Yang and Santosh, 2017). The NCC basement consists predominantly of Archean–Paleoproterozoic Tonalite–Trondhjemite–Granodiorite (TTG) gneisses and greenschist–granulite facies metamorphic rocks, overlain by Sinian–Ordovician marine carbonates and shales, Carboniferous–Permian continental clastic rocks, and Mesozoic continental-basin sedimentary rocks (Zhao et al., 2001, 2005). Volcanism has occurred in the ENCC since the Paleozoic,

whereas magmatism is rare in the WNCC (Xu et al., 2009).

The Luxi terrane is located in the central part of the ENCC, bounded by the Tan–Lu Fault to the east, the Liaocheng–Lankao Fault to the west, the Fengpei Fault to the south, and the Qihe–Guangrao Fault to the north (Zhang et al., 2007; Fig. 1b). Strata of Archean–Proterozoic basement rocks, and Paleozoic marine sedimentary rocks and Mesozoic–Cenozoic sedimentary–volcanic rocks, which are typical of NCC stratigraphy (BGMRS, 1991), are well developed and preserved in the region. Mesozoic intrusive rocks in the Luxi terrane include gabbro, diorite, monzonite, granite, and alkaline complexes (Zhang et al., 2005; Liu et al., 2008a; Lan et al., 2011, 2012, 2013; Huang et al., 2012; Guo et al., 2013a; Wang et al., 2015, 2016; Li et al., 2017; Fig. 1b). In contrast, Mesozoic volcanic rocks are dominantly andesitic (Qiu et al., 1997), and are located in faulted basins such as the Zouping, Fangcheng, and Feixian basins (Qiu et al., 2002; Zhang et al., 2002; Guo et al., 2003; Ling et al., 2009; Fig. 1b).

The 25 km² Tiezhai complex, located in the northeastern part of the Luxi terrane, ~20 km southwest of Linqu County, intruded the Cambrian basement (Fig. 1c). Strata in the Tiezhai area are mainly Cambrian carbonates and Archean basement, with minor Ordovician marine sedimentary rocks. The area is cut by NE–SW and NW–SE striking faults. The intrusion comprises mainly gabbrodiorite and syenite porphyry, with minor gabbro and monzonite porphyry.

3. Sample descriptions and analytical methods

Eight syenite porphyry samples (TZ-2, TZ-3, TZ-4, TZ-5, TZ-7, TZ-8, TZ-10 and TZ-11,) and five mafic rock samples (TZ-13, TZ-14, TZ-15, TZ-16 and TZ-17) were collected from the Tiezhai complex. The syenite porphyry samples have a porphyritic texture and are not foliated.

Most phenocrysts are feldspars but minor hornblende and quartz phenocrysts are also present. Feldspar phenocrysts, ~45 vol.% are 0.4–5.0 mm in size and are euhedral–subhedral, mainly albite. Minor K-feldspar phenocrysts (0.4–0.8 mm, ~5 vol.%) are also present. The matrix feldspar (~40 vol.%) is subhedral–anhedral, mainly K-feldspar and albite. Euhedral to subhedral hornblende, ~5 vol.%, occurs mainly as phenocrysts 0.3–1.0 mm in size. Quartz, ~3 vol.%, occurs mainly in the matrix with a few phenocrysts; all the quartz is anhedral. A few anhedral magnetite and euhedral titanite grains are also present (Fig. 2a, b).

The Tiezhai monzonite is fine-grained and massive. The mineral assemblage is dominated by plagioclase (euhedral–subhedral, 0.2–1.2 mm, 45%) and K-feldspar (anhedral, 0.4–1.5 mm, ~35%), with minor amount of hornblende (subhedral–anhedral, 0.1–0.8 mm, ~10%), biotite (subhedral–anhedral, 0.2–1.0 mm, ~7%), and quartz (anhedral, 0.1–0.4 mm, ~3%). The accessory minerals present are magnetite, zircon, and apatite (Fig. 2c, d).

The Tiezhai monzodiorite is also have a fine-grained and massive. They contain plagioclase (euhedral–subhedral, 0.2–0.8 mm, 35%), and K-feldspar (subhedral–anhedral, 0.4–0.8 mm, ~30%), with subordinate hornblende (subhedral–anhedral, 0.2–0.8 mm, 15%), biotite (subhedral–anhedral, 0.2–0.4 mm, ~7%), clinopyroxene (subhedral–anhedral, 0.2–0.4 mm, <5%), and quartz (anhedral, 0.1–0.2 mm, <3%). The accessory minerals are magnetite, titanite, zircon, and apatite (Fig. 2e).

The gabbros in the Tiezhai complex are again fine-grained and massive. They contain euhedral–subhedral plagioclase (0.1–0.9 mm, ~40 vol.%), subhedral–anhedral K-feldspar (0.2–0.8 mm, ~20 vol.%), and euhedral–subhedral pyroxene (0.2–1.0 mm, ~20 vol.%). Most of the pyroxene is clinopyroxene. The gabbros also contain subhedral–anhedral hornblende (0.2–0.6 mm, ~15%), biotite (0.2–0.5 mm, ~5 vol.%), and minor quartz with accessory magnetite, titanite,

zircon, and apatite (Fig. 2f).

Whole-rock major element compositions were analyzed in the Modern Analysis Center (MAC) at Nanjing University by X-ray fluorescence (XRF) on fused glass beads, with precision better than $\pm 5\%$. Whole-rock trace element contents were determined using a Finnigan Element II inductively coupled plasma mass spectrometry (ICP-MS) in the SKLMDR at Nanjing University. Detailed analytical procedures are described in Gao et al. (2003). Analytical precision is better than $\pm 10\%$, with the majority better than $\pm 5\%$.

Whole-rock Rb–Sr, Sm–Nd and Pb isotopic compositions were measured using a Finnigan Triton TI TIMS in the SKLMDR, following the procedure described in detail by Pu et al. (2004, 2005). The mass fractionation corrections for $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were based on $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, respectively. Repeated analysis of the Nd standard JNDi-1 during the analytical session yielded $^{143}\text{Nd}/^{144}\text{Nd} = 0.512117 \pm 0.000005$ (2σ , $n = 12$), and the Sr standard NBS-987 yielded $^{87}\text{Sr}/^{86}\text{Sr} = 0.710260 \pm 0.000010$ (2σ , $n = 30$). Total analytical blanks were 5×10^{-11} g for Sm and Nd, and $(2\text{--}5) \times 10^{-10}$ g for Rb and Sr. Analytical reproducibility of $\pm 0.01\%$ (2σ) for $^{206}\text{Pb}/^{204}\text{Pb}$, $\pm 0.01\%$ for $^{207}\text{Pb}/^{204}\text{Pb}$ and $\pm 0.02\%$ for $^{208}\text{Pb}/^{204}\text{Pb}$ was attained in this study. Mass fractionation was calibrated against the NBS-981 standard (Todt et al., 1996).

Single zircon grains for LA–ICP–MS U–Pb dating were separated using conventional magnetic and heavy liquid separation methods, and then handpicked under a binocular microscope. Zircon grains were mounted in epoxy, and then polished for scanning electron microscope–cathodoluminescence (SEM–CL) imaging. CL images were taken in the State Key Laboratory of Continental Dynamics at Northwest University, Xi'an. Zircon U–Th–Pb isotopes were analyzed using an Agilent 7500a ICP–MS, coupled with a 213nm wave-length laser

microprobe in the SKLMDR. Analytical procedures are similar to that described in Jackson et al. (2004) and all the results were calibrated against the zircon standard, GEMOC/GJ-1 (609 Ma). Common Pb was corrected according to the method introduced by Andersen (2002). The analyses presented here have been corrected assuming recent lead loss. Data processing was carried out using the Isoplot/Ex 2.49 program of Ludwig (2001).

In situ Hf isotopic analyses of the above zircons were conducted on spots adjacent to U–Pb age determinations spots. Hf isotopic compositions were determined with a Thermo Scientific Neptune Plus Multi Collector ICP–MS coupled to a New Wave UP193 solid-state laser ablation system in the SKLMDR. Zircons were ablated with a beam diameter of 35 μm , an 10 Hz laser repetition rate, and with an energy of 10.91 to 11.05 J/cm². During Hf isotopic analysis, the Mud Tank standard was analyzed in every run as an independent control on reproducibility and instrument stability. Mud Tank zircon has an average $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282522 ± 0.000042 (2σ , $N = 2335$) (Griffin et al. 2007). Repeated measurements of this zircon standard during our study yielded $^{176}\text{Hf}/^{177}\text{Hf} = 0.282514 \pm 0.000012$ (2σ , $N = 16$), which is identical to the previously reported literature value. Detailed procedures are provided by Hou et al. (2007a).

In situ O isotope analyses of zircons were conducted on spots adjacent to U–Pb age determination spots. O isotopic compositions were determined with a Sensitive High-Resolution Ion Micro-Probe Stable Isotope (SHRIMP SI) ion microprobe in Research School of Earth Sciences, the Australian National University (Ireland et al. 2008). Analytical conditions were similar to those outlined in detail by Ickert et al. (2008). Corrected $^{18}\text{O}/^{16}\text{O}$ ratios are reported in $\delta^{18}\text{O}$ notation, in permil variations relative to Vienna standard mean ocean water (VSMOW). All $\delta^{18}\text{O}$ values were calibrated against standard zircon FC-1 (5.61‰; J.W. Valley, unpublished data).

The spot-to-spot reproducibility (external precision) was typically better than $\pm 0.50\%$ (2σ). After calibration, average $\delta^{18}\text{O}$ for secondary standard zircon Temora-2, varying from 7.97 to 8.50‰, is identical within analytical uncertainty to 8.20‰ that reported by Valley (2003) and Black et al. (2004).

4. Results

4.1. LA-ICP-MS zircon U-Pb ages

Zircons from syenite porphyry (sample TZ-2 and TZ-5), gabbro (sample TZ-13) and monzonite (sample TZ-16) are between 100–300 μm long (Fig. 3), with aspect ratios of 1.2:1 to 3:1. Most are euhedral–subhedral and are gray in CL images with wide oscillatory zoning in the cores and narrow oscillatory zoning in the rims. Most crystals from the syenite porphyry and the gabbro have white cores mantled by gray overgrowths (Fig. 3a, b, c), whereas the zircons from the monzonite (sample TZ-16) rarely have white cores (Fig. 3d).

Zircon LA-ICP-MS U-Pb ages have been determined for the syenite porphyry, gabbro, and monzonite; the results are listed in Table S1 and illustrated in Fig. 4. The Th/U ratios in the CL-dark rims and CL-white cores are 0.06–2.54 (mostly > 0.1) and 0.15–5.05, respectively. Ages of the rims of zircons from the syenite porphyry samples TZ-2 and TZ-5 are concordant with weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 130.4 ± 1.1 Ma (2σ , $n = 17$) and 129.9 ± 1.0 Ma (2σ , $n = 16$), respectively (Fig. 4a, b). These ages represent the time syenite magma crystallized. Ages of zircon rims from the gabbro (sample TZ-13) and the monzonite (sample TZ-16) are concordant with weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 128.5 ± 1.5 Ma (2σ , $n = 15$) and 128.8 ± 0.9 Ma (2σ , $n = 20$), respectively (Fig. 4c, d). These ages are slightly younger than the syenite porphyry. Analyses conducted on white cores and inherited zircons from three samples (TZ-2, TZ-5 and TZ-13)

yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2.95–2.34 Ga (mostly ~2.5 Ga; Fig. 4 and Table S1), indicating that Mesoarchean–Paleoproterozoic inherited zircons were captured in the syenite porphyry and gabbro.

4.2. Whole-rock major and trace elements

Results of major and trace element analyses of the Tiezhai complex are listed in Table S2. The syenite porphyry has a limited range of SiO_2 contents (63.5–65.5 wt.%) and total-alkalis ($\text{K}_2\text{O} + \text{Na}_2\text{O}$) contents (10.4–11.1 wt.%). The mafic rocks have lower SiO_2 (48.4–58.4 wt.%) and total-alkalis (5.58–8.07 wt.%). All the syenite porphyry data plot in the syenite field on the $\text{K}_2\text{O} + \text{Na}_2\text{O}$ vs. SiO_2 diagram (Fig. 5a), whereas the mafic rocks plot in the fields for gabbro (TZ-13), monzodiorite (TZ-15), or monzonite (TZ-14, TZ-16, and TZ-17) (Fig. 5a). Both the syenite porphyry and the mafic rocks are metaluminous, with alumina saturation index [$\text{Al}_2\text{O}_3/(\text{Ca}_2\text{O} + \text{Na}_2\text{O} + \text{K}_2\text{O})$] values of 0.70–0.94 (Table S2). The syenite porphyry has higher K_2O contents (3.50–4.26 wt.%) than the mafic rocks (1.44–2.65 wt.%), with most samples plotting in the high-K calc-alkaline field on the K_2O vs. SiO_2 diagram (Fig. 5b).

On the chondrite-normalized rare earth elements (REEs) plot (Fig. 6a) and the primitive mantle-normalized spidergram (Fig. 6b), the patterns formed by the analytical data from the Tiezhai mafic rocks are similar, indicating that their magmas may have originated from the same source. The syenite porphyry and mafic rocks are enriched in light REEs (LREEs), with $(\text{La}/\text{Yb})_{\text{N}}$ ratios of 24.8–36.2 and 6.85–9.92, respectively (Fig. 6a; Table S2). The mafic rocks have higher contents of heavy REEs (HREEs) than the syenite porphyry (Fig. 6a). Neither the Tiezhai syenite porphyry nor the mafic rocks have obviously Eu anomalies, with δEu values for those two rock types ranging from 1.04–1.28 and 0.79–0.89, respectively (Fig. 6a). On the primitive

mantle-normalized spidergram (Fig. 6b), both syenite porphyry and mafic rocks are enriched in large ion lithophile elements (LILEs, e.g. Rb, Sr, Ba) and depleted in high field strength elements (HFSEs, e.g. Nb, Ta, P and Ti), as is typical of subduction-related igneous rocks. In addition, the syenite porphyry has high Sr/Y (196–292) and (La/Yb)_N ratios, and low Y (6.05–7.61 ppm) and Yb (0.53–0.65 ppm) contents, suggesting an affinity with adakites (Fig. 7a, b; Defant and Drummond, 1990).

4.3. Whole-rock Sr–Nd–Pb isotopic compositions

Whole-rock Sr–Nd–Pb isotopic compositions of the Tiezhai complex rocks are listed in Table 1 and illustrated in Figs. 8 and 9. The Tiezhai syenite porphyry has relatively narrow ranges of measured $^{87}\text{Sr}/^{86}\text{Sr}$ (0.705960–0.706120) and $^{143}\text{Nd}/^{144}\text{Nd}$ (0.512101–0.512210) ratios, with calculated ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ratios of 0.7057–0.7059 and $\epsilon_{\text{Nd}}(t)$ values of –8.9 to –6.8 (Fig. 8). It also has relatively high and limited ranges of radiogenic Pb isotopic compositions with ($^{206}\text{Pb}/^{204}\text{Pb}$)_i = 18.36–18.37, ($^{207}\text{Pb}/^{204}\text{Pb}$)_i = 15.54–15.55, and ($^{208}\text{Pb}/^{204}\text{Pb}$)_i = 38.25–38.31 (Table 1; Fig. 9).

Based on their Sr–Nd–Pb isotopic compositions, the Tiezhai mafic rocks can be divided into two groups, 1 and 2. Group 1 mafic rocks include the monzodiorite (sample TZ-15) and one monzonite (TZ-16). These two specimens have relatively narrow ranges of measured $^{87}\text{Sr}/^{86}\text{Sr}$ (0.704843–0.704993) and $^{143}\text{Nd}/^{144}\text{Nd}$ (0.512162–0.512205) ratios, with calculated ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ratios of 0.7046–0.7047 and $\epsilon_{\text{Nd}}(t)$ values of –8.0 to –7.1 (Fig. 8). They also have relatively low and limited ranges of radiogenic Pb isotopic compositions with ($^{206}\text{Pb}/^{204}\text{Pb}$)_i = 17.68–17.71, ($^{207}\text{Pb}/^{204}\text{Pb}$)_i = 15.44–15.45, and ($^{208}\text{Pb}/^{204}\text{Pb}$)_i = 37.72–37.98 (Table 1; Fig. 9).

Group 2 mafic rocks include the gabbro (TZ-13) and one monzonite (TZ-17) specimen. These specimens have relatively wide ranges of measured $^{87}\text{Sr}/^{86}\text{Sr}$ (0.707608–0.709832) and

$^{143}\text{Nd}/^{144}\text{Nd}$ (0.512019–0.512149) ratios, with calculated $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios of 0.7072–0.7096 and $\epsilon_{\text{Nd}}(t)$ values of –10.8 to –8.2 (Fig. 8). They have relatively high and limited ranges of radiogenic Pb isotopic compositions with $(^{206}\text{Pb}/^{204}\text{Pb})_i = 18.08\text{--}18.12$, $(^{207}\text{Pb}/^{204}\text{Pb})_i = 15.52\text{--}15.53$, and $(^{208}\text{Pb}/^{204}\text{Pb})_i = 40.17\text{--}40.60$ (Table 1; Fig. 9). Compared with group 1, group 2 mafic rocks have higher $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios, higher radiogenic Pb isotopic compositions, and slightly lower $\epsilon_{\text{Nd}}(t)$ values, indicating they may have assimilated material that had high $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios, high radiogenic Pb, and low $\epsilon_{\text{Nd}}(t)$.

4.4. In situ zircon Hf–O isotopic compositions

In situ zircon Hf–O isotopic data for the Tiezhai syenite porphyry (sample TZ-2) and gabbro (TZ-13) are listed in Table 2 and plotted in Fig. 10. Rims on zircon in the Tiezhai syenite porphyry (TZ-2) have relatively narrow ranges of Hf isotopic contents, with $\epsilon_{\text{Hf}}(t)$ values (age-corrected using the U–Pb ages for the individual zircon grains) of –4 to +2 (except two spots; Fig. 10a). Zircon rims have relatively high $\delta^{18}\text{O}$ values of 6.9‰–9.0‰ (mean $8.2 \pm 0.8\text{‰}$; Table 2). The cores of inherited Paleoproterozoic–Mesoarchean zircons in the syenite porphyry have relatively narrow ranges of $\epsilon_{\text{Hf}}(t)$ (+2.6 to +10) and $\delta^{18}\text{O}$ values (5.8‰ to 6.9‰, mean $6.4 \pm 0.3\text{‰}$; Table 2).

Rims on zircon from the Tiezhai gabbro (sample TZ-13) also show a relatively narrow range of Hf isotopic compositions, with the majority of the $\epsilon_{\text{Hf}}(t)$ values in the –4 to +2 range (except for three spots; Fig. 10d; Table 2). Tiezhai gabbro zircon rims have relatively high oxygen isotope values, with $\delta^{18}\text{O}$ values ranging from 7.7 ‰ to 9.1‰ (mean $8.4 \pm 0.4\text{‰}$). Paleoproterozoic–Mesoarchean inherited zircons/cores from the gabbro have relatively narrow ranges of $\epsilon_{\text{Hf}}(t)$ (–0.5 to +5.6) and $\delta^{18}\text{O}$ values (5.7‰ to 6.9‰, mean $6.1 \pm 0.5\text{‰}$; Table 2).

5. Discussion

5.1. Petrogenesis

According to their similar intrusive ages (gabbro sample TZ-13, 128.5 ± 1.5 Ma; monzonite sample TZ-16, 128.8 ± 0.9 Ma), normalized REE and trace element spidergrams (Fig. 6), and close spatial relationship, we regard that gabbro, monzodiorite and monzonite are the members of one intrusion, despite their different major element compositions. The Tiezhai syenite porphyries also have similar intrusive ages (sample TZ-2, 130.4 ± 1.1 Ma; sample TZ-5, 129.9 ± 1.0 Ma), major and trace elements patterns (Figs. 5 and 6), and whole-rock Sr-Nd-Pb isotopic compositions (Figs. 8 and 9), indicating that their magmas came from the same source.

5.1.1 Fractional crystallization

The Tiezhai syenite porphyry exhibit negative correlations between SiO_2 and CaO, Ba, and Sr (Fig. 11e, i, j), and positive correlations between CaO and Sr, and δEu (Fig. 11m, n), suggesting that fractional crystallization of plagioclase took place in the parental magmas. Negative correlations between SiO_2 and MgO , FeO^T , CaO, Cr, and Ni (Fig. 11d, e, f, k, l) for the Tiezhai mafic rocks suggest olivine and clinopyroxene may have been fractionated during magma evolution. The obvious decrease of P_2O_5 and TiO_2 with increasing SiO_2 in the Tiezhai mafic rocks (Fig. 11g, h) indicates fractional crystallization of P- and Ti-rich minerals such as apatite, monazite, titanite, and ilmenite.

5.1.2 Crustal assimilation

The very old inherited zircons/cores preserved in the syenite porphyry and gabbro (Figs. 3 and 4) indicate that ancient crustal material must have participated in the formation of the Tiezhai complex. The Mesoarchean–Paleoproterozoic ages of the inherited zircons/cores correspond to the ages of zircons from Neoproterozoic gneisses of the NCC (Wang et al., 2009), and the $\varepsilon_{\text{Hf}}(t)$ values of

the Tiezhai inherited zircons/cores are mostly within the 2.5–3.0 Ga continental-crust evolution lines (Fig. 12a), further suggesting that ancient crustal material were derived from the NCC basement rocks.

The Hf isotopic compositions of the rims on zircons from the Tiezhai complex are above the 2.5–3.0 Ga continental upper crust and mafic lower crust evolution lines (Fig. 12a), and the $\delta^{18}\text{O}$ values of the zircon rims are higher than those of the inherited zircons/cores (Table 2; Fig. 12b). Therefore, it is reasonable to conclude that the Tiezhai complex was not derived directly from partial melting of NCC ancient crust. The relatively homogeneous chemical and zircon Hf–O isotopic compositions of the Tiezhai complex indicate that magma mixing played an insignificant role in the development of the Tiezhai magma (Table 2; Fig. 10). The lack of correlation between $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios and $\epsilon_{\text{Nd}}(t)$ values with SiO_2 contents for the syenite porphyry and mafic rocks (Fig. 12c, d), suggests that the magmas that formed these intrusions assimilated continental crustal material might via melting, assimilation, storage, and homogenization (MASH) processes rather than by assimilation and fractional crystallization (AFC) processes. As discussed in section 4.3, group 2 mafic rocks (specimens TZ-13 and TZ-17) have higher $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios, higher radiogenic Pb isotopic compositions, and slightly lower $\epsilon_{\text{Nd}}(t)$ values than group 1 mafic rocks (specimens TZ-15 and TZ-16), indicating that group 2 may have assimilated more ancient crust material (with high $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios, high radiogenic Pb isotopic compositions, and low $\epsilon_{\text{Nd}}(t)$ values). In addition, sample TZ-13 contains abundant inherited zircons/cores (Fig. 3c), whereas sample TZ-16 contains almost no (Fig. 3d), further indicating that the group 2 mafic rocks may have assimilated more ancient crust than group 1 rocks.

The ancient lower crust of the NCC has relatively low $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios (Huang et al., 2004)

and low radiogenic Pb isotopic compositions (Ying et al., 2010), whereas the upper crust has high and varied ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ratios (Jahn et al., 1988), and high radiogenic Pb isotopic compositions (Li et al., 2001). This indicates that the Tiezhai complex assimilated varied amounts of NCC upper crust material. The extremely high ($^{208}\text{Pb}/^{204}\text{Pb}$)_i ratios of group 2 mafic rocks may have been inherited from the NCC upper crust, which was possibly affected by a Th-enrichment event in the source area.

The $\delta^{18}\text{O}$ values for the inherited zircons/cores from the Tiezhai complex are 5.7‰–6.9‰ (mean 6.4‰), assuming that the protolith for the inherited zircons/cores was upper crust with an average SiO_2 content of 53.4 wt.% (Rudnick and Gao, 2003). Taking the SiO_2 content of the host rocks into account and using the linear function for the relationship between $\Delta^{18}\text{O}_{(\text{zircon} - \text{WR})}$ and SiO_2 composition (Lackey et al., 2008): $\Delta^{18}\text{O}_{(\text{zircon} - \text{WR})} \approx -0.0612 \times (\text{wt.\% SiO}_2) + 2.5$, the $\delta^{18}\text{O}$ values for the upper crust material are calculated to be 6.5‰–7.7‰ (mean 7.2‰), similar to the minimum value for upper continental crust $\delta^{18}\text{O}$ (7‰–30‰; Kempton and Harmon, 1992; Eiler et al., 2001; Simon and Lécuyer, 2005).

5.1.3 Origin of the Tiezhai mafic rocks

The low SiO_2 contents (48.4–58.4 wt.%) of the Tiezhai mafic rocks suggests that they were derived from an ultramafic source. This view is supported by high MgO (4.21–7.49 wt.%), Cr (23.5–149.3 ppm), and Ni (35.2–59 ppm) contents of these rocks, and their Mg# (54–61). Crustal rocks can be excluded as possible sources because experimental studies indicate that the partial melting of older exposed crustal rocks, or intermediate granulites in the deep crust, would produce high-Si granitoid rocks (Hirajima et al., 1990; Zhang et al., 1995; Kato et al., 1997; Rapp et al., 2003). Furthermore, the whole-rock Sr–Nd–Pb isotopic compositions of the Tiezhai mafic rocks

are quite different from those of the NCC and YB crust (Jahn et al., 1988, 1999; Figs. 8 and 9).

The isotopic composition of the Tiezhai rocks also precludes their formation by direct re-melting of crustal sources. Thus, it seems that the Tiezhai mafic rocks probably originated in the mantle.

The zircon $\epsilon_{\text{Hf}}(t)$ values of the inherited-zircon-rich Tiezhai gabbro (with more assimilated upper crustal material) are mostly -4 to $+2$ (Fig. 10d; Table 2), slightly lower than inherited-zircon-free gabbrodiorite sample (LQTZ10-12, with less assimilated upper crustal material; Zhong and Huang, 2012; Fig. 10c). Therefore, the $\epsilon_{\text{Hf}}(t)$ values for the primary magma of the Tiezhai mafic rocks can be represented by an inherited-zircon-free gabbrodiorite sample ($\epsilon_{\text{Hf}}(t)$ value of about -1 ; Fig. 10c), with these values also being higher than those of the Luxi EM1-derived mafic rocks ($\epsilon_{\text{Hf}}(t) = -20$ to -16 ; Huang et al., 2012; Fig. 10d), the Luxi EM2-derived mafic rocks ($\epsilon_{\text{Hf}}(t) = -18$ to -16 ; Xu et al., 2007; Lan et al., 2011; Yang et al., 2012b; Guo et al., 2013c, 2014; Fig. 10b), and the Luxi EM1 ($\epsilon_{\text{Hf}}(t)$ about -9 ; Yang et al., 2006; Fig. 10).

The Sr–Nd isotopic compositions of Tiezhai group 1 mafic rocks (that assimilated less crust) are similar to those of the Luxi EM1-derived mafic rocks (Fig. 8), indicating that the primary magma of the Tiezhai mafic rocks may have been derived from partial melting of Luxi EM1 units. However, the radiogenic Pb isotopic compositions of group 1 mafic rocks are higher than those of the Luxi EM1-derived mafic rocks, and the primary magma of the Tiezhai mafic rocks had higher $\epsilon_{\text{Hf}}(t)$ values than Luxi EM1-derived mafic rocks. Therefore, we conclude that the sources of the Tiezhai mafic rocks must have contained some asthenospheric mantle material with high $\epsilon_{\text{Hf}}(t)$ values, and high radiogenic Pb isotopic compositions.

As discussed in section 5.1.2, the primary magmas of the Tiezhai mafic rocks may have assimilated different amounts of NCC upper crustal material. The $\delta^{18}\text{O}$ values of zircon rims in the

Tiezhai gabbro are 7.7‰–9.1‰ (mean 8.4‰; Table 2). Based on the SiO₂ content of the host rock (Lackey et al., 2008), the $\delta^{18}\text{O}$ values for the Tiezhai gabbro are estimated to be 8.2‰–9.6‰ (mean 8.9‰), and values for the upper crust are 6.5‰–7.7‰ (mean 7.2‰). The $\delta^{18}\text{O}$ values of the primary magma for the Tiezhai gabbro are greater than 8.2‰–9.6‰, and thus higher than values for; a) Luxi EM1-derived mafic rocks (Jinan and Zouping gabbros, $\delta^{18}\text{O} = 5.8\text{‰}–7.1\text{‰}$; Yang et al., 2012a; Guo et al., 2013a), b) Luxi EM2-derived mafic rocks (Yinan gabbro, $\delta^{18}\text{O} = 6.5\text{‰}–7.0\text{‰}$; Yang et al., 2012b), c) Luxi EM2 (Fangcheng and Feixian mantle xenolith, $\delta^{18}\text{O} = 6.9\text{‰}–8.9\text{‰}$; Guo et al., 2013b; Xu et al., 2013; Fig. 13), and d) asthenospheric mantle ($5.7 \pm 0.5\text{‰}$; Hoefs, 2009). Therefore, we conclude that the primary magma of the Tiezhai mafic rocks was derived from partial melting of Luxi EM1-asthenospheric mantle that had previously been metasomatized by fluids or melts with high $\delta^{18}\text{O}$ values. Basaltic rocks and sediments in the upper part of the oceanic crust typically have high $\delta^{18}\text{O}$ (~9‰–20‰; Bindeman et al., 2005), and these rocks therefore seem to be the most likely source for the metasomatic fluids/melts. The magma derived from partial melting of mantle that has been previously metasomatized by subducted oceanic-crust-derived melts always has an affinity with adakite (Martin et al, 2005), and this is inconsistent with our observations. Therefore, the agent of metasomatism was possibly fluids derived from subducted upper oceanic crust and sediments, rather than melts.

Whole-rock $(^{87}\text{Sr}/^{86}\text{Sr})_{i-\text{Nd}}(t)$, $(^{206}\text{Pb}/^{204}\text{Pb})_{i-(^{207}\text{Pb}/^{204}\text{Pb})_i}$ and zircon $\varepsilon_{\text{Hf}}(t)$ – $\delta^{18}\text{O}$ modeling calculations indicate that magma derived from metasomatized Luxi EM1-asthenospheric mantle, with ~5% and 20%–30% assimilated NCC ancient upper crust material, produces whole-rock Sr–Nd–Pb and zircon Hf–O isotopic compositions similar to those of the Tiezhai group 1 and group 2 mafic rocks, respectively (Figs. 8, 9 and 12b).

Therefore, we conclude that the Tiezhai mafic rocks were derived from Luxi EM1-asthenospheric mantle that was previously metasomatized by subducted upper oceanic crust and sediments-derived fluids. The magmas that produced the Tiezhai mafic rocks assimilated different amounts of NCC ancient upper crust material during magma ascent (group 1 mafic magmas assimilated ~5%, and group 2 mafic magmas assimilated 20%–30%).

5.1.4 Origin of the Tiezhai syenite porphyry

The geochemical characteristics of the Tiezhai syenite porphyry are similar to that of adakites (Fig. 7; Defant and Drummond, 1990). Adakite is believed to be generated by the partial melting of subducted young and hot oceanic slab material (Defant and Drummond, 1990). However, some rocks with adakitic affinities are considered to have different origins, including (1) crustal AFC involving basaltic magma (Castillo et al., 1999; Macpherson et al., 2006; Gao et al., 2012; Chen et al., 2013); (2) partial melting of a thickened/delaminated mafic lower crust (Atherton and Petford, 1993; Petford and Atherton, 1996; Chung et al., 2003; Gao et al., 2004; Hou et al., 2004; Ma et al., 2013; Wang et al., 2016); and (3) low-degree partial melting of metasomatized mantle (Martin et al., 2005; Jiang et al., 2006).

The Tiezhai syenite porphyry has whole-rock $\epsilon_{\text{Nd}}(t)$ values, zircon $\epsilon_{\text{Hf}}(t)$ values, and $\delta^{18}\text{O}$ values similar to those of the Tiezhai mafic rocks (Figs. 8, 10 and 12b), indicating that the syenite porphyry might have been produced from the mafic rock's parent magma by fractional crystallization. High-pressure fractional crystallization (HPFC) precipitating garnet could efficiently raise the Sr/Y and La/Yb ratios, and also increase the Dy/Yb ratio (Macpherson et al., 2006). However, neither the syenite porphyry nor the mafic rocks show evidence of high-pressure fractional crystallization trends involving garnet (Fig. 14a, b). Lower-pressure fractional

crystallization (LPFC) precipitating hornblende could also raise the Sr/Y and La/Yb ratios, but unlike garnet fractionation, hornblende precipitation will decrease the Dy/Yb ratio (Davidson et al., 2008). In addition, in natural systems, hornblende fractionation is typically accompanied by plagioclase removal. Fractional crystallization of hornblende and plagioclase could increase the La/Yb ratio, decrease the Dy/Yb ratio, and moderate increases or even decrease the Sr/Y ratio. On the SiO₂ vs. Sr/Y diagram, neither the syenite porphyry nor the mafic rocks show hornblende fractional crystallization trends (Fig. 14a). On the SiO₂ vs. Dy/Yb diagram, only the mafic rocks show hornblende fractional crystallization trends (Fig. 14b). However, there is no negative correlation between SiO₂ and Al₂O₃ for the mafic rocks (Fig. 11c), indicating that insignificant hornblende fractional crystallization. Furthermore, many of the adakites generated by AFC process reported by Richards and Kerrich (2007) exhibit only limited ranges of Sr/Y ratios, typically from 30 to 60, which is inconsistent with the high Sr/Y ratio of the syenite porphyry (196–292). Therefore, we conclude that the syenite porphyry is not generated by AFC process involving basaltic magma.

The Tiezhai syenite porphyry has high SiO₂ contents, 63.5–65.5 wt.%, and low MgO contents, 0.84–1.08 wt.%, with insignificant fractional crystallization, much higher than the adakitic rocks derived from metasomatized mantle (SiO₂ <56 wt.%, MgO >4 wt.%; Martin et al., 2005). This precludes a metasomatized mantle origin for the Tiezhai syenite.

Whole-rock Sr–Nd–Pb and Zircon Hf isotopic signatures from the Tiezhai syenite porphyry are different than those of adakitic rocks derived from ancient thickened mafic lower crust or delaminated mafic lower crust in the Luxi terrane and Jiaodong Peninsula (Hou et al., 2007b; Liu et al., 2008b, 2009; Ma et al., 2013; Wang et al., 2016; Fig. 8). This observation suggested that the

syenite porphyry did not originate from ancient NCC lower crust, but might have been produced from a newly underplated thick lower crust. The similar whole-rock Nd and Zircon Hf–O isotopic compositions of the syenite porphyry and the mafic rocks indicating that the newly thick lower crust may have been created by Tiezhai mafic magmas underplating. The differences in the $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios between the syenite porphyry and the mafic rocks may result from varied crustal contamination. Whole-rock $(^{87}\text{Sr}/^{86}\text{Sr})_{i-\varepsilon_{\text{Nd}}(t)}$ modeling calculations indicate that the syenite porphyry magma might assimilate ~10% Neoarchean–Paleoproterozoic upper crust material (Fig. 8). The syenite porphyry has higher $(^{206}\text{Pb}/^{204}\text{Pb})_i$ and $(^{207}\text{Pb}/^{204}\text{Pb})_i$ ratios than the mafic rocks (Fig. 9), indicating that the source of the newly underplated thick lower crust may have more asthenospheric mantle component.

We conclude that the Tiezhai syenite porphyry was formed by partial melting of newly underplated thick lower crust, and the magma assimilated ~10% NCC Neoarchean–Paleoproterozoic upper crust material during its ascent.

5.2 Implication for destruction of the NCC

As mentioned in the Introduction, three main mechanisms have been proposed for the destruction of the NCC. These are thermo–mechanical–chemical erosion (Xu et al., 2004b, 2009; Zheng et al., 2007), delamination (Gao et al., 2004; Windley et al., 2010; Jiang et al., 2010; Ma et al., 2014a), and the addition of fluids to the overlying SCLM during subduction (Niu, 2005; Windley et al., 2010; Zhu et al., 2012).

The data presented in this study of the Tiezhai mafic rocks indicates that fluids derived from subducted oceanic crust+sediments may have metasomatized the overlying EM1 units. Therefore, we conclude that fluids added to the overlying SCLM during subduction weakened and thinned

the lithospheric mantle, raised the geothermal gradient, and thereby lowered the mantle solidus to trigger hydrous melting of the mantle source rocks (Niu, 2005; Windley et al., 2010). It follows that the introduction of fluids derived from subducted oceanic crust into the overlying SCLM during subduction was an important mechanism for destruction of the NCC.

Previous studies have shown that in eastern China, the Early Cretaceous was a time of wide development of metamorphic core complexes (Ratschbacher et al., 2000; Yang et al., 2007b), widespread A-type granites and alkaline rocks (Wang et al., 1995), and extensional sedimentary basins (Guo et al., 2005; Wang et al., 2005). These processes indicate that this was a period of extensional tectonics. This extensional tectonic framework in the lithospheric mantle and asthenosphere upwelling may have been inherited during post-collisional processes following the Triassic collision between the NCC and the YB (Zhao and Zheng, 2009; Yang et al., 2012a, 2012b) and the effects of paleo-Pacific subduction (Wu et al., 2005; Sun et al., 2007; Jiang et al., 2010; Ma et al., 2014a). It is also possible that a combination of both processes triggered this intense tectonic extension.

Based on global models of the evolution of orogenic belts, Ma et al. (2006) concluded that about 20–50 Myr elapses between collisional orogeny and orogen collapse. Therefore, Early Cretaceous magmatic activities cannot be explained by post-collision effects following the NCC–YB continental collision in the Triassic, for which the time lag is >100 Myr. In addition, Late Triassic magmatism, which may have been related to post-collision extension between the NCC and the YB, occurred only at the eastern margin of the NCC (Jiaodong and Liaodong peninsulas; Yang et al., 2005; 2007a, 2010; Fig. 15c), indicating that the collision affected only the

eastern margin. The points enumerated below indicate that subduction of the paleo-Pacific slab may have played a critical role in the extension of the eastern NCC during the Early Cretaceous.

1) The eastward younging trend of the Early Cretaceous large-scale magmatism beginning (Taihang, 146 Ma; Chen et al., 2004; Luxi terrane, 144 Ma; Liu et al., 2008a; Jiaodong Peninsula, 134 Ma; Yang et al., 2005; Fig. 15), peak (Taihang, 132–130 Ma; Luxi terrane, 130–128 Ma; Jiaodong Peninsula, 124–122 Ma; Fig. 15), and end (Taihang, 123 Ma; Li et al., 2014b; Luxi terrane, 119 Ma; Yang et al., 2012b; Jiaodong Peninsula, 108 Ma; Cai et al., 2013; Fig. 15) are consistent with the paleo-Pacific slab roll-back model. An eastward younging trend of magmatism has also been observed in the Great Xing'an Range, northeastern China (Wang et al., 2006; Zhang et al., 2010b) and North Korea (Wu et al., 2007). 2) The occurrence of adakites derived from an oceanic slab at Zouping (Wangjiazhuang quartz monzonite; Wang et al., 2015) and Mengyin (Wang et al., 2016) in the Luxi terrane and the Xu-Huai adakite in the southeastern NCC (Ling et al., 2013) provide evidence for the existence of the paleo-Pacific slab beneath the eastern NCC. 3) The NW–SE extensional direction is consistent with the NW-trending subduction of the paleo-Pacific plate (Wang et al., 2011; Zhu et al., 2012; Zhang et al., 2014).

Here, we propose the following tectonic model. The model is illustrated in Fig. 16.

(1) The onset of subduction of the paleo-Pacific slab beneath the eastern NCC began at ~180 Ma (Wu et al., 2007; Jiang et al., 2010; Ma et al., 2014a), forming a continental arc at the eastern margin of the NCC. The fluids derived from dehydration of the upper part of subducted oceanic crust metasomatized the overlying lithospheric and asthenospheric mantle (Fig. 16a).

(2) During 146–130 Ma, slab rollback beneath the Taihang Mountain belt caused back-arc extension, and upwelling of the asthenospheric mantle (Fig. 16b). Partial melting of the EM1

caused by upwelling asthenospheric heating generated voluminous mafic magmas (Chen et al., 2004; Gao et al., 2012; Li et al., 2013a; Zhang et al., 2017) and the associated Au-Ag-Cu-Mo ore deposits exposed in the Taihang Mountain (Li et al., 2013b, 2014b, 2015; Shen et al., 2013; Zhang et al., 2015).

(3) During 130–125 Ma, progressive slab rollback, combined with ridge subduction, caused extension of the lithospheric mantle and upwelling of the asthenospheric mantle beneath the Luxi terrane. The magma derived from partial melting of the metasomatized Luxi EM1-asthenospheric mantle underplated the lower crust to form new thick garnet-amphibolite lower crust. Until ca. 130 Ma, partial melting of this thick lower crust formed the primary adakitic magma that evolved into the Tiezhai syenite porphyry by assimilating ~10% of ancient NCC upper crust material during upwelling. Until ca. 129 Ma, partial melting of the metasomatized Luxi EM1-asthenospheric mantle generated mafic magmas, and these magmas assimilated varied amounts of ancient NCC upper crust material (group 1 mafic rocks assimilated ~5%; group 2 mafic rocks assimilated 20%–30%) to form the Tiezhai mafic rocks (Fig. 16c).

(4) During 130–125 Ma, rapidly increasing slab rollback, caused foundering of the EM2 and thinning of the lithosphere beneath the Jiaodong and Liaodong peninsulas (Fig. 16d). Partial melting of this delaminated lithospheric mantle and the decompression melting of upwelling asthenospheric mantle caused voluminous mafic magmatism on these peninsulas (Yang et al., 2005, 2008; Cai et al., 2013, 2016; Ma et al., 2014a; Meng et al., 2015). This was contemporaneous with the partial melting of crustal material as a result of basaltic magmatic underplating, that formed the widespread granitoids and associated gold deposits on these

peninsulas (Wu et al., 2005; Guo et al., 2013c; Li and Santosh, 2013, 2017a; Ma et al., 2013; Yang et al., 2012d, 2014; He et al., 2016; Cai et al., 2018).

6. Conclusions

(1) Zircon U–Pb dating results indicate that the igneous rocks of the Tiezhai complex were intruded into the Luxi terrane at 130–129 Ma. Both the syenite porphyry and the gabbro contain abundant Mesoarchean–Paleoproterozoic inherited zircon/cores with ages of 2.95–2.34 Ga.

(2) The Tiezhai mafic rocks were derived from partial melting of the Luxi EM1-asthenospheric mantle that has been previously metasomatized by fluids derived from subducted oceanic crust+sediments, and assimilated varied amounts of ancient NCC upper crust material during the magma ascent (group 1 mafic rocks assimilated ~5%; group 2 mafic rocks assimilated 20%–30%). The Tiezhai syenite porphyry is characterized by high Sr/Y and La/Yb ratios, and low Yb and Y contents, similar to adakites. Geochemical and Sr–Nd–Pb–Hf–O isotopic data indicate that the syenite porphyry was derived from partial melting of newly underplated thick lower crust, and that it assimilated ~10% of NCC Neoproterozoic–Paleoproterozoic upper crust material.

(3) The addition of fluids derived from oceanic crust and sediments to the overlying subcontinental lithospheric mantle during subduction is an important mechanism for lithosphere destruction in the eastern NCC.

(4) Early Cretaceous large-scale magmatism and lithospheric thinning in the eastern NCC may be related to the slab rollback accompanying ridge subduction of the paleo-Pacific slab beneath the eastern NCC.

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Figure captions

Fig. 1. Geological maps. (a) Geological and tectonic map of the North China Block, modified after Zhang et al. (2005) and Lan et al., (2011); (b) Geological map of the Luxi Terrane, modified after Zhang et al. (2007) and Lan et al. (2011); (c) Geological sketch map of the Tiezhai basin, modified after Zhong and Huang (2012).

Fig. 2. Representative photomicrographs (cross-polarized light) of thin sections of the Tiezhai complex. (a and b) syenite porphyry; (c and d) monzonite; (e) monzodiorite; (f) gabbro; Qtz, quartz; Pl, plagioclase; Kfs, K-feldspar; Hbl, hornblende; Bt, biotite; Cpx, clinopyroxene.

Fig. 3. Cathodoluminescence images of zircons from Tiezhai (a) syenite porphyry (sample TZ-2), (b) syenite porphyry (sample TZ-5), (c) gabbro (sample TZ-13) and (d) monzonite (sample TZ-16). Small circles mark LA-ICP-MS analysis spots for U-Pb dating, large red circles mark the sites of Hf isotopic analyses, and small yellow ellipses mark the sites of O isotopic analyses, with the $^{206}\text{Pb}/^{238}\text{U}$ or $^{206}\text{Pb}/^{207}\text{Pb}$ ages, $\varepsilon_{\text{Hf}}(t)$, and $\delta^{18}\text{O}$ values shown.

Fig. 4. Concordia diagram for zircons from the Tiezhai (a) syenite porphyry (sample TZ-2), (b) syenite porphyry (sample TZ-5), (c) gabbro (sample TZ-13) and (d) monzonite (sample TZ-16).

Fig. 5. Rock classification diagrams for the Tiezhai complex. (a) $\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. SiO_2 (Middlemost, 1994); (b) K_2O vs. SiO_2 (Peccerillo and Taylor, 1976).

Fig. 6. (a) Chondrite-normalized rare earth element patterns, and (b) primitive mantle-normalized spider diagrams for the Tiezhai complex. The chondrite and primitive mantle normalization values are from McDonough and Sun (1995).

Fig. 7. (a) Sr/Y vs. Y and (b) $(\text{La}/\text{Yb})_{\text{N}}$ vs. Yb_{N} for the Tiezhai complex. (a) is after Defant and Drummond (1990), (b) is after Drummond et al. (1996).

Fig. 8. $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs. $\epsilon_{\text{Nd}}(t)$ diagrams for the Tiezhai complex. Data sources: Luxi EM1-derived mafic rocks from the Jinan and Zouping gabbros (Guo et al., 2013a; Huang et al., 2012; Yang et al., 2012a); Luxi EM2-derived mafic rocks from the Fangcheng basalts (Zhang et al., 2002), Feixian basalts (Yang et al., 2012c), Yinan gabbros (Xu et al., 2004b), Tongjin diorite (Xu et al., 2004b), Longbaoshan monzonite–syenite (Zhang et al., 2005; Lan et al., 2011); Cenozoic subducted oceanic-crust-derived adakites (Defant and Drummond, 1990; Kay and Kay, 1993; Sajona et al., 2000; Aguillo'n-Robles et al., 2001); Early Cretaceous delaminated/thickened ancient lower-crust-derived adakitic rocks at Luxi and Jiaodong (Hou et al., 2007b; Liu et al., 2008b, 2009; Wang et al., 2016). The lower and upper crust of the NCC field, and lower crust of the YB field are from Jahn et al. (1999); the P-MORB (Pacific MORB) field is from Barry et al. (2003) and Vervoort et al. (1999); the depleted mantle (DM) field is from Zindler and Hart (1986). All the $(^{87}\text{Sr}/^{86}\text{Sr})_i$ and $\epsilon_{\text{Nd}}(t)$ values are recalculated to 130 Ma. Also shown are the binary mixing curves between the Luxi EM1+asthenospheric mantle-derived melts and NCC upper crust material (UC). The mixing parameters are assumed to be:

	$(^{87}\text{Sr}/^{86}\text{Sr})_i$	Sr (ppm)	$\epsilon_{\text{Nd}}(t)$	Nd (ppm)
NCC UC	0.730	320	−25	27
Luxi EM1+asthenospheric mantle-derived melts	0.704	600	−6.9	30

- (1) Sr–Nd isotopes: UC, Taishan group (Jahn et al., 1988); Luxi EM1+asthenospheric mantle-derived melts, assumed.
- (2) Sr–Nd concentrations: UC, average continental upper crust (Rudnick and Gao, 2003); Luxi EM1+asthenospheric mantle-derived melts, assumed.

Fig. 9. $(^{207}\text{Pb}/^{204}\text{Pb})_i$ vs. $(^{206}\text{Pb}/^{204}\text{Pb})_i$ diagram for the Tiezhai complex. Data sources: Luxi EM1-derived mafic rocks from the Jinan and Zouping gabbros (Li et al., 2007; Yang et al., 2012c); Luxi EM2-derived mafic rocks from the Fangcheng basalts (Zhang et al., 2002), Feixian basalts (Yang et al., 2012c), Yinan gabbros (Xu et al., 2004b; Yang et al., 2012b), Longbaoshan monzonite–syenite (Zhang et al., 2005; Lan et al., 2011). The fields for P–MORB (Pacific MORB) are from Vervoort et al (1999). All $(^{207}\text{Pb}/^{204}\text{Pb})_i$, $(^{206}\text{Pb}/^{204}\text{Pb})_i$ and $(^{87}\text{Sr}/^{86}\text{Sr})_i$ values were recalculated to 130 Ma. Also shown are the binary mixing curves between the Luxi EM1+asthenospheric mantle-derived melts and NCC UC. The mixing parameters are assumed as follows:

	$(^{206}\text{Pb}/^{204}\text{Pb})_i$	$(^{207}\text{Pb}/^{204}\text{Pb})_i$
NCC UC	19.6	15.75
Luxi EM1+asthenospheric mantle-derived melts	17.59	15.44

Pb isotopes: UC, Taishan Group (Jhan et al., 1988; Li et al., 2001); Luxi EM1+asthenospheric mantle-derived melts, assumed.

Fig. 10. Frequency distributions of $\varepsilon_{\text{Hf}}(t)$ values for zircons from (a) Tiezhai syenite porphyry, (b) Luxi EM2-derived mafic rocks, (c) Tiezhai gabbrodiorite, (d) Tiezhai gabbro and Luxi EM1-derived mafic rocks. All the $\varepsilon_{\text{Hf}}(t)$ values were recalculated to 130 Ma. In (b), the Luxi EM2-derived mafic rocks are from the Shangyu gabbros (Xu et al., 2007; Yang et al., 2012b), the Longbaoshan monzonite–syenite (Lan et al., 2011) and the Tongjin diorites (Xu et al., 2007; Guo et al., 2013c). In (c), the Tiezhai gabbrodiorite is from Zhong and Huang (2012). In (d), the Luxi EM1-derived mafic rocks are from the Jinan and Zouping gabbros (Huang et al., 2012)

Fig. 11. Harker diagrams for the Tiezhai complex.

Fig. 12. (a) Zircon $\epsilon_{\text{Hf}}(t)$ vs. T, (b) zircon $\delta^{18}\text{O}$ vs. $\epsilon_{\text{Hf}}(t)$, (c) $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs. SiO_2 , and (d) $\epsilon_{\text{Nd}}(t)$ vs.

SiO_2 for the Tiezhai complex. In (b), also shown is the mixing trends between the NCC upper crust material ($\epsilon_{\text{Hf}}(t) = -40$, $\delta^{18}\text{O} = 6.3\text{‰}$; calculated from inherited zircons in Tiezhai complex) and melts derived from Luxi EM1+asthenospheric mantle ($\epsilon_{\text{Hf}}(t) = +2$, $\delta^{18}\text{O} = 9.5\text{‰}$), and depleted mantle material ($\epsilon_{\text{Hf}}(t) = +15$, $\delta^{18}\text{O} = 5.3\text{‰}$). The $\text{Hf}_{\text{ea}}/\text{Hf}_{\text{uc}}$ term signifies the ratio between the Hf concentration in the Luxi EM1+asthenospheric mantle (ea) and the Hf concentration in the NCC upper crust (uc). The $\text{Hf}_{\text{dm}}/\text{Hf}_{\text{uc}}$ term signifies the ratio between the Hf concentration in the depleted mantle (dm) and the Hf concentration in the NCC upper crust (uc). The small open circles along the curve indicate 10% mixing increments.

Fig. 13. Comparison of whole-rock $\delta^{18}\text{O}$ values for the Tiezhai syenite porphyry, gabbro, and assimilated NCC upper crust with calculated $\delta^{18}\text{O}$ values for rocks in the NCC. The whole-rock $\delta^{18}\text{O}$ values for the Tiezhai syenite porphyry and gabbro were calculated from zircon rim $\delta^{18}\text{O}$ values and whole-rock SiO_2 contents. Whole-rock $\delta^{18}\text{O}$ values for assimilated NCC upper crust were calculated from $\delta^{18}\text{O}$ values of inherited zircons/cores from the Tiezhai syenite porphyry and gabbro, and average SiO_2 content of upper continental crust (53.4 wt.%; Rudnick and Gao, 2003). Data sources: normal mantle ($5.7 \pm 0.5\text{‰}$; Hoefs, 2009); upper portion of the oceanic crust (9‰–20‰; Bindeman et al., 2005); upper crust (7‰–30‰; Eiler et al., 2001; Kempton and Harmon, 1992; Simon and Lécuyer, 2005); NCC lower crust (4.7‰–8.9‰; Zhang, 2014); Luxi EM1-derived mafic rocks (Jinan and Zouping gabbros; 5.8‰–7.1‰; Yang et al., 2012a; Guo et al., 2013a); Luxi EM2-derived mafic rocks

(Yinan gabbros; 6.5‰–7.0‰; Yang et al., 2012b); Luxi EM2 (Feixian websterite xenolith; 7.5‰–8.9‰; Xu et al., 2013; Fangcheng pyroxenite xenolith; 6.9‰–7.4‰; Guo et al., 2013b).

Fig. 14. (a) Sr/Y vs. SiO₂, and (b) Dy/Yb vs. SiO₂ for the Tiezhai complex.

Fig. 15. Frequency histograms showing the ages of mantle-derived and crust-derived rocks from (a) Taihang Mountain, (b) Luxi terrane and (c) Jiaodong Peninsula. Data sources are list in Table S3.

Fig. 16. Schematic diagram showing the generation and emplacement of the Tiezhai complex (modified from Liu et al., 2015).

Tables

Table 1. Sr–Nd–Pb isotopic data for the Tiezhai complex.

Table 2. Zircon Lu–Hf and O isotope data for the Tiezhai complex.

Table 1
Sr–Nd–Pb isotopic data for samples from the Tiezhai complex.

Sample	TZ-2	TZ-5	TZ-7	TZ-8	TZ-11	TZ-13	TZ-15	TZ-16	TZ-17
Rock type	Syenite porphyry					Gabbro	Monzodiorite	Monzonite	
Age(Ma)	130	130	130	130	130	129	129	129	129
Rb(ppm)	73.0	74.0	73.8	73.0	81.2	39.8	42.6	52.0	40.6
Sr(ppm)	1542	1378	1755	1675	1494	776	637	1131	545
$^{87}\text{Rb}/^{86}\text{Sr}$	0.1369	0.1554	0.1216	0.126	0.1572	0.1483	0.1936	0.1329	0.2157
$^{87}\text{Sr}/^{86}\text{Sr}(2\sigma)$	0.706120(10)	0.705968(11)	0.705960(5)	0.706063(5)	0.705965(6)	0.709832(6)	0.704993(10)	0.704843(6)	0.707608(5)
$(^{87}\text{Sr}/^{86}\text{Sr})_i$	0.7059	0.7057	0.7057	0.7058	0.7057	0.7096	0.7047	0.7046	0.7072
Sm(ppm)	3.93	3.83	4.00	3.90	4.30	6.73	5.30	9.46	4.33
Nd(ppm)	24.0	22.7	22.7	23.8	24.9	35.3	27.3	49.2	22.3
$^{147}\text{Sm}/^{144}\text{Nd}$	0.0991	0.102	0.1063	0.099	0.1043	0.1151	0.1174	0.1161	0.1172
$^{143}\text{Nd}/^{144}\text{Nd}$	0.512101(12)	0.512210(6)	0.512188(19)	0.512201(5)	0.512198(7)	0.512149(7)	0.512205(9)	0.512162(6)	0.512019(9)
$\epsilon_{\text{Nd}}(t)$	-8.9	-6.8	-7.3	-6.9	-7.1	-8.2	-7.1	-8.0	-10.8
$T_{2\text{DM}}(\text{Ga})$	1.65	1.48	1.52	1.49	1.50	1.59	1.51	1.57	1.80
$^{206}\text{Pb}/^{204}\text{Pb}(2\sigma)$	18.5102(6)	18.5163(5)			18.4853(5)	18.2159(6)	17.8529(6)	17.8079(4)	18.1911(6)
$^{207}\text{Pb}/^{205}\text{Pb}(2\sigma)$	15.5564(6)	15.5586(6)			15.5566(5)	15.5334(6)	15.4581(5)	15.4466(9)	15.5312(6)
$^{208}\text{Pb}/^{206}\text{Pb}(2\sigma)$					38.426(1)	40.250(1)		37.891(1)	40.741(2)
$(^{206}\text{Pb}/^{204}\text{Pb})_i$	18.36	18.36	—	—	18.37	18.12	17.71	17.68	18.08
$(^{207}\text{Pb}/^{204}\text{Pb})_i$	15.55	15.54	—	—	15.55	15.53	15.45	15.44	15.52
$(^{208}\text{Pb}/^{204}\text{Pb})_i$	38.31	38.25	—	—	38.28	40.19	37.98	37.72	40.60

Chondrite Uniform Reservoir (CHUR) values ($^{87}\text{Rb}/^{86}\text{Sr} = 0.0847$, $^{87}\text{Sr}/^{86}\text{Sr} = 0.7045$, $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$, $^{143}\text{Nd}/^{144}\text{Nd} = 0.512638$) are used for the calculation. $\lambda_{\text{Rb}} = 1.42 \times 10^{-11} \text{ year}^{-1}$, $\lambda_{\text{Sm}} = 6.54 \times 10^{-12} \text{ year}^{-1}$ (Lugmair and Marti, 1978). $\lambda_{\text{U}238} = 1.55125 \times 10^{-10} \text{ year}^{-1}$, $\lambda_{\text{U}235} = 9.8485 \times 10^{-10} \text{ year}^{-1}$, $\lambda_{\text{Th}232} = 4.9475 \times 10^{-11} \text{ year}^{-1}$ (Steiger and Jäger, 1977).

Table 2

Zircon Lu–Hf and O isotopic data for samples from the Tiezhai complex.

Spot	Age(Ma)	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm 1\sigma$	$\varepsilon_{\text{Hf}}(t)$	$T_{\text{DM1}}(\text{Ga})$	$T_{\text{DM2}}(\text{Ga})$	$(^{176}\text{Hf}/^{177}\text{Hf})$	$\delta^{18}\text{O}(\text{‰})$	± 2
Syenite porphyry (TZ-2)											
TZ-2-01	129	0.017628	0.000471	0.282747	0.00002	1.9	0.71	1.06	0.282746	8.1	0.1
TZ-2-02	2946	0.014848	0.000377	0.281198	0.00002	10.0	2.82	2.74	0.281177	5.8	0.1
TZ-2-03	128	0.010034	0.000305	0.282659	0.00003	-1.2	0.83	1.26	0.282659	7.5	0.1
TZ-2-04	2523	0.015069	0.000374	0.281318	0.00002	4.5	2.66	2.74	0.281300	6.7	0.1
TZ-2-05	130	0.011835	0.000327	0.282656	0.00002	-1.3	0.83	1.26	0.282656	8.3	0.1
TZ-2-06	2547	0.007792	0.000179	0.281337	0.00002	6.1	2.62	2.66	0.281328	6.4	0.1
TZ-2-07	132	0.013658	0.000371	0.282613	0.00002	-2.8	0.89	1.36	0.282612	8.6	0.1
TZ-2-08	129	0.014426	0.000393	0.282645	0.00002	-1.7	0.85	1.29	0.282644	7.0	0.1
TZ-2-10	130	0.017635	0.000478	0.282507	0.00003	-6.6	1.04	1.60	0.282505	7.1	0.1
TZ-2-11	2533	0.015295	0.000385	0.281359	0.00002	6.2	2.60	2.65	0.281340	6.9	0.1
TZ-2-12	130	0.013687	0.000393	0.282719	0.00001	0.9	0.75	1.12	0.282718	8.9	0.1
TZ-2-13	2338	0.010023	0.000236	0.281376	0.00002	2.6	2.57	2.72	0.281365	6.4	0.1
TZ-2-15	2588	0.040108	0.000954	0.281359	0.00002	6.5	2.64	2.67	0.281312	6.0	0.1
TZ-2-16	130	0.023523	0.000670	0.282723	0.00002	1.1	0.74	1.12	0.282722	8.6	0.1
TZ-2-17	2571	0.036435	0.000922	0.281381	0.00002	6.9	2.61	2.63	0.281335	6.4	0.1
TZ-2-18	132	0.010399	0.000272	0.282716	0.00002	0.9	0.75	1.13	0.282716	8.1	0.1
TZ-2-19	130	0.023216	0.000627	0.282729	0.00002	1.3	0.74	1.10	0.282727	9.0	0.1
TZ-2-20	132	0.024706	0.000670	0.282584	0.00002	-3.8	0.94	1.43	0.282582	7.0	0.1
TZ-2-21	130	0.011683	0.000327	0.282597	0.00002	-3.4	0.91	1.40	0.282596	8.8	0.1
TZ-2-22	129	0.012666	0.000390	0.282648	0.00002	-1.6	0.84	1.28	0.282647	8.3	0.1
TZ-2-23	2551	0.019183	0.000498	0.281433	0.00002	9.1	2.51	2.49	0.281409	6.6	0.2
TZ-2-24	133	0.013941	0.000427	0.282673	0.00003	-0.6	0.81	1.22	0.282672	8.8	0.1
TZ-2-25	133	0.018658	0.000487	0.282631	0.00003	-2.1	0.87	1.32	0.282630	8.7	0.1
TZ-2-26	133	0.011887	0.000317	0.282648	0.00002	-1.5	0.84	1.28	0.282647	9.0	0.1
TZ-2-27	131	0.013463	0.000365	0.282393	0.00002	-10.	1.20	1.85	0.282393	6.9	0.2
TZ-2-28	2574	0.007186	0.000172	0.281273	0.00003	4.5	2.70	2.79	0.281265	6.7	0.1
TZ-2-29	2494	0.020767	0.000574	0.281367	0.00002	5.3	2.61	2.67	0.281340	6.1	0.1
Gabbro (TZ-13)											
TZ-13-0	128	0.010804	0.000295	0.282596	0.00002	-3.4	0.91	1.40	0.282596		
TZ-13-0	2459	0.005496	0.000128	0.281321	0.00002	3.6	2.64	2.75	0.281315		
TZ-13-0	132	0.013902	0.000383	0.282713	0.00002	0.8	0.75	1.14	0.282712	7.7	0.1
TZ-13-0	2465	0.008461	0.000194	0.281326	0.00002	3.8	2.64	2.74	0.281317	6.0	0.2
TZ-13-0	130	0.017811	0.000438	0.282679	0.00002	-0.5	0.80	1.21	0.282678	7.8	0.1
TZ-13-0	128	0.008895	0.000275	0.282605	0.00003	-3.1	0.90	1.38	0.282605	8.8	0.1
TZ-13-0	2465	0.014853	0.000355	0.281385	0.00002	5.6	2.57	2.63	0.281368	6.9	0.1
TZ-13-0	132	0.031279	0.000748	0.282785	0.00002	3.3	0.66	0.98	0.282783	8.4	0.1
TZ-13-1	2467	0.031596	0.000745	0.281342	0.00002	3.5	2.65	2.76	0.281307	5.7	0.1
TZ-13-1	2493	0.025654	0.000605	0.281321	0.00002	3.6	2.67	2.78	0.281292		
TZ-13-1	2500	0.016145	0.000366	0.281257	0.00002	1.9	2.74	2.89	0.281240		
TZ-13-1	125	0.010689	0.000267	0.282587	0.00002	-3.8	0.93	1.42	0.282587	8.5	0.1
TZ-13-1	129	0.015316	0.000409	0.282584	0.00002	-3.9	0.93	1.43	0.282583	8.6	0.1
TZ-13-1	132	0.027935	0.000645	0.282705	0.00002	0.5	0.77	1.16	0.282703	8.4	0.1
TZ-13-1	2432	0.009599	0.000260	0.281229	0.00002	-0.5	2.77	2.98	0.281217	5.8	0.1
TZ-13-1	2465	0.013642	0.000406	0.281222	0.00002	-0.2	2.79	2.99	0.281203		

TZ-13-2	154	0.010197	0.000250	0.282569	0.00002	-3.8	0.95	1.44	0.282568		
TZ-13-2	128	0.008262	0.000226	0.282548	0.00002	-5.1	0.98	1.51	0.282548	9.1	0.1
TZ-13-2	2478	0.012329	0.000293	0.281293	0.00002	2.7	2.69	2.82	0.281279		
TZ-13-2	2479	0.018107	0.000432	0.281328	0.00003	3.8	2.65	2.75	0.281308		
TZ-13-2	124	0.031775	0.000785	0.282703	0.00002	0.2	0.78	1.17	0.282701	8.0	0.1
TZ-13-2	2431	0.014250	0.000341	0.281285	0.00002	1.3	2.70	2.87	0.281269		
TZ-13-2	125	0.014032	0.000414	0.282597	0.00003	-3.5	0.92	1.40	0.282596		
TZ-13-2	128	0.128416	0.003245	0.282404	0.00004	-10.	1.28	1.84	0.282396		
TZ-13-3	2376	0.026275	0.000755	0.281372	0.00003	2.5	2.61	2.75	0.281338		

The following parameters were applied to the calculation: $(^{176}\text{Lu}/^{177}\text{Hf})_{\text{CHUR}} = 0.0332$, $(^{176}\text{Hf}/^{177}\text{Hf})_{\text{CHUR},0} = 0.282772$, $(^{176}\text{Lu}/^{177}\text{Hf})_{\text{DM}} = 0.0384$, $(^{176}\text{Hf}/^{177}\text{Hf})_{\text{DM},0} = 0.28325$ (Blichert-Toft and Albarède, 1997; Griffin et al., 2000), ^{176}Lu decay constant $\lambda = 1.867 \times 10^{-11} \text{ a}^{-1}$ (Söderlund et al., 2004).

Highlights

- Tiezhai mafic rocks were derived from modified Luxi EM1-asthenospheric mantle
- Tiezhai syenite porphyry was derived from newly underplated thick lower crust
- Addition of fluids were important mechanism for NCC destruction
- NCC destruction may be related to rollback of the subducted paleo-Pacific slab

130-125 Ma

Graphics Abstract

Figure 1

Figure 2

Tieshai syenite porphyry (TZ-2)

Tieshai syenite porphyry (TZ-5)

Tieshai gabbro (TZ-13)

Tieshai monzonite (TZ-16)

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12

Figure 13

Figure 14

Figure 15

(a) ~ 180 Ma

(b) 140-130 Ma

(c) 130-125 Ma

(d) 125-120 Ma

Figure 16