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Composition and geochronology of the deep-seated xenoliths from the southeastern margin of the North China Craton.

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1 1 **Composition and geochronology of the deep-seated xenoliths**
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4 2 **in the southeastern margin of the North China Craton, and**
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6 3 **geological implications**

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ABSTRACT

In the southeastern margin of the North China Craton, a variety of deep-seated xenoliths, which have various mineral assemblages depending on their formation times, bulk compositions and metamorphic histories, occur with different sizes from 2 cm to 20 cm in diameter in Mesozoic dioritic porphyry. In this study, we present a combined petrologic, geochronological, Hf isotope and geochemical study on different types of xenoliths sampled from the same locality (Jiagou) and use these data to better constrain the composition and age of the deep crust beneath the area, thus shading new light on its late-Archean to Palaeoproterozoic crustal growth and evolution. Most of the xenoliths are mafic meta-igneous rocks, among which garnet-bearing lithologies are common. They can be classified into three broad petrographic groups: spinel-bearing garnet clinopyroxenite / phlogopite clinopyroxenite / spinel pyroxenite (Group 1), garnet amphibolite or hornblendite / garnet granulite / basic gneiss lacking pyroxene (Group 2), and garnet-bearing felsic (intermediate-acid) gneiss (Group 3). Generally, mafic-ultramafic rocks predominate whereas felsic rocks are rare. The protoliths of the studied xenoliths are igneous and range from basalt through andesite to dacite. Geochemical and Hf-isotope data indicate that most xenoliths for Groups 2 and 3 resemble those formed at a convergent continental margin arc setting, and a few (most of Group 1) represent mantle-derived products. Multiple metasomatic imprints have been recognized by multistage metamorphic mineral assemblages (e.g., different generations of amphibole) and ages. The occurrence of hydrous minerals such as multiple hornblendes and phlogopites in the xenoliths is the most important witness of a water-rich precursor and multistage metasomatism triggered by subduction-related or mantle-derived fluids or melts.

SHRIMP zircon U–Pb dating, Hf isotope and element geochemical data provide the solid evidence for the existence of subduction-related adakite-like and arc-related rocks in the southeastern margin of the North China Craton at ca. 2.5 Ga and 2.1 Ga, and confirm the occurrence of high-pressure granulite-facies metamorphism at ca. 1.8 Ga. These data suggest an episodic growth of Precambrian lower crust beneath the

1 region in response to two stages of subduction–accretion and one vertical accretion of
2 mantle-derived basaltic magma at the base of the lower crust. Additionally, a
3 previously unknown late mantle-derived basaltic magmatism at 393 ± 7 Ma has been
4 recognized. Thus, the data presented in this paper demonstrate that the deep crust
5 beneath the southeastern margin of the North China Craton is composed of a
6 combination of different protoliths with a variety of Paleozoic, Palaeoproterozoic and
7 late Neoproterozoic sources, rather than rock types with a single origin as previously
8 supposed, and is hybrid in composition. Therefore, the studied xenoliths have
9 different origins, formed at different ages and originated at different levels of the deep
10 crust.

11
12 *Keywords:* U-Pb dating; Hf isotope; lower crust; crustal growth; deep-seated xenoliths

1. Introduction

The formation of continental crust has been ascribed to two distinct plate tectonic settings: active continental margin and intra-plate (e.g., Rudnick, 1995). Growth of continents in within-plate settings occurs in response to plume-related magmatism, whereas the continental crust currently grows at the plate boundaries through magmatic additions and island arc collisions. Furthermore, convergent continental margins are well assessed to be a major site of lower crustal accretion through different mechanisms including basaltic underplating at the base of the crust and subduction–accretion (e.g., Weber et al., 2002, and references therein). However, relatively fewer examples of lower crustal xenoliths coming from active continental margins have been reported (e.g., Kempton et al., 1997; Weber et al., 2002). Crucial to all models of continental growth is knowledge of the petrology and geochemistry of the lower crust (Condie, 1997, 1999; Weber et al., 2002). However, our understanding of the Precambrian crust formation, especially lower continental crust, is generally hampered by the scarcity of data concerning deep crustal processes. At present, information on the lower continental crust is basically derived from studies on xenoliths in volcanic rocks, observations on exhumed granulite terrains and geophysical data, each one having its own advantages and limitations. Few, if any, exposed metamorphic terrains retain evidence of having formed near the base of the crust, and many have been modified by deformation and re-crystallization during uplift. Geophysical data are all remotely sensed and thus their interpretation without geological constraints can be speculative. As a consequence, xenoliths brought to the surface mainly by basaltic or dioritic volcanics were thought to provide direct windows into the present and/or fossil lower crust (e.g., Rudnick et al., 1986; Loock et al., 1990; Moser and Heaman, 1997; Upton et al., 2001; Weber et al., 2002; Zheng et al., 2003; Gao et al., 2004; Downes et al., 2007). In particular, mafic and ultramafic xenoliths are generally considered to represent current samples of inaccessible upper mantle and lower crustal lithologies (Montanini and Harlov, 2006).

Zircon retains a high-fidelity record of its crystallization history due to its

1 sluggish kinetic properties for U, Th and Pb (Cherniak and Watson, 2003) and is able
2 to preserve multiple stages of magmatic and metamorphic records in high-grade
3 metamorphic rocks involved in complex processes (e.g., Ayers et al., 2002; Möller et
4 al., 2002; Hermann and Rubatto, 2003; Liu et al., 2011b), thereby readily amenable to
5 be used for obtaining U–Pb ages in poly-metamorphic rocks (Liu et al., 2009, and
6 references therein). Many studies have shown that U–Pb zircon geochronology is
7 ideally suited for elucidating the temporal evolution of lower crustal xenoliths (e.g.,
8 Hanchar and Rudnick, 1995; Moser and Heaman, 1997). In addition, the Lu–Hf
9 isotope system of zircon is generally more resistant to alteration processes than the
10 U–Pb isotope system (Amelin et al., 2000; Cherniak and Watson, 2003; Zheng et al.,
11 2006; Gerdes and Zeh, 2009). As a consequence, zircon preserves a high-quality
12 record of near-initial Hf-isotope ratios, which can be used both in provenance studies
13 and as a petrogenetic indicator (e.g., Griffin et al., 2004; Woodhead et al., 2004). The
14 combined U–Pb and Lu–Hf isotope data for single zircon grains have been proven to
15 provide reliable and very detailed information about the timing of magmatic and
16 metamorphic events, as well as about the petrogenesis and the evolution of crust and
17 mantle over time (e.g., Vervoort and Blichert-Toft, 1999; Amelin et al., 2000; Griffin
18 et al., 2000, 2004; Andersen et al., 2002; Condie et al., 2005; Iizuka et al., 2005;
19 Hawkesworth and Kemp, 2006; Wu et al., 2006; Zheng et al., 2006; Gerdes and Zeh,
20 2009; Kemp et al., 2009; Zeh et al., 2009).

21 The North China Craton (NCC) is one of the oldest cratons in the world and there
22 are numerous U–Pb zircon geochronological data on its Precambrian metamorphosed
23 basement or lower crust. The data show that except for minor >3.6 Ga components
24 (Liu et al., 1992; Zheng et al., 2004a; Wu et al., 2008; Zhang et al., 2012), rocks of the
25 NCC basement have U–Pb zircon ages clustering at 1.8–1.9 and ~2.5 Ga (e.g., Zhao
26 et al., 2001; Gao et al., 2004; Zheng et al., 2004b; Santosh et al., 2007b; Liu et al.,
27 2009, 2011a; Zhai and Santosh, 2011). The 2.5 Ga peak was considered to coincide
28 with the major crustal growth period of the NCC (Gao et al., 2004). In addition, the
29 basement rocks show a large range of Nd and Hf model ages with a peak at ~ 2.7 Ga,
30 which is considered to be a major crustal growth period in the NCC (Jiang et al.,

1 2010). Based on Nd model ages, Wu et al. (2005) suggested that 2.8 Ga is the best
2 estimate of the major mantle extraction age for the NCC and Zhai et al. (2005)
3 considered that ca. 2.9–2.7 Ga is the main crust-forming episode in the NCC and that
4 ~2.5 Ga age peak reflects a high-grade metamorphic event. Whether the most
5 prominent ~2.5 Ga zircon age-peak or the ~2.8 Ga Nd model age-peak or ~2.7 Ga
6 zircon Hf-model age-peak registers the major crustal growth period of the NCC is still
7 uncertain.

8 However, in the southeastern margin of the NCC, there are numerous Mesozoic
9 granitic or dioritic intrusions in the Xuzhou-Suzhou region and voluminous Mesozoic
10 and Cenozoic volcanic rocks in the northeastern Anhui province and western
11 Shandong Province which yield xenoliths of deep-seated rock types (Zhou et al., 1992;
12 Xu et al., 1998, 2002, 2006, 2009; Zhang et al., 2002; Yu et al., 2003; Zheng et al.,
13 2003, 2009; Huang et al., 2004; Zhang, 2005; Guo and Li, 2009; Liu et al., 2009). In
14 this study, the target region is focused on the Xuzhou-Suzhou area and the object is
15 the deep-seated xenoliths entrained in the Mesozoic Jiagou intrusion near Suzhou.
16 Although the xenoliths in the studied area have previously been described and are
17 considered to be formed at ~2.5 Ga and suffered from 1.8–1.9 Ga granulite-facies
18 overprinting (Xu et al., 2002, 2006, 2009; Guo and Li, 2009; Liu et al., 2009),
19 integrated petrographic, geochronological and geochemical studies on various
20 xenoliths from the same locality are scarce. As a consequence, the composition and
21 age of the deep crust beneath this portion of the NCC are not well-constrained;
22 nevertheless, this information is essential for a robust understanding of the
23 Precambrian crustal evolution and tectonic history of the area. Moreover, the nature
24 and petrogenesis of the xenoliths also bear on the question as to whether or not they
25 shared a common formational (~2.5 Ga arc origin) and evolutionary history as
26 proposed by Xu et al. (2009). In addition, so far, **convincing evidence of a Paleozoic
27 tectonothermal event has not been reported in the area except for the early Paleozoic
28 (480 Ma) diamondiferous kimberlites from the intermediate segment of the Eastern
29 Block of the NCC (Li et al., 2011 and references therein).**

30 In this study, we present a combined petrologic, geochronological, Hf isotope and

1 geochemical study on different types of xenoliths from Jiagou in the southeastern
2 margin of the NCC and use these data to better constrain the composition and age of
3 the deep crust beneath the area, thus shading new light on its late Archean to
4 Palaeoproterozoic crustal accretion and evolution. In particular, this study suggests a
5 new interpretation of the composition, nature and age of the deep crust in the studied
6 region, providing solid documentation for: (i) multiple crustal growth, (ii) its
7 modification during the subsequent tectono-metamorphic events, and (iii) its
8 interaction with the ascending mantle-derived magma. In addition, a Paleozoic
9 mantle-derived magmatic event, leading to metasomatism on rocks of the deep crust
10 beneath the region, has been recognized for the first time.

11 12 **2. Geological setting**

13 The NCC (Fig. 1) is one of the largest and oldest cratonic blocks in the world, as
14 evidenced by the presence of > 3.6 Ga crustal remnants **exposed at the surface or in**
15 **lower crustal xenoliths (e.g., Liu et al., 1992; Zheng et al., 2004a; Wu et al., 2008;**
16 **Zhang et al., 2012)**. The NCC underwent a series of tectonothermal events in late
17 Archean and Palaeoproterozoic (e.g., Zhai et al., 2000; Zhai and Liu, 2003; Zhao et al.,
18 2000, 2001; Wilde et al., 2002; Kusky and Li, 2003; Guo et al., 2005; Kröner et al.,
19 2005; Hou et al., 2006, 2008; Tang et al., 2007; Guo and Li, 2009; Liu et al., 2009,
20 2011a; Jiang et al., 2010; Zhai and Santosh, 2011), and was stabilized in the late
21 Palaeoproterozoic (e.g., Zhai et al., 2000) but experienced widespread tectonothermal
22 reactivation during Mesozoic and Cenozoic (Zhou et al., 1992; Xu et al., 1998; Yu et
23 al., 2003; Zheng et al., 2003, 2009; Huang et al., 2004; Xu et al., 2006; Liu et al.,
24 2004, 2012; Zhang, 2012; Zhang et al., 2012). The NCC is bounded by faults and
25 younger orogenic belts. The early Paleozoic Qilianshan orogen and the late Paleozoic
26 Tianshan–Inner Mongolia–Daxinganling orogen bound the craton to the west and the
27 north, respectively, whereas in the south the Mesozoic Qinling–Dabie–Sulu high- to
28 ultrahigh-pressure (HP to UHP) belt separates the NCC from the Yangtze Craton. The
29 NCC can be subdivided into the Eastern Block, the Western Block and the intervening

1 Trans-North China Orogen or Central Orogenic Zone (e.g., Zhao et al., 2000, 2001;
2 Kusky and Li, 2003; Zhai and Liu, 2003; Kröner et al., 2005).

3 The Bengbu and Xuzhou–Suzhou areas lie in the Eastern Block along the
4 southeastern margin of the NCC, about 100 km west of the Tan-Lu fault zone on the
5 southwestern termination of the Sulu orogen and about 300 km north of the Dabie
6 orogen and Hefei Basin (Fig. 1). The deformed Neoproterozoic to late Paleozoic
7 sedimentary cover and late Archean to Palaeoproterozoic metamorphic basement in
8 the region were intruded by several small Mesozoic intrusions (e.g. Jiagou, Banjing
9 and Ligu) (Fig. 1) mainly composed of dioritic to monzodioritic porphyry. The
10 Precambrian metamorphic basement in the studied area is mainly exposed at the
11 surface in Bengbu, while no exposed basement occurs in the Xuzhou-Suzhou area
12 where abundant deep-sourced enclaves occur in the Mesozoic intrusions (Xu et al.,
13 2002, 2006, 2009; Guo and Li, 2009; Liu et al., 2009).

14 The studied samples were collected at the southeastern margin of the NCC from
15 the xenoliths, which occur with different sizes from 2 cm to 20 cm in diameter in the
16 Mesozoic dioritic porphyry at Jiagou near Suzhou (Figs. 1 and 2). While a lower
17 crustal origin is evident for mafic garnet-bearing granulite or amphibolite and gneiss
18 xenoliths (Huang et al., 2004; Liu et al., 2004, 2009), this is less evident for
19 spinel-bearing garnet clinopyroxenite (see below). Moreover, it is not clear whether or
20 not all the xenoliths of various lithologies share a common origin and formed in the
21 same period. In order to better constrain the composition and age of the deep crust in
22 this portion of the NCC, seventeen representative samples or xenoliths collected from
23 the same locality at Jiagou were investigated through petrology and whole-rock
24 geochemistry and five of them were also studied by *in situ* SHRIMP U-Pb dating,
25 including two samples previously reported by Liu et al. (2009). At the same time, all
26 the five dated xenoliths from Jiagou and a garnet amphibolite from the exposed
27 Precambrian metamorphic basement in the Bengbu region dated by Liu et al. (2009)
28 have been studied by means of zircon Hf-isotope analyses with the aim to better
29 constrain the formation and evolution of lower crust in the region.

3. Petrography and samples

The xenoliths in the studied area are metamorphic rocks of deep-seated origin and comprise abundant mafic rocks, less common feldspar-free ultramafic types and rare felsic gneisses. No evidence was found of the eclogite xenoliths reported by Xu et al. (2002) in the same area. From a lithological viewpoint, the xenoliths include spinel-bearing garnet clinopyroxenite, phlogopite clinopyroxenite, spinel pyroxenite (websterite), garnet granulite, garnet amphibolite, garnet-bearing hornblendite and garnet-bearing mafic and felsic gneiss. Details of the petrography and mineral chemistry of the studied xenoliths were given in a separate paper (Liu et al., 2009) and are only summarized here. They can be classified into three broad petrographical groups: ultramafic spinel-bearing garnet clinopyroxenite / phlogopite clinopyroxenite / spinel pyroxenite (Group 1), mafic garnet amphibolite or hornblendite / garnet granulite / **basic gneiss lacking pyroxene** (Group 2) and felsic garnet-bearing gneiss (Group 3). Representative mineral compositions are reported in Table 1. Mineral abbreviations in figures and tables are after Whitney and Evans (2010).

Different mineral assemblages and compositions in the xenoliths reflect either different bulk-rock compositions or different metamorphic evolution. Group 1 includes spinel-bearing garnet clinopyroxenite and phlogopite clinopyroxenite (Fig. 3a–c). Spinel-bearing garnet clinopyroxenite is composed mainly of garnet, clinopyroxene and spinel with minor hornblende and biotite or phlogopite. Garnet occurs as exsolution lamellae in clinopyroxene, coronae around spinel and in the matrix. The occurrence of garnet exsolution is common in aluminous clinopyroxene from the mantle xenoliths of eclogite / garnet clinopyroxenite / pyroxenite found in kimberlite and alkali basalt (e.g., Sobolev and Sobolev, 1964; Lappin and Dawson, 1975). The development of this exsolution has been interpreted as the result of a slow isobaric cooling at upper mantle depths from high-temperature (e.g., Aoki et al., 1980; Sautter and Harte, 1988; Jerde et al., 1993). In addition, the occurrence of garnet coronae around spinel (Fig. 2a) suggests that Group 1 xenoliths formed under spinel-garnet-facies conditions. *P–T* conditions of spinel–garnet phase transition were

1 experimentally documented at 1.6–2 GPa and 900–1100 °C (e.g. O’Neill, 1981; Fan
2 et al., 1997, and references therein). Therefore, such a spinel-bearing garnet
3 clinopyroxenite was most likely of mantle origin.

4 In Group 2, the typical mineral assemblage is garnet, clinopyroxene, plagioclase,
5 hornblende, quartz, ilmenite and rutile (Liu et al., 2009) (Fig. 3d, e). Interestingly
6 enough, mafic granulites and garnet amphibolites or hornblendites coexist and have
7 similar mineral assemblages and metamorphic evolution histories (Liu et al., 2009),
8 most likely due to their different whole-rock compositions, local differences in fluid
9 activity and various degrees of equilibration. Group 3 comprises plagioclase,
10 hornblende, quartz, ilmenite and minor garnet and rutile (Fig. 3f). As a whole, Groups
11 2 and 3 xenoliths show a consistent peak metamorphic assemblage consisting of
12 garnet, clinopyroxene, rutile, plagioclase, quartz, Ti-rich hornblende and Ti-rich
13 biotite, indicative of HP granulite-facies metamorphism (Yardley, 1989; Pattison,
14 2003; O’Brien and Rötzler, 2003; Liu et al., 2009). Microstructural observations and
15 mineral relationships suggest that this peak assemblage was followed by: (i)
16 plagioclase + green hornblende + ilmenite + titanite; (ii) chlorite + calcite + magnetite.
17 These assemblages are representative of amphibolite- and greenschist-facies
18 metamorphic stages, respectively. Thermobarometric data and mineral assemblage
19 suggested *P–T* conditions of 1.0–1.2 GPa and 800–860 °C for the HP granulite-facies
20 metamorphic stage at *c.* 1.8 Ga (Liu et al., 2009).

21 All the xenoliths have undergone some retrogression promoted by hydration, with
22 formation of mica (phlogopite and/or biotite), amphibole and chlorite but lacking any
23 significant evidence of eclogite-facies overprint. The most abundant mineral is
24 amphibole, followed by clinopyroxene and plagioclase. Garnet, rutile, plagioclase,
25 ilmenite, quartz, apatite, spinel and corundum have also been identified; no olivine
26 has been observed. The pervasive occurrence of amphibole and biotite or phlogopite
27 implies a water-rich precursor and/or late modification by fluid and/or mantle
28 metasomatism (see below). Furthermore, in Groups 2 and 3 xenoliths there are
29 multiple generations of amphibole such as brown pargasite to ferro-pargasite and
30 green magnesio-hastingsite or edenite as reported by Liu et al. (2009). Brown and

1 green hornblendes contain higher and lower amounts of TiO₂, respectively; they are
2 inferred to have formed under different conditions, because it has been demonstrated
3 that Ti increases with metamorphic grade (Raase, 1974; Pattison, 2003). This multiple
4 stage of amphibole growth is also supported by textural evidence, with green
5 hornblende occurring in retrograde symplectites or rimming the clinopyroxene margin
6 and brown hornblende occurring as inclusions. It is thus evident that the xenoliths
7 suffered multiple episodes of metamorphism and metasomatism.

8 K-feldspar is commonly associated with plagioclase or occurs along fractures
9 within clinopyroxene (Fig. 3e) or garnet, suggesting a possible metasomatic origin
10 due to solid-state infiltration of K-rich fluids/melts. These lines of evidence suggest
11 that formation of K-feldspar in the mafic xenoliths reflects metasomatic processes,
12 requiring an external K-rich fluid source, which operated in the lower crust during and
13 after *in situ* high-T recrystallization of relatively dry rocks (Montanini and Harlov,
14 2006).

15 16 **4. Analytical methods**

17 *4.1. Whole-rock major and trace elements*

18 Whole-rock major element analysis was performed by wet chemical methods at
19 the Langfang Laboratory, Hebei Bureau of Geology and Mineral Resources; trace
20 element compositions were measured by an Elan DRCII ICP-MS at the CAS Key
21 Laboratory of Crust-Mantle Materials and Environments, University of Science and
22 Technology of China in Hefei. Analytical uncertainties range from ± 1 to $\pm 5\%$ for
23 major elements and $\pm 5\%$ to $\pm 10\%$ for trace elements. Detailed analytical procedures
24 and instrument parameters for trace element analyses are reported in Hou and Wang
25 (2007).

26 27 *4.2. Zircon U-Pb dating*

28 Zircons were separated from the whole-rock samples by crushing and sieving,
29 followed by conventional magnetic and heavy-liquid separation and hand-picking

1 under binoculars. Approximately 200 zircon grains for samples 07JG32 and 07JG34
2 and 50 grains for sample 07JG30 were mounted using epoxy, together with a zircon
3 U–Pb standard TEM (417 Ma) (Black et al., 2003). The mount was then polished until
4 all zircon grains were approximately cut in half. The internal zoning patterns of the
5 crystals were observed by cathodoluminescence (CL) imaging at Beijing SHRIMP
6 Center, Chinese Academy of Geological Sciences (CAGS).

7 Zircons were dated on a SHRIMP II at Beijing SHRIMP Center. Uncertainties in
8 ages are quoted at the 95% confidence level (2σ). A spot size of about $30\mu\text{m}$ was used.
9 Common Pb corrections were made using measured ^{204}Pb . The SHRIMP analyses
10 followed the procedures described by Williams (1998). Both optical
11 photomicrographs and CL images were taken as a guide to select the U–Pb dating
12 spots. Five scans through the mass stations were made for each age determination.
13 Standards used were SL13, with an age of 572 Ma and U content of 238 ppm, and
14 TEM, with an age of 417 Ma (Williams, 1998; Black et al., 2003). The U–Pb isotope
15 data were treated following Compston et al. (1992) with the ISOPLOT program of
16 Ludwig (2001). Measurement of standard zircon TEMORA 1 yielded a weighted
17 $^{206}\text{Pb}/^{238}\text{U}$ age of 416.3 ± 3.2 Ma, which is in good agreement with the recommended
18 isotope dilution-thermal ionization mass spectrometry (ID-TIMS) age of $416.75 \pm$
19 0.24 Ma (Black et al., 2003).

20 Mineral inclusions in zircon were identified using Raman spectroscopy at the
21 Continental Dynamics Laboratory, CAGS and/or substantiated using the electron
22 microprobe analyzer (EMPA) at the Institute of Mineral Resources, CAGS in China
23 and at the Department of Mineralogical and Petrological Sciences, University of
24 Torino in Italy. The analytical conditions on the Raman and EMPA were reported by
25 Liu et al. (2009).

27 *4.3. Zircon Lu-Hf isotope analysis*

28 The laser ablation multi-collector inductively coupled plasma mass spectrometry
29 (LA-MC-ICPMS) Lu–Hf isotope analysis was carried out at the Institute of Geology
30 and Geophysics, the Chinese Academy of Sciences in Beijing. Instrumental conditions

1 and data acquisition followed those described by Xu et al. (2004) and Wu et al. (2006).
2 A Geolas-193 laser-ablation microprobe was attached to a Neptune multi-collector
3 ICPMS. Typical ablation times were 30 to 90 s with a 10 Hz repetition rate and laser
4 power of 100 mJ/pulse, resulting in a beam depth of 30 to 50 μm . A stationary spot
5 was used for core analyses with a beam diameter of about 63 μm . In contrast, a linear
6 ablation was employed for rim and mantle analyses with a beam width of about 31.5
7 μm . Both He and Ar carrier gases were used to transport the ablated sample from the
8 laser-ablation cell via a mixing chamber to the ICPMS torch. A signal collection
9 model for one block with 200 cycles, in which one cycle has 0.131 s integration time
10 and bulk time for one measurement lasts about 30 s. During analytical processes, the
11 $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of two standard zircons Mud Tank and GJ-1 were 0.282502 ± 5 and
12 0.282008 ± 7 , respectively, which are in good agreement with the referenced
13 $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of 0.282507 ± 6 and 0.281999 ± 8 measured by using the solution
14 method (Woodhead and Hergt, 2005; Wu et al., 2006). The main parameters used in
15 the $\epsilon\text{Hf}(t)$ (t = crystallization time of zircon) and model age calculations are listed as
16 follows: $(^{176}\text{Lu}/^{177}\text{Hf})_{\text{CHUR}}=0.0332$ and $(^{176}\text{Hf}/^{177}\text{Hf})_{\text{CHUR},0}=0.282772$ for chondritic
17 uniform reservoir (Blichert-Toft & Albarede, 1997), $(^{176}\text{Lu}/^{177}\text{Hf})_{\text{DM}}=0.0384$ and
18 $(^{176}\text{Hf}/^{177}\text{Hf})_{\text{DM}}=0.28325$ for depleted mantle reservoir (Griffin et al., 2000). ^{176}Lu
19 decay constant $\lambda = 1.865 \times 10^{-5}$ m.y was used for calculations (Soderlund et al.,
20 2004).

21 22 **5. Results**

23 *5.1. Whole-rock major and trace elements*

24 Seventeen samples of xenoliths, representative of the three different groups
25 mentioned above, have been analyzed in this study and show a broad range in major
26 element composition (Table 2; Figs. 4 and 5). Based on the discriminant function
27 (D.F.) proposed by Shaw (1972) and their high MgO contents, all the protoliths of the
28 studied xenoliths are magmatic, in agreement with the igneous cores of their zircons
29 in CL images (see below). The SiO_2 contents range between 37.92 and 67.06 wt.%,

1 spanning from mafic-ultramafic, minor intermediate and rare acid compositions with
2 variable Mg-numbers (77–38) (Fig. 5). The studied xenoliths mostly derived from
3 basaltic protoliths, ranging in composition from alkaline to subalkaline types, and a
4 few from andesitic or dacitic protoliths (Fig. 4). The samples of the Group 1 may be
5 basalts or basaltic rocks. For the Group 2, their protoliths are mostly basalts and
6 basaltic andesites; among this group, samples 08JG08, 07JG34 and 07JG05-2 have
7 higher Nb contents of 8.17-42.3 ppm (Table 2 and Fig. 5b), suggesting that their
8 precursors were high-Nb (HNB) and Nb-enriched (NEB) basalts, respectively
9 according to Sajona et al. (1996) and Martin et al. (2005). Among group 3, sample
10 07JG32 displays magnesian andesite characteristics with SiO₂ 53.58 wt.% and MgO
11 5.48 wt.% (Calmus et al., 2003) whereas sample 08JG03 has the highest SiO₂ content
12 (67.06 wt.%) and MgO 5.18 wt.%, suggestive of a dacite precursor (Fig. 4).

13 The REE patterns for the three types of xenoliths are also distinctive (Fig. 6). In
14 Group 1, most samples have relatively flat REE patterns and a few show a weak
15 LREE enrichment, generally associated with negative Eu anomalies. For Group 2, the
16 samples display negative, positive or even no Eu anomalies and most of them show
17 near-flat REE to slightly enriched LREE patterns, except for sample 07JG07 which
18 exhibits a pronounced HREE enrichment which may be related to its higher modal
19 amount of garnet. Within the investigated xenoliths, amphibole-rich ones (e.g., sample
20 07JG05-2) are more enriched in total REEs than the other xenoliths. In contrast, the
21 two samples of Group 3 are characterized by the lack of Eu anomalies and by strongly
22 fractionated LREE to HREE trends (Fig. 6).

23 Most of the Groups 2 and 3 xenoliths exhibit primitive-mantle-normalized
24 negative Ta–Nb anomalies typical of an arc signature (Fig. 7b and c), this signature
25 being also supported by Sr/Y–Y diagram in Fig. 8. Positive Ti anomalies, consistent
26 with the occurrence of Fe–Ti oxides, are present in some of the Group 2 and mirrored
27 by negative Ti anomalies in the Group 3 xenoliths.

28 Cr and Ni contents for the studied samples display a large variation between
29 4.5–448 ppm and 18.7–248 ppm, most probably resulted from their different origin or
30 from the processes they suffered (e.g., metasomatism, see below); the highest values

1 (e.g. samples 07JG06, 07JG12 and 07JG21) are typical of primitive to slightly
2 fractionated mantle-derived magmas (Embey-Isztin et al., 2003). This chance is also
3 supported by the relatively high Mg-number of some samples, i.e. 54–77. In addition,
4 the samples exhibit various Sr, Ba, Nb and Y contents, some of them having
5 comparatively high ones (Table 2; Figs. 8 and 9).

7 5.2. SHRIMP zircon U–Pb dating

8 5.2.1. CL imaging and mineral inclusions in zircon

9 SHRIMP U-Pb dating, combined with CL information and mineral inclusions for
10 the selection of specific locations in zircon, is one of the most powerful techniques to
11 unravel complex histories of individual zircon grains occurring in multistage
12 metamorphic rocks (e.g., Vavra et al., 1996; Gebauer et al., 1997; Hermann et al.,
13 2001; Liu et al., 2009, 2011b). Although mineral inclusions in zircon, especially from
14 lower crustal rocks or deep-seated xenoliths, are rare, they can be occasionally found
15 and provide a direct link between zircon formation and metamorphism (e.g., Gebauer
16 et al., 1997; Hermann et al., 2001; Liu et al., 2007, 2009, 2011b).

17 Sample 07JG30 (Group 1) contains rare zircon grains, most probably resulted
18 from a mantle genesis. Except for three inherited zircons (Fig. 10c), others have
19 similar CL images with typical oscillatory zoning, indicative of igneous origin (e.g.,
20 Hanchar and Rudnick, 1995; Gebauer et al., 1997). Some of them show a core-rim
21 structure, but the rims are usually very thin ($< 20 \mu\text{m}$) and do not contain mineral
22 inclusions (Fig. 10b).

23 On the basis of CL images and mineral inclusions, core-rim domains have been
24 clearly recognized in zircon from samples 07JG34 (Group 2) and 07JG32 (Group 3)
25 (Fig. 10d–p). Most of the zircons cores exhibit oscillatory growth zoning, which is
26 typical of igneous origin. However, some cores are truncated, embayed or irregularly
27 shaped (Fig. 10k, n), suggesting that they were partially or completely reabsorbed,
28 probably in the presence of a fluid (e.g., Ayers et al., 2002). The rim domains contain
29 garnet, clinopyroxene, plagioclase, quartz and rutile inclusions (Table 3; Fig. 10h, k, l,
30 o, p), similar to those previously described by Liu et al. (2009), indicative of a typical

1 HP granulite-facies assemblage (e.g., O'Brien and Rötzler, 2003; Pattison, 2003).

3 5.2.2. U–Pb ages

4 Twelve U-Pb spot analyses were made on 11 zircon grains from sample 07JG30
5 (Group 1: Table 4 and Fig. 11a), including 3 inherited and 8 igneous zircon dots.
6 Except for 2 grains, other 6 analyses of igneous zircons record $^{206}\text{Pb}/^{238}\text{U}$ concordant
7 ages ranging from 384 to 410 Ma with a weighted mean age of 393 ± 7 Ma (MSWD =
8 0.83). This age is in good agreement with the age of 387 ± 13 Ma obtained from a
9 metamorphic overgrowth rim of zircon previously dated from sample 07JG14 (Liu et
10 al., 2009) within error, recording a thermal event induced by intrusion of concurrent
11 mantle-derived magma.

12 Twenty-three U-Pb spot analyses were made on 22 zircon grains from sample
13 07JG34 (Group 2: Table 4 and Fig. 11b). 6 analyses of inherited zircons record
14 $^{207}\text{Pb}/^{206}\text{Pb}$ concordant ages ranging from 2559 to 2696 Ma with a weighted mean age
15 of 2631 ± 43 Ma (MSWD = 0.72). Except for spot 20.1, 8 analyses of igneous zircons
16 record $^{207}\text{Pb}/^{206}\text{Pb}$ concordant ages ranging from 2389 to 2567 Ma with a weighted
17 mean age of 2480 ± 49 Ma (MSWD = 0.59). 4 metamorphic rim domains of zircon
18 record $^{207}\text{Pb}/^{206}\text{Pb}$ concordant ages ranging from 2147 to 2215 Ma with a weighted
19 mean age of 2167 ± 58 Ma (MSWD = 0.14), consistent with the upper intercept age of
20 2165 ± 68 Ma defined by 6 spot analyses of rim domains within error. In addition,
21 granulite-facies mineral inclusions such as garnet, clinopyroxene and plagioclase
22 (Table 3; Fig. 10h, k) were found in metamorphic domains. In this regard, the age of
23 2167 ± 58 Ma records the timing of granulite-facies metamorphism whilst the $2480 \pm$
24 49 Ma age records the formation time of its igneous precursor.

25 Twenty U-Pb spot analyses were made on 20 zircon grains from sample 07JG32
26 (Group 3: Table 4; Fig. 11c). Two different textural populations of zircon can be easily
27 discerned by means of integrated CL imaging, i.e. core and rim, each one recording
28 discrete ages (Fig. 11c). The 5 analyses of the zircon core domains record $^{206}\text{Pb}/^{238}\text{U}$
29 concordant ages with a weighted mean age of 2121 ± 32 Ma (MSWD = 0.86) and 15
30 spot analyses of rim domains give a $^{206}\text{Pb}/^{238}\text{U}$ concordant age of 1805 ± 43 Ma. As

1 mentioned above, there are igneous zoning in the core domains (Fig. 10 m and n) and
2 inclusions of garnet + rutile + quartz + plagioclase + clinopyroxene in the rim
3 domains (Table 3; Fig. 10i, o and p) for zircon from the sample 07JG32. Therefore,
4 these data suggest that the 2121 ± 32 Ma age represents its formation time whereas
5 the 1805 ± 43 Ma age indicates the timing of the granulite-facies metamorphic stage.

6 In conclusion, zircons from the 3 dated xenolith samples, especially samples
7 07JG34 and 07JG32, exhibit clear core-rim patterns evidenced by CL images (and
8 mineral inclusions). Samples 07JG30, 07JG34 and 07JG32 were formed at 393 ± 7
9 Ma, 2480 ± 49 Ma and 2121 ± 32 Ma, and the latter two samples experienced
10 granulite-facies metamorphism at 2167 ± 58 Ma and 1805 ± 43 Ma, respectively. If
11 combined with the reported sample 07JG14 from the same area, which recorded three
12 metamorphic stages at 176 ± 2 Ma, 387 ± 13 Ma and 1811 ± 19 Ma (Liu et al., 2009),
13 it is therefore consistent that the studied xenoliths derived from igneous precursors of
14 various age (ca. 2.5 Ga, 2.1 Ga and 393 ± 7 Ma) which underwent variable degrees of
15 metamorphic overprinting or reworking at ca. 2.1 Ga, 1.8 Ga, 393 ± 7 Ma and 176 ± 2
16 Ma respectively, these ages strongly depending on their formation time and derivation
17 depths or crustal levels.

18 19 5.3. Zircon Lu–Hf isotope data

20 Some of the zircons analyzed for U–Pb were also analyzed for their Hf isotopic
21 composition. As already mentioned, the Hf isotopic composition of zircons provides
22 valuable and reliable information on the source and age of the magmatic protolith, and
23 it is also a very useful way to discriminate between magmatic and inherited
24 components. Ablation spots for the Hf analyses were sited at the same spots used for
25 U–Pb dating analyses. The *in situ* Hf isotopic analyses of zircons from six samples are
26 listed in Table 5 and shown in Figs. 12 and 13.

27 Zircon $\epsilon_{\text{Hf}}(t)$ values (+ 0.5 to + 6.7) with T_{DMI} ages of 786 ± 40 Ma (MSWD =
28 0.47, n = 4), ranging from 754 to 1003 Ma for sample 07JG30 of Group 1 (Table 5
29 and Fig. 12a) are significantly indicative of mantle-derived components but probably
30 contaminated by older crust because of the existence of inherited zircons (Fig. 10c).

1 The wide range of positive $\epsilon_{\text{Hf}}(t)$ values (+1.0 to +12.1) for the samples of Group 2
2 suggests the derivation of the parental magmas of the zircons by a range of processes,
3 including the melting of minor ancient crust and the addition of juvenile material. The
4 Hf-isotope data for most zircons with ages 2.5–2.4 Ga indicate the early generation of
5 juvenile crust, followed by reworking. However, the samples of Group 2 have older
6 T_{DM2} ages, e.g. 2777 ± 31 Ma (MSWD = 2, n = 18) and 2720 ± 53 Ma (MSWD = 1.5,
7 n = 4) for samples 07JG34 and 07JG12, respectively (Table 5 and Fig. 12b and c). In
8 contrast, the sample 07JG32 of Group 3 have variable $\epsilon_{\text{Hf}}(t)$ values of –5.1 to +3.8
9 with a weighted mean T_{DM2} age of 2518 ± 47 Ma (MSWD = 0.43, n = 7) (Table 5 and
10 Fig. 12e).

11 In comparison, the sample 07FY01 from the exposed meta-basement at Bengbu
12 displays high $\epsilon_{\text{Hf}}(t)$ values of +8.1 to +10.3 and +1.3 to +3.5 at 2.1 and 1.8 Ga,
13 respectively, with a Hf T_{DM2} age of 2137 ± 31 Ma (MSWD = 1.08, n = 8) (Table 5 and
14 Fig. 12f). This Hf T_{DM2} age of 2137 ± 31 Ma probably shows its emplacement age
15 while the timing of its HP granulite-facies metamorphic stage is 1839 ± 31 Ma (Liu et
16 al., 2009).

18 **6. Discussion**

19 *6.1. Composition and petrogenesis of the protoliths*

20 The composition of the deep crust at the southeastern margin of the NCC can be
21 constrained by investigations of chemical compositions, metamorphic evolution and
22 isotope dating of the xenoliths. As mentioned above, most of the protoliths for the
23 studied samples are basalt or basaltic andesite, and a few are magnesian andesite and
24 dacite. Plagioclase-rich xenoliths (e.g., sample 07JG06) are characterized by
25 pronounced positive Eu and Sr anomalies (Figs. 6 and 7) and relatively enriched
26 LREE patterns, in addition to higher contents of Al_2O_3 and Ba (Table 2 and Fig. 9).
27 Altogether, these features indicate that plagioclase accumulation played a significant
28 role in the fractionation process and their protoliths were gabbros.

29 The precursors of Group 1 xenoliths are mainly composed of basalts or basaltic

1 rocks with extremely high Mg-number values (62–77), relatively low REE contents,
2 high positive $\epsilon_{\text{Hf}}(t)$ values (+ 0.5 to + 6.7) and young formation time, strongly
3 suggesting a mantle-derived magma origin at 393 ± 7 Ma. This is in agreement with
4 the above petrographic observations such as garnet occurrence as coronae around
5 spinel (Fig. 3a) and exsolution lamellae in clinopyroxene (Fig. 3b), suggestive of
6 spinel-garnet-facies conditions. In addition, incompatible elements, such as high field
7 strength elements (HFSE, e.g. Nb, Ta, Zr and Hf), are generally insensitive to
8 fractional crystallization, secondary alteration and degrees of partial melting in the
9 source region and thus can be used to evaluate the petrogenesis of the xenoliths. In the
10 spider diagrams (Fig. 7a), some of the xenoliths show Nb depletion and Ba and Zr–Hf
11 enrichment as the possible result of lower crustal contamination (e.g. Sun &
12 McDonough, 1989), probably during their ascent; in fact, such contamination would
13 produce negative Nb–Ta anomalies but positive Zr–Hf anomalies in a spider diagram.

14 Particularly distinctive for the xenoliths of Group 2 is the presence of large rutile
15 grains and abundant garnet (Liu et al., 2009), similar in composition to the
16 rutile-bearing garnet granulites from Lesotho (Rogers, 1977) and Kola (Kempton et
17 al., 2001), most likely suggesting a similar petrogenesis. The protoliths of Group 2
18 xenoliths were basalt or basaltic andesite or gabbro (Figs. 4–6) and formed at ~ 2.5 Ga
19 with subsequent granulite-facies metamorphic overprinting at ~ 2.1 and/or ~ 1.8 Ga,
20 strongly depending on their crustal-levels or depths of provenance. Some samples
21 have typically high Sr contents (>400 ppm) and Sr/Y ratios, suggestive of adakitic
22 affinity (Defant and Drummond, 1990) (Table 2; Figs. 7 and 8). In view of the
23 possible occurrence of an adakitic rock–basaltic andesite–Nb-enriched basalt
24 association (Figs. 4–8), it is suggested that at least part of Group 2 xenolith’s
25 protoliths formed in a subduction-related arc setting which however probably
26 involved a multistage process (e.g. Calmus et al., 2003; Wang et al., 2008;
27 Manikyamba and Kerrich, 2011, and references therein). This interpretation is also
28 supported in these samples, by multiple hydrous minerals such as Ti-rich and Ti-poor
29 amphiboles, whereas the Ti-rich amphibole generally occurs as inclusions in garnet or
30 plagioclase (Liu et al., 2009) indicative of hydrous precursors. However, the primitive

1 mantle-normalized spider diagrams (Fig. 7b) of garnet granulites and amphibolites
2 from the Group 2 display variable geochemical characteristics, probably resulting
3 from metamorphic overprinting and reworking of various intensity. Some samples
4 have pronounced troughs at Nb–Ta, Th and Zr, and peaks at Zr and Sr which could be
5 ascribed to the metasomatic enhancement of LIL elements in comparison to HFSE.
6 Most samples of this type display flat REE patterns. However, samples 07JG05-2 and
7 07JG34 are light REE-enriched, while sample 07JG07 exhibits the enrichment of
8 heavy REEs. Such variations of REE suggest a metasomatic overprint (see below).

9 Two samples of Group 3 have magnesian andesite and dacite protoliths with
10 similar geochemical characteristics (Figs. 4 and 7c), characterized by high Sr contents
11 and Sr/Y ratios typical of adakitic rocks (Table 2 and Fig. 8), providing evidence for a
12 similar source and petrogenetic history. They display marked negative anomalies at
13 Nb-Ta, P and Ti anomalies, and positive Ba anomalies, typical of an arc signature. In
14 addition, the magnesian andesite is generally considered as deriving from the melting
15 of arc mantle wedge peridotites having interacted with acidic melts originated from
16 the subducted oceanic crust (e.g., Yogodzinski et al., 1995; Calmus et al., 2003). High
17 Cr (204 ppm) and Ni (97.8 ppm) contents for sample 07JG32 of Group 3 xenoliths
18 (Table 2) strongly support this scenario. In this context, the precursors for Group 3
19 most likely generated from the northward oceanic subduction beneath the
20 southeastern margin of the NCC at ~2.1 Ga, giving way to metamorphic overprinting
21 of the ≥ 2.5 Ga lower crustal rocks. Therefore, combined with Hf T_{DM2} ages with a
22 peak at 2518 ± 47 Ma (MSWD = 0.43, n = 7) (Fig. 12d) and $\epsilon_{Hf}(t)$ values of -5.1 to
23 $+0.3$ (Table 5), the Group 3 xenoliths were probably derived from arc mantle wedge
24 peridotites metasomatized by adakitic melts, and contaminated by ca. 2.5 Ga ancient
25 crustal materials during their emplacement at ca. 2.1 Ga. In addition, similar to the
26 Group 2, Group 3 xenoliths also recorded granulite-facies metamorphism at 1805 ± 43
27 Ma, originated from coeval large-scale mantle-derived magma activity as proposed by
28 Liu et al. (2009).

29 In particular for the xenoliths of Groups 2 and 3, the predominance of garnet-rich
30 mineral assemblages is consistent with a HP petrogenesis because of the thicker crust,

1 probably related to large igneous provinces which were documented to be connected
2 to the presence of mantle plumes at ca. 1.8 Ga (Liu et al., 2009), especially if
3 accompanied by rifting (e.g., White et al., 1987; Kempton et al., 2001). The higher
4 temperatures associated with an ascending plume would also lead to more magnesian
5 parental magmas, and the plume source would result in high Cr and Ni contents and
6 Mg-number values, in agreement with the data presented in this study. Thus, most
7 xenoliths of Groups 2 and 3, especially those with adakite-like Nb-enriched basalt and
8 magnesian andesite precursors, contain significant contributions from slab melts,
9 slab-derived fluids and the mantle wedge (Martin et al., 2005), involved in multistage
10 metasomatism rather than a simple, single stage history.

11 However, although sample 07JG32 of Group 3 and sample 07FY01 from the
12 metamorphosed basement formed at the same age of ~2.1 Ga, they have different ϵ_{Hf}
13 (t) values of - 5.1 to + 0.3 and + 8.1 to + 10.3 (Table 5 and Fig. 13). We suggest that
14 the latter sample probably represents a contemporaneous mantle-derived magmatism
15 with the ~2.1 Ga subduction-related arc magmatism, but both of them subsequently
16 experienced different evolutionary processes (Fig. 13).

17 In summary, the above petrologic, geochronological and geochemical data have
18 turned out to be complex with significantly different assemblages of mafic-ultramafic
19 and intermediate-acid rocks in the deep crust beneath the region. It is difficult to relate
20 these xenoliths to a single mantle source or petrogenetic process. However, it is clear
21 that the xenoliths of Groups 2 and 3 are dominated by hornblende-bearing
22 assemblages with hydrous precursors, similar to those formed at convergent
23 continental margins such as the northern Andes (Weber et al., 2002). **In addition,**
24 subduction generally brings large quantities of water and hydrous minerals deep into
25 the mantle, resulting in a large volume of aqueous fluids in the mantle at continental
26 convergent margins. As a consequence, the basaltic magma **was** rich in hydrous fluids
27 and the consequently formed rocks contain a large quantity of hydrous minerals such
28 as amphibole. Therefore, based on these lithologic associations in the xenoliths and
29 their geochronological and geochemical characteristics, it is suggested that there were
30 at least two stages of northward oceanic subduction beneath the region at ~2.5 and

1 ~2.1 Ga, respectively. In this regard, at least before 2.1 Ga the southern margin of the
2 North China Craton must have faced an open ocean, and should not have been
3 connected to any other continental blocks. Moreover, the Group 1 xenoliths were
4 resulted from a young mantle-sourced magmatism characterized by positive $\epsilon_{\text{Hf}}(t)$
5 values of + 0.5 to + 6.7 and extremely high Mg-number (62–77) values, giving rise to
6 various degrees of enrichment in MgO, Cr and Ni (see below) on the Precambrian
7 lower crust at about 390 Ma.

9 *6.2. Precambrian episodic crustal growth*

10 The compositional, geochronological and geochemical data of the xenoliths
11 mentioned above suggest that both arc and plume components were present in the
12 deep crust and the lower crust comprises chiefly subducted-related components,
13 which can be related to the regional geology to produce a model of how and when the
14 deep crust formed in the southeastern margin of the NCC.

15 The data on zircon U–Pb ages and Hf isotopes presented in this work clearly
16 attest multistage thermal events during the crustal evolution in this region. At least
17 three major groups of igneous zircons, suggesting different magma crystallization
18 ages and interpreted as the formation time for the precursors of the xenoliths, can be
19 evidently recognized. The first zircon population, recorded in the samples of Group 2,
20 indicates that they formed at about 2.5 Ga, although the formation of the primordial
21 lower crust of the region might have occurred even earlier, **at ca. 2.7–2.8 Ga (e.g.,**
22 **Tang et al., 2007; Zheng et al., 2009; Jiang et al., 2010; Zhai and Santosh, 2011;**
23 **Zhang, 2012; this study)**. The second zircon population from the Group 3 xenoliths
24 defined ages of ca. 2.1 Ga and the third population gave ages of ca. 390 Ma.
25 Combined with the petrographic and geochemical data provided in this paper, the
26 former two ages represent the formation age of a subduction-related arc at a
27 convergent continental margin and the latter age is related to a young mantle-sourced
28 magmatism. Thus it is proposed that multiple crustal growth and subsequent
29 metamorphic events, and interactions between the ascending mantle-derived magma
30 and the lower crust result in high Cr and Ni contents for some xenoliths of Groups 2

1 and 3 (e.g., samples 07JG06, 07JG12 and 07JG32) (Table 2).

2 Moreover, as previously proposed by Liu et al. (2009), the ca. 1.8 Ga HP
3 granulite-facies metamorphic event may be ascribed to large-scale crustal heating and
4 thickening related to mantle-derived basaltic underplating at the base of the lower
5 crust, as evidenced by widespread extension, rifting and related mafic magma
6 emplacement in the NCC during this period, in response to a vertical accretion in the
7 region. This metamorphic event can thus be considered to be plume-related,
8 eventually contributing to the initiation of the break-up of the NCC from Columbia at
9 ca. 1.8 Ga (Zhai and Santosh, 2011).

10 Therefore, the ages derived from zircon in the deep-seated xenoliths correlate
11 well with the lower crustal history in this area, providing robust evidence of at least
12 three major periods of crust formation attributed to subduction–accretion at ca. 2.5 Ga
13 and 2.1 Ga, and magmatic underplating at ca. 1.8 Ga. The former event is consistent
14 with the dominant age of the NCC Archean basement (e.g., Zhai et al., 2000; Zhao et
15 al., 2000; Kusky and Li, 2003; Zhai and Liu, 2003; Gao et al., 2004; Guo et al., 2005;
16 Kröner et al., 2005; Zhai et al., 2005; Hou et al., 2006; Liu et al., 2011a; Zhai and
17 Santosh, 2011; Zhang, 2012). In view of predominant two-stage Hf model ages of
18 2777 ± 31 Ma and 2720 ± 53 Ma (Fig. 12a and b) and inherited zircon ages of $2631 \pm$
19 43 Ma (Fig. 11b), the 2.7 - 2.8 Ga event might represent an older crustal growth event
20 in the area or in the whole of NCC as suggested by some researchers (Tang et al.,
21 2007; Zheng et al., 2009; Jiang et al., 2010; Zhai and Santosh, 2011), but remains to
22 be further constrained by more data. Consequently, different types of xenoliths
23 suffered various degrees of metamorphic overprinting and modification, strongly
24 depending on their precursor ages and derivation depths.

25 26 *6.3. Multiple metasomatism and modification of the lower crust*

27 Although metasomatism occurs commonly in mantle-derived ultramafic xenoliths
28 (e.g., Hawkesworth et al., 1990; Kempton et al., 2001), there is little evidence for its
29 widespread occurrence in the lower crust. Notable exceptions are the

1 amphibole-bearing Pl + Cpx ± Grt ± Opx granulites from the Eifel, Germany (Stosch
2 and Lugmair, 1984) and amphibole-bearing Cpx + Grt + Pl ± Opx granulites from
3 north Queensland (Stolz and Davis, 1989), as well as the Kola granulite xenoliths
4 where phlogopite is the predominant hydrous phase (Kempton *et al.*, 2001). In
5 contrast, K-rich (i.e. mica-rich) metasomatism of the lowermost crust–mantle
6 boundary may be common for thick continental roots, although these are probably lost
7 through delamination and recycling in tectonically active areas, and may only rarely
8 be preserved (Kempton *et al.*, 2001). However, the southeastern margin of the NCC
9 experienced a complex crustal evolution involved in multistage tectonothermal events
10 and related metamorphism, thus the deep-seated xenoliths in this area record
11 petrographic and geochemical evidence for multiple metasomatic overprinting,
12 resulting in variable degrees of modification. As an example, **the pervasive occurrence**
13 **of multiple amphiboles and biotites or phlogopites** in various xenoliths, especially in
14 Groups 2 and 3, implies a water-rich precursor and multistage fluid or melt or mantle
15 metasomatism and widespread modification of the lower crust in this region.

16 Small amounts of Ti-rich biotite in textural equilibrium with a HP granulite-facies
17 assemblage in the xenoliths of this region have been already reported (Liu *et al.*,
18 2009). The Ti-rich biotite- and hornblende-bearing HP granulite-facies metamorphism
19 have been documented to take place at about 1.8 Ga, with green hornblende and
20 chlorite being originated from the late stages (see before). The green hornblende
21 probably formed at 176 ± 2 Ma as defined by U-Pb dating of zircon overgrowth rim
22 (Liu *et al.*, 2009) and Ar-Ar dating of hornblende (Liu Y.-C. *et al.*, unpublished data),
23 whereas the chlorite was possibly produced by a more recent fluid-driven
24 metasomatic event according to the petrographic observation. In this regard, the
25 xenoliths suffered from at least three episodes of metasomatism: an older (late
26 Palaeoproterozoic) one responsible for the growth of Ti-biotite and hornblende, and
27 two younger (middle Jurassic to early Cretaceous) ones related to their exhumation
28 and the related magmatism causing the growth of Ti-poor amphibole and the
29 formation of chlorite. The former metasomatism was probably the result of the
30 introduction of a fluid phase during HP metamorphism as suggested by Otten (1984).

1 Hydrous phase assemblages (e.g. green hornblende) partially replacing the
2 granulite-facies minerals (e.g. clinopyroxene) (Fig. 3d and e) also suggest a fluid
3 influx at amphibolite-facies conditions. An additional metasomatic event may be
4 related to mantle-derived fluids or melts that resulted in the formation of phlogopite
5 and metamorphic overprinting and modification on the rocks beneath the deep crust in
6 this region, which produced MgO, Cr and Ni enrichments in some samples of Groups
7 2 and 3 during recent events such as 393 ± 7 Ma mantle-derived magmatism. Also, the
8 LREE enrichment exhibited by some samples of Group 1 (sample 08JG09) and by
9 amphibolites of Group 2 (samples 07JG05-2 and 07JG34) is more likely the result of
10 local metasomatic modifications (Dessai et al., 2004), causing the formation of
11 phlogopite and/or amphibole and the consequent enrichment in LILE (e.g. Sr, Ba and
12 K) and REE. The samples containing abundant amphiboles tend to be richer in REEs
13 and show more LREE enrichments than the others (Fig. 6). Similar cases have been
14 previously observed (e.g., Al-Mishwat and Nasir, 2004).

15 In addition, as mentioned above some xenoliths have been metasomatised by
16 subduction-related fluid or melt, leading to varying degrees of Ba and Nb enrichments
17 (Table 2 and Fig. 9) at a convergent continental margin setting. Furthermore, during
18 the ascent/uplift process, some of the xenoliths might have been metasomatized by
19 K-rich fluids or melts, probably derived from phlogopite breakdown at depth, leading
20 to the development of K- and Rb-rich mica and K-feldspar (Fig. 3c and e) and
21 large-ion lithophile (LILE) (Ba, Rb, K, and Sr) elemental enrichment (Table 2 and Fig.
22 7).

23 Therefore, multiple metasomatic episodes have been recognized in the xenoliths
24 from the southeastern margin of the NCC. Moreover, many xenoliths of Groups 2 and
25 3 suffered from a variety of multiple mantle metasomatic processes during their
26 evolution. However, more detailed geochronological and petrological studies will be
27 required in order to define the precise features and timing of metasomatism in the
28 area.

30 *6.4. Possible eastern extension of the northern Qinling orogen beneath the*

1 *southeastern margin of the NCC*

2 As mentioned above, a previously unknown late mantle-derived basaltic
3 magmatic event at 393 ± 7 Ma has been recognized and resulted in metamorphic
4 overprinting **and modification** on some rocks of the Precambrian lower crust at depth
5 in the southeastern margin of the NCC. However, in the northern Qinling orogen,
6 there is an extensive tectono-thermal event at about 400 Ma as reported by Mattauer
7 et al. (1985), Li et al. (1989), Lerch et al. (1995), Xue et al. (1996), Hacker et al.
8 (1998) and Zhai et al. (1998). Recently, Liu et al. (2006) also reported a SHRIMP
9 U-Pb age of 386 ± 20 Ma dated on one metamorphic overgrowth rim in zircon from
10 the garnet-bearing epidote-mica-quartz schist of 465 Ma **with an arc affinity** in the
11 Beihuaiyang zone of the Dabie orogen. In this context, it seems plausible that the
12 northern Qinling orogen continues to the eastern segment of the Dabie orogen and
13 then beneath the southeastern margin of the NCC in the Suzhou region or even further
14 to the East. **This extension might be presently covered** by the Neoproterozoic to late
15 Paleozoic strata and by the Mesozoic basin sediments. Based on 786 ± 40 Ma
16 Hf-model ages and positive $\epsilon_{\text{Hf}}(t)$ values ranging mostly from +3 to +6 (Table 5 and
17 Fig. 12) for magmatic zircons from 393 ± 7 Ma spinel-bearing garnet clinopyroxenite,
18 it is inferred that there was a Neoproterozoic oceanic crust (now vanished) between
19 the NCC and Yangtze Craton. This possibility, recently demonstrated by Peng et al.
20 (2012) at the northern margin of the Yangtze Craton, however remains to be further
21 investigated and constrained by more tight evidences.

22 23 **7. Conclusions**

24 The occurrence and characteristics of diverse xenoliths at Jiagou near Suzhou
25 implies that the actual deep crustal structure and composition of the southeastern
26 margin of the NCC may be the result of multiple magmatic activities and
27 metamorphic processes. Based on integrated petrologic, geochronological, Hf isotope
28 and geochemical data, a major result of our study is that different types of xenoliths in
29 the region are firstly documented to be the end-products of multistage petrogenetic

1 processes involving several sources and protolith ages, as well as variable degrees of
2 metamorphic overprinting and modification, strongly depending on the rock types and
3 their evolutionary histories.

4 The deep-seated xenoliths from the Mesozoic Jiagou dioritic porphyry in the
5 southeastern margin of the NCC comprise a suit of mostly mafic rock types such as
6 garnet amphibolite, garnet granulite, spinel-bearing garnet clinopyroxenite,
7 garnet-bearing mafic gneiss and spinel pyroxenite (websterite) with minor
8 garnet-bearing hornblendite and felsic gneiss, some of which are characterized by HP
9 granulite-facies metamorphism, probably reflecting major crustal thickening and
10 representing the lowermost part of lower crust. Whole-rock major, trace element and
11 rare earth element characteristics show that the xenoliths include ultra-mafic, mafic
12 and intermediate-acid groups. Most of the protoliths are alkaline and subalkaline
13 basalt, while a few are gabbro, **basaltic or magnesian andesite** and dacite. Some of the
14 xenoliths probably had the same mantle source but experienced different modification
15 by late events (e.g., metasomatism), and formed at the base of the lower continental
16 crust by mantle-derived magma underplating.

17 Geochronological and geochemical features of different types of xenoliths testify
18 the significant compositional complexity of the deep crust in the southeastern margin
19 of the NCC. SHRIMP zircon U–Pb dating and geochemical data of the xenoliths
20 presented in this study, combined with their tectonic setting, provide the solid
21 evidence for the existence of subduction-related adakite-like and arc-related rocks in
22 the southeastern margin of the NCC at ca. 2.5 Ga and 2.1 Ga. Our data further
23 confirm the occurrence of HP granulite-facies metamorphism at ca. 1.8 Ga, involved
24 in a multistage process of continental growth in response to two stages of
25 subduction–accretion and one vertical accretion of basaltic magma at the base of the
26 lower crust, respectively. The 2.7 to 2.8 Ga ages defined by inherited zircons and
27 depleted mantle zircon-Hf model ages could record an earlier crustal growth episode
28 in the area.

29 The compositional and age variability shown by the xenoliths suggests that their
30 igneous protoliths are the products of several episodes of subduction or magmatic

1 underplating and tectonic activity, though most samples of Groups 2 and 3, excluding
2 Group 1, shared a common metamorphic evolution after 1.8 Ga. Therefore, the
3 studied xenoliths have different origins, formed at different ages and originated at
4 different levels of the deep crust, and thus are unlikely to share a common formational
5 and evolutionary history as previously claimed.

6 Furthermore, on the base of these data the northern Qinling orogen most likely
7 extends into the southeastern margin of the NCC and is covered by the
8 Neoproterozoic to late Paleozoic strata and by the Mesozoic basin sediments
9 widespread in the region.

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Figure captions

Figure 1 Schematic geological map of the Qinling–Dabie–Sulu collision zone and adjacent parts of the North China Craton (modified after Xu *et al.*, 2006), with inset showing the major tectonic divisions of China and the location of the study area. YZ and SC denote the Yangtze Craton and South China Orogen, respectively. Also shown are the subdivisions of the North China Craton (Zhao *et al.*, 2000), where WB, TNCO and EB denote the Western Block, Trans-North China Orogen and Eastern Block, respectively.

Figure 2 Photographs showing the field occurrence of deep-seated xenoliths of spinel-bearing garnet clinopyroxenite (a), garnet granulite (b), garnet amphibolite (c) and basic gneiss (d) in the Mesozoic dioritic porphyry at Jiagou near Suzhou.

Figure 3 Microstructures of deep-seated xenoliths from Jiagou in the southeastern margin of the North China Craton. (a) Garnet corona surrounding spinel and oriented garnet needles in clinopyroxene, spinel-bearing garnet clinopyroxenite (sample 07JG03), plane polarized light (PPL); (b) Oriented garnet needles in clinopyroxene, spinel-bearing garnet clinopyroxenite (sample 07JG30), PPL; (c) Phlogopite and intergrowth of clinopyroxene + phlogopite + K-feldspar in phlogopite clinopyroxenite (sample 08JG06), PPL; (d) Clinopyroxene inclusion in garnet and its transformation to a symplectite of hornblende + plagioclase in the matrix, garnet granulite (sample 08JG15), PPL; (e) Back scattered electron (BSE) image showing clinopyroxene (darker) partially replaced by hornblende; K-feldspar and plagioclase occur in the fractures, garnet-bearing amphibolite (sample 07JG12); (f) Garnet + plagioclase + hornblende + ilmenite after rutile, garnet-bearing hornblende-plagioclase gneiss (sample 07JG32), PPL.

Figure 4 $\text{SiO}_2\text{--Zr/TiO}_2$ diagram (after Winchester and Floyd, 1977) for the studied xenoliths in the Mesozoic dioritic porphyry at Jiagou. Symbols used for the three

1 groups of xenoliths are the same as in the following figures.

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4 **Figure 5** SiO₂-Mg# (a) and Nb-Mg# (b) diagrams for the deep-seated xenoliths from
5 Jiagou. HNB and NEB denote high-Nb basalt and Nb-enriched basalt, respectively
6 based on Sajona et al. (1996) and Martin et al. (2005). Symbols as in Fig. 4.
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12 **Figure 6** Chondrite-normalized rare earth element diagrams for Groups 1 (a), 2 (b)
13 and 3 (c) of the xenoliths from Jiagou. Normalization values are from Sun and
14 McDonough (1989).
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21 **Figure 7** Primitive-mantle-normalized spider diagrams for Groups 1 (a), 2 (b) and 3
22 (c) of the xenoliths from Jiagou. Normalization values are from Sun and McDonough
23 (1989).
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29 **Figure 8** Sr/Y-Y diagram (Defant and Drummond, 1990; Martin, 1999) for Groups 2
30 and 3 of the xenoliths from Jiagou.
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35 **Figure 9** Plot of Ba-Nb/Y for the deep-seated xenoliths from Jiagou. Symbols as for
36 Fig. 4.
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42 **Figure 10** Cathodoluminescence (CL) images for zircon from samples 07JG30 (a-c),
43 07JG34 (d-k) and 07JG32 (l-p). The open circles are spot analysis with available
44 ²⁰⁶Pb/²³⁸U ages.
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50 **Figure 11** Zircon SHRIMP U-Pb dating for the xenoliths from samples 07JG30 (a),
51 07JG34 (b) and 07JG32 (c) at Jiagou. All errors are given at 2σ level.
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56 **Figure 12** Histograms of single-stage depleted mantle Hf model ages (T_{DM1}) for the
57 zircons from sample 07JG30 (a) and two-stage depleted mantle Hf model ages (T_{DM2})
58 for the zircons from samples 07JG34 (b), 07JG12 (c), 07JG14 (d) and 07JG32 (e) of
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1 the xenoliths at Jiagou and the exposed meta-basement (sample 07FY01) (f) at
2 Bengbu.

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4 **Figure 13** The relationship between U-Pb age and $\epsilon_{\text{Hf}}(t)$ values for the igneous core
5 and metamorphic rim of zircons as identified by the SHRIMP U-Pb dating for four
6 xenoliths (samples 07JG34, 07JG12, 07JG14 and 07JG32) from Jiagou and one
7 garnet amphibolite (sample 07FY01) from the exposed basement at Bengbu. Deleted
8 Mantle (DM) and CHUR evolution lines are also shown. The solid and open spots for
9 different samples denote $\epsilon_{\text{Hf}}(t)$ values at their precursor and corresponding
10 metamorphic ages, respectively. **Dashed arrow lines indicate the core–rim analyses of**
11 **zircons with the arrow pointing to the rim.**

Table 1 Electron microprobe analyses (wt. %) of representative minerals from the xenoliths at Jiagou in the southeastern margin of the North China Craton

Mineral	Garnet			Pyroxene				Amphibole		Phlogopite		Feldspar	
	07JG03	07JG03	07JG30	07JG03	07JG30	08JG06	08JG15	07JG30	08JG15	08JG06	08JG06	07JG12	07JG12
Sample	07JG03	07JG03	07JG30	07JG03	07JG30	08JG06	08JG15	07JG30	08JG15	08JG06	08JG06	07JG12	07JG12
Site	m	i	i	m	m	m	i	m	m	Sy	m	Sy	Sy
SiO ₂	38.91	38.98	38.88	51.10	49.66	53.23	52.58	40.80	43.78	39.10	39.77	65.08	62.92
TiO ₂	0.00	0.00	0.11	0.00	1.23	0.03	0.12	1.12	0.51	2.02	1.89	0.00	0.00
Al ₂ O ₃	21.84	21.62	21.79	4.11	7.72	1.71	2.77	17.08	11.55	13.85	15.37	18.45	23.12
Fe ₂ O ₃	1.77	1.83	0.86	3.85	0.00	0.00	0.00					0.00	0.00
FeO	16.59	16.95	14.68	1.97	5.32	5.67	8.88	9.94	18.01	11.56	11.77	0.00	0.00
MnO	0.00	0.00	0.00	0.00	0.05	0.11	0.10	0.00	0.12	0.07	0.06	0.00	0.00
MgO	5.24	5.55	3.84	14.83	11.61	14.40	12.46	12.19	9.91	18.81	16.59	0.00	0.00
CaO	16.08	15.43	19.51	23.60	22.71	24.43	21.37	11.99	10.88	0.06	0.01	0.00	4.80
Na ₂ O	0.00	0.00	0.00	0.53	1.42	0.27	1.36	2.75	1.80	0.08	0.13	1.80	8.58
K ₂ O	0.00	0.00	0.00	0.00	0.01	0.02	0.00	1.07	0.75	8.94	10.17	14.65	0.57
Total	100.43	100.36	99.67	99.99	99.73	99.87	99.64	96.94	97.31	94.49	95.76	99.98	99.99
O	12	12	12	6	6	6	6	23	23	12	12	8	8
Si	2.97	2.97	2.99	1.88	1.83	1.97	1.96	6.01	6.49	2.90	2.88	2.98	2.79
Al ^{IV}	0.03	0.02	0.01	0.12	0.17	0.03	0.04	1.02	1.00	1.32	1.20	1.00	1.21
Al ^{VI}	1.93	1.92	1.95	0.05	0.16	0.04	0.08	0.97	0.51				
Fe ³⁺	0.10	0.10	0.05	0.10	0.04	0.01	0.06	0.00	0.78			0.00	0.00
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.12	0.06	0.10	0.11	0.00	0.00
Fe ²⁺	1.06	1.08	0.94	0.06	0.12	0.16	0.22	1.22	1.46	0.72	0.71	0.00	0.00
Mg	0.59	0.63	0.44	0.81	0.64	0.79	0.69	2.68	2.19	1.80	2.06	0.00	0.00
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00
Ca	1.31	1.26	1.60	0.93	0.89	0.97	0.85	1.89	1.73	0.00	0.00	0.00	0.23
Na	0.00	0.00	0.00	0.04	0.10	0.02	0.10	0.78	0.52	0.02	0.01	0.16	0.74
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.20	0.14	0.95	0.84	0.86	0.03

Note: m, matrix; i, inclusion or exsolution; Sy, symplectite. Garnet/pyroxene stoichiometries and the amount of Fe³⁺ and Fe²⁺ were estimated on the base of eight/four cations and the charge-balance constraint; Fe³⁺ content in amphibole was calculated as Si + Al + Ti + Mg + Fe + Mn = 13 for O = 23.

Table 2 Major and trace element compositions of the xenoliths from Jiagou in the southeastern margin of the North China Craton. Major oxides are in wt. %, trace elements in ppm.

Type	Group 1						Group 2		
Sample	07JG03	07JG17	07JG21	07JG30	08JG06	08JG09	07JG05-2	07JG06	07JG07
SiO ₂	42.88	46.06	48.08	42.02	47.42	45.4	46.08	52.72	39.10
Al ₂ O ₃	11.80	9.29	6.45	13.24	17.62	9.96	12.18	16.89	14.90
Fe ₂ O ₃	3.17	1.66	3.79	3.21	1.67	2.58	6.24	2.80	7.90
TiO ₂	1.02	1.27	0.48	1.04	1.04	0.64	1.98	0.38	3.01
FeO	7.33	7.07	8.45	7.59	6.32	5.13	10.35	6.87	11.86
CaO	17.36	20.48	13.83	18.52	7.00	19.93	12.31	6.06	7.79
MgO	11.64	11.32	14.63	9.75	8.00	14.02	5.90	6.65	9.56
K ₂ O	0.5	0.07	0.26	0.21	5.55	0.06	0.59	0.73	0.89
Na ₂ O	1.16	0.93	1.23	0.99	2.00	0.44	1.71	4.03	1.65
MnO	0.14	0.14	0.23	0.15	0.07	0.09	0.21	0.18	0.44
P ₂ O ₅	0.03	0.07	0.06	0.02	0.13	0.05	0.33	0.04	0.06
H ₂ O ⁺	1.14	0.36	0.74	0.73	1.55	0.62	0.68	2.26	2.09
LOI	2.88	1.66	2.09	2.77	2.8	2.14	2.22	3.09	3.03
Mg#	67	70	69	62	65	77	40	56	47
Cr	3.99	42.24	447.4	93.7	136.1	36.1	46.8	97.5	163.4
Ni	21.8	18.4	178.7	64.2	24.0	173.0	76.1	90.5	48.4
Rb	20.6	0.87	1.43	11.5	76.3	1.26	29.8	27.4	13.5
Sr	119.9	57.4	91.8	57.7	196.8	88.4	218.9	573.9	218.9
Y	15.4	18.9	15.1	13.8	20.9	12.6	79.9	6.37	77.7
Zr	99.0	103.0	68.0	83.9	68.7	202.9	248.0	38.8	46.2
Nb	2.08	1.27	1.34	2.54	3.96	1.86	42.3	1.32	4.21
Ba	190.6	19.4	34.3	100.6	1425.5	22.4	127.6	279.6	214.6
La	4.79	3.54	19.89	2.21	6.51	10.7	35.0	7.74	6.72
Ce	10.9	9.88	42.3	6.56	14.0	30.3	89.4	13.5	13.6
Pr	1.34	1.35	5.16	0.87	1.86	4.13	13.50	1.43	1.76
Nd	6.37	7.06	22.5	4.69	9.34	18.7	72.4	6.54	9.57
Sm	1.75	2.19	4.37	1.51	2.73	3.82	21.10	1.37	3.60
Eu	0.54	0.63	1.04	0.44	0.83	0.57	3.32	1.16	1.23
Gd	2.20	2.80	4.25	1.95	3.15	3.18	20.6	1.43	6.11
Tb	0.36	0.47	0.51	0.34	0.53	0.40	2.69	0.18	1.26
Dy	2.39	3.17	2.82	2.39	3.45	2.09	14.2	1.08	9.96
Ho	0.51	0.68	0.52	0.52	0.73	0.38	2.55	0.21	2.48
Er	1.45	1.95	1.44	1.51	2.06	1.06	6.75	0.59	7.90
Tm	0.22	0.30	0.20	0.23	0.28	0.14	0.96	0.09	1.32
Yb	1.61	2.14	1.44	1.65	1.78	0.94	6.80	0.66	10.30
Lu	0.25	0.33	0.21	0.25	0.26	0.14	1.02	0.10	1.68
Hf	1.57	2.26	1.70	1.77	1.67	3.83	5.79	0.75	1.08
Ta	0.30	0.29	0.13	0.39	0.20	0.55	2.23	0.07	0.25
Pb	66.9	21.6	56.1	110.0	14.7	5.34	23.2	29.6	38.5

Th	0.73	0.63	0.30	0.42	0.99	2.02	7.93	0.09	0.94
U	3.74	1.01	1.04	0.26	2.95	2.60	2.62	0.32	3.04
Type	Group 2				Group 3				
Sample	07JG12	07JG14	07JG29	07JG34	08JG08	08JG15	07JG32	08JG03	
SiO ₂	43.48	48.46	37.92	51.3	46.68	45.8	53.58	67.06	
Al ₂ O ₃	10.84	12.59	12.81	14.28	15.65	13.09	15.64	15.46	
Fe ₂ O ₃	6.61	1.87	8.51	3.20	3.23	3.82	4.08	0.88	
TiO ₂	5.21	1.70	1.95	1.33	1.47	1.03	0.53	0.21	
FeO	12.48	14.99	10.73	8.33	11.86	13.89	5.77	1.70	
CaO	7.79	10.44	8.43	7.95	10.92	11.24	8.37	3.60	
MgO	7.92	7.17	11.97	5.18	5.10	6.78	5.48	1.29	
K ₂ O	0.48	0.11	0.91	1.3	0.28	0.37	0.48	1.79	
Na ₂ O	2.21	0.79	1.84	3.96	2.84	1.8	3.77	5.86	
MnO	0.26	0.30	0.21	0.19	0.27	0.30	0.17	0.04	
P ₂ O ₅	0.12	0.05	0.03	0.36	0.19	0.16	0.019	0.11	
H ₂ O ⁺	1.56	0.47	6.06	1.45	0.80	0.45	1.61	0.89	
LOI	2.76	1.72	3.69	2.51	1.79	1.92	2.16	1.37	
Mg#	43	43	54	45	39	41	51	48	
Cr	252.4	25.9	97.7	80.0	62.4	78.7	202.1	79.4	
Ni	247.7	109.7	47.2	98.3	21.6	92.0	96.8	29.5	
Rb	3.97	1.42	8.77	23.8	2.00	3.68	6.46	38.7	
Sr	297.9	76.2	225.9	833.8	235.8	253.8	421.8	1473	
Y	14.9	37.2	18.5	22.3	25.0	35.8	14.8	4.98	
Zr	72.2	30.1	23.0	115.9	92.6	79.8	49.9	96.5	
Nb	15.4	2.92	2.20	9.39	8.17	4.13	2.24	5.11	
Ba	74.7	26.2	147.6	476.5	52.5	89.7	93.4	1299	
La	7.42	11.7	8.68	36.4	11.4	7.92	12.7	8.94	
Ce	17.5	19.2	14.7	80.3	24.9	16.6	19.2	17.4	
Pr	2.33	2.16	1.79	9.84	3.23	2.06	3.13	1.94	
Nd	11.7	9.76	9.45	42.8	15.6	10.3	14.2	8.06	
Sm	2.94	2.74	2.71	8.20	4.15	3.23	3.07	1.62	
Eu	0.99	1.04	0.91	1.66	1.46	0.96	0.90	0.53	
Gd	3.11	4.07	3.16	6.63	4.51	4.14	2.87	1.31	
Tb	0.43	0.75	0.49	0.77	0.71	0.73	0.40	0.15	
Dy	2.49	5.41	3.12	3.82	4.30	4.91	2.34	0.79	
Ho	0.48	1.25	0.66	0.67	0.88	1.09	0.47	0.15	
Er	1.31	3.65	1.85	1.84	2.48	3.25	1.35	0.46	
Tm	0.19	0.57	0.27	0.23	0.35	0.49	0.19	0.07	
Yb	1.36	4.16	1.81	1.51	2.28	3.32	1.27	0.46	
Lu	0.21	0.65	0.27	0.22	0.35	0.52	0.20	0.07	
Hf	1.62	0.87	0.70	2.35	2.26	1.80	1.09	2.56	
Ta	0.73	0.09	0.12	0.34	0.53	0.24	0.07	0.17	
Pb	15.7	4.06	75.5	20.2	1.56	6.05	13.8	27.6	
Th	0.21	0.36	0.37	0.71	1.18	0.74	0.29	1.51	

U	0.74	0.34	1.10	1.86	0.38	0.97	0.11	1.07
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Table 3 Electron microprobe analyses (wt. %) of representative minerals in zircon from the xenoliths at Jiagou in the southeastern margin of the North China Craton

Mineral	Garnet		Rutile	Plagioclase		
	07JG32	07JG34		07JG32		07JG34
No.						
Site	Zir-R	Zir-R	Zir-R	Zir-R	Zir-R	Zir-R
SiO ₂	39.24	50.62	0.28	58.75	59.62	60.65
TiO ₂	0.00	0.21	98.72	0.00	0.00	0.01
Al ₂ O ₃	22.11	3.83	0.00	25.97	25.56	24.64
FeO*	24.90	11.49	0.88	0.04	0.04	0.06
MnO	0.80	0.18	0.03	0.02	0.00	0.04
MgO	6.04	11.21		0.04	0.03	0.03
CaO	6.76	21.24		8.13	7.26	6.29
Na ₂ O	0.00	1.15		6.87	7.17	7.54
K ₂ O	0.01	0.00		0.16	0.21	0.57
Total	99.95	99.96	99.91	99.98	99.89	99.83
O	12	6		8	8	8
Si	3.01	1.94		2.63	2.66	2.70
Al ^{IV}	0.00	0.06		1.37	1.34	1.29
Al ^{VI}	1.91	0.04				
Fe ³⁺	0.06	0.07		0.00	0.00	0.00
Ti	0.00	0.00		0.00	0.00	0.00
Fe ²⁺	1.80	0.18		0.00	0.00	0.00
Mg	0.55	0.78		0.00	0.00	0.00
Mn	0.10	0.00		0.00	0.00	0.00
Ca	0.56	0.88		0.39	0.35	0.30
Na	0.00	0.05		0.59	0.62	0.65
K	0.00	0.00		0.01	0.01	0.03

Note: Zir-R denotes the rim domain of zircon. Garnet/pyroxene stoichiometries and the amount of Fe³⁺ and Fe²⁺ were estimated based on eight/four cations and the charge-balance constraint. * represent all Fe.

Table 4 SHRIMP zircon U-Pb data for the xenoliths at Jiagou in the southeastern margin of the North China Craton

Sample 07JG30											
spot	²⁰⁶ Pb _c %	U ppm	Th ppm	Th/U	²⁰⁶ Pb* ppm	²⁰⁷ Pb*/ ²³⁵ U	±%	²⁰⁶ Pb*/ ²³⁸ U	±%	²⁰⁶ Pb/ ²³⁸ U	Age(Ma) ±1σ
1.1	7.49	264	127	0.48	15.4	0.68	23	0.0629	2.5	393	10
2.1	1.13	506	48	0.10	56.1	1.265	3.9	0.1276	2.5	774	18
3.1	1.93	288	171	0.60	15.5	0.487	16	0.0614	2.3	384	8.6
4.1	0.15	800	86	0.11	97.1	1.429	2.7	0.1411	2.0	851	16
5.1	1.64	491	293	0.60	27.2	0.481	11	0.0634	2.2	397	8
6.1	0.33	255	130	0.51	59.5	4.00	2.6	0.2699	2.1	1541	29
7.1	1.59	444	129	0.29	25.4	0.474	8.2	0.0656	2.6	410	10
8.1	1.79	458	274	0.60	24.9	0.612	12	0.0622	2.2	389	8
9.1	5.38	285	159	0.56	16.2	0.62	16	0.0625	2.5	391	9
10.1	3.10	566	335	0.59	34.2	0.71	16	0.0681	2.3	425	10
11.1	2.97	286	343	1.20	18.8	0.660	12	0.0744	2.2	462	10
4.2	0.16	1058	121	0.11	111	1.191	2.8	0.1220	2.0	742	14
Sample 07JG34											
spot	²⁰⁶ Pb _c %	U ppm	Th ppm	Th/U	²⁰⁶ Pb* ppm	²⁰⁷ Pb*/ ²³⁵ U	±%	²⁰⁶ Pb*/ ²³⁸ U	±%	²⁰⁷ Pb/ ²⁰⁶ Pb	Age(Ma) ±1σ
1.1	0.00	13	16	1.23	3.91	7.91	5.4	0.358	4.2	2458	57
2.1	0.37	21	35	1.67	9.67	12.73	3.9	0.529	3.2	2602	38
3.1	0.64	34	56	1.65	14.50	11.02	4.0	0.491	2.6	2485	51
4.1	0.78	17	18	1.06	5.18	6.54	7.3	0.344	4.0	2203	110
5.1	0.93	22	48	2.18	9.15	11.50	6.1	0.488	3.9	2567	77
6.1	0.00	12	16	1.33	5.30	12.68	4.8	0.521	3.9	2622	46
7.1	1.83	11	12	1.09	4.22	9.45	7.1	0.445	4.1	2389	99
8.1	1.18	12	12	1.00	5.77	12.61	5.8	0.538	3.7	2559	74
9.1	0.00	13	23	1.77	4.10	6.95	6.8	0.362	5.0	2215	81
10.1	1.36	13	16	1.23	6.01	13.72	5.3	0.539	3.6	2696	65
11.1	1.26	14	19	1.36	5.33	10.49	6.5	0.452	4.2	2540	82
12.1	2.78	12	11	0.92	5.13	10.8	13	0.488	7.1	2466	190
13.1	1.18	11	11	1.00	5.30	14.11	5.5	0.560	4.1	2677	61
14.1	1.55	11	14	1.27	5.06	13.85	6.2	0.548	4.2	2683	75
15.1	0.19	147	9	0.06	51.80	8.20	1.4	0.4103	1.1	2287	16
16.1	1.72	23	20	0.87	6.52	6.02	7.7	0.329	3.1	2135	120
17.1	0.27	56	5	0.09	19.80	7.67	2.8	0.4125	2.0	2161	34
18.1	1.54	32	64	2.00	12.50	9.56	4.4	0.4467	2.2	2403	65
19.1	1.49	20	17	0.85	6.90	7.18	7.7	0.390	4.7	2148	110
20.1	1.34	12	15	1.25	4.76	12.42	7.8	0.469	3.3	2759	120
21.1	3.45	14	14	1.00	5.56	10.60	10	0.460	3.9	2526	160
22.1	1.34	13	17	1.31	4.10	6.61	11	0.359	3.7	2147	170

sample 07JG32											
spot	$^{206}\text{Pb}_c$ %	U ppm	Th ppm	Th/U	$^{206}\text{Pb}^*$ ppm	$^{207}\text{Pb}^*/^{235}\text{U}$	$\pm\%$	$^{206}\text{Pb}^*/^{238}\text{U}$	$\pm\%$	$^{206}\text{Pb}/^{238}\text{U}$ Age(Ma) $\pm 1\sigma$	
1.1	2.71	12	17	1.49	3.41	5.42	8.0	0.333	3.7	1853	59
2.1	0.20	78	117	1.49	25.7	6.84	2.6	0.3812	2.1	2082	38
3.1	5.84	21	26	1.23	6.37	4.47	15	0.330	3.5	1840	55
4.1	5.34	14	19	1.32	3.90	4.15	18	0.302	4.8	1700	72
5.1	2.62	7	16	2.34	2.12	6.41	9.1	0.344	4.8	1908	80
6.1	2.13	11	22	2.01	3.44	6.37	11	0.350	4.6	1937	76
7.1	2.92	21	21	1.02	5.82	4.91	8.8	0.318	3.2	1778	50
8.1	5.60	13	19	1.46	4.14	4.6	25	0.346	4.6	1918	74
9.1	1.49	20	42	2.09	5.71	5.18	7.1	0.321	3.2	1795	50
10.1	4.26	12	16	1.34	3.43	4.11	20	0.314	4.2	1761	63
11.1	0.71	52	32	0.62	14.3	5.00	4.3	0.3200	2.4	1790	37
12.1	1.14	63	75	1.18	21.8	6.90	3.7	0.3948	2.2	2145	41
13.1	0.31	194	223	1.15	65.3	7.12	2.2	0.3903	1.9	2124	34
14.1	6.38	27	27	1.00	7.90	4.61	15	0.319	3.3	1785	51
15.1	0.37	192	168	0.88	63.6	6.78	2.2	0.3836	2.0	2093	35
16.1	7.46	11	22	1.92	3.60	4.5	26	0.339	4.9	1881	77
17.1	1.49	38	28	0.74	9.92	4.37	8.0	0.2998	2.8	1690	42
18.1	0.60	99	58	0.58	34.3	7.34	2.8	0.3993	2.1	2166	38
19.1	0.60	24	21	0.89	7.19	5.97	5.4	0.351	3.2	1941	54
20.1	8.75	11	18	1.65	3.27	4.7	26	0.308	4.7	1729	68

Pb_c and Pb^* indicate the common and radiogenic portions, respectively. Pb^* : corrected for common ^{204}Pb using measured ^{204}Pb . All errors are 1σ . me - metamorphic zircon, ma - magmatic zircon, c - core, m - mantle, c - rim.

Table 5 Hf isotopic compositions of zircons from the xenoliths at Jiagou in the southeastern margin of the North China Craton

Sample 07FY01 (t1 = 2100Ma, t2=1800Ma)												
No.	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm(2\sigma)$	Age (Ma)	$\epsilon_{\text{Hf}(t2)}$ $\pm(2\sigma)$	$\epsilon_{\text{Hf}(t1)}$ $\pm(2\sigma)$	T_{DM1} (Ma)	$\pm(2\sigma)$	$f_{\text{Lu/Hf}}$	T_{DM2} (Ma)	$\pm(2\sigma)$
1.1	0.000058	0.000003	0.281702	0.000024	1599	2.2±0.4	9.1±0.5	2120	32	-1.00	2132	53
5.1	0.000289	0.000009	0.281711	0.000021	1814	2.6±0.4	9.4±0.4	2107	28	-1.00	2111	45
6.1	0.000391	0.000011	0.281674	0.000022	1771	1.3±0.4	8.1±0.4	2157	30	-1.00	2193	48
12.1	0.000411	0.000012	0.281686	0.000021	1700	1.7±0.4	8.5±0.4	2141	28	-1.00	2167	46
13.1	0.000180	0.000005	0.281687	0.000017	1812	1.7±0.3	8.6±0.3	2139	23	-1.00	2164	38
14.2	0.000378	0.000011	0.281681	0.000022	1886	1.5±0.4	8.3±0.4	2148	29	-1.00	2179	48
15.1	0.000699	0.000021	0.281712	0.000020	1745	2.6±0.4	9.4±0.4	2107	27	-1.00	2111	44
16.1	0.000126	0.000005	0.281737	0.000019	1849	3.5±0.4	10.3±0.4	2073	25	-1.00	2055	41
Sample 07JG12 (t1 = 2500Ma, t2=1800Ma)												
No.	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm(2\sigma)$	Age (Ma)	$\epsilon_{\text{Hf}(t2)}$ $\pm(2\sigma)$	$\epsilon_{\text{Hf}(t1)}$ $\pm(2\sigma)$	T_{DM1} (Ma)	$\pm(2\sigma)$	$f_{\text{Lu/Hf}}$	T_{DM2} (Ma)	$\pm(2\sigma)$
2.1	0.000640	0.000024	0.281338	0.000031	2291	-10.7±0.6	5.3±0.6	2607	41	-1.00	2675	66
3.1	0.000624	0.000023	0.281473	0.000019	1796	-5.9±0.3	10.1±0.3	2426	25	-1.00	2380	41
4.1	0.005753	0.000231	0.281226	0.000042	2363	-14.9±0.7	1.0±0.7	2770	56	-0.99	2938	91
5.1	0.000333	0.000012	0.281434	0.000020	1786	-7.3±0.4	8.8±0.4	2478	27	-1.00	2464	45
8.1	0.000332	0.000012	0.281428	0.000020	1766	-7.5±0.4	8.5±0.4	2486	26	-1.00	2478	43
9.1	0.005874	0.000266	0.281510	0.000022	no	-4.9±0.4	11.0±0.4	2392	29	-0.99	2325	48
9.2	0.000648	0.000025	0.281347	0.000021	2187	-10.4±0.4	5.6±0.3	2595	28	-1.00	2656	46
12.1	0.000401	0.000014	0.281478	0.000027	1819	-5.7±0.5	10.3±0.4	2420	36	-1.00	2369	59
13.2	0.002150	0.000086	0.281294	0.000021	2221	-12.3±0.4	3.7±0.4	2669	29	-1.00	2776	47
14.1	0.003887	0.000167	0.281301	0.000027	1783	-12.2±0.5	3.8±0.4	2666	36	-0.99	2770	58
15.1	0.010489	0.000463	0.281499	0.000022	1786	-5.5±0.4	10.3±0.4	2419	30	-0.99	2370	49
Sample 07JG14 (t1 = 2500Ma, t2=1800Ma)												
No.	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm(2\sigma)$	Age (Ma)	$\epsilon_{\text{Hf}(t2)}$ $\pm(2\sigma)$	$\epsilon_{\text{Hf}(t1)}$ $\pm(2\sigma)$	T_{DM1} (Ma)	$\pm(2\sigma)$	$f_{\text{Lu/Hf}}$	T_{DM2} (Ma)	$\pm(2\sigma)$
3.1	0.029836	0.001108	0.281582	0.000025	1822	-3.4±0.5	12.1±0.5	2346	35	-0.97	2256	55
3.2	0.026136	0.000979	0.281555	0.000023	1717	-4.2±0.4	11.4±0.4	2376	31	-0.97	2302	50
4.1	0.006538	0.000263	0.281449	0.000024	1787	-7.0±0.5	8.9±0.5	2474	33	-0.99	2458	53
5.1	0.011601	0.000535	0.281351	0.000026	2563	-10.9±0.5	4.9±0.4	2624	35	-0.98	2700	56
5.2	0.013299	0.000517	0.281487	0.000034	387	-6.0±0.6	9.8±0.5	2439	46	-0.98	2402	74
6.1	0.008101	0.000298	0.281296	0.000022	1830	-12.5±0.4	3.4±0.4	2682	29	-1.0	2795	47
7.1	0.006782	0.000302	0.281504	0.000028	2400	-5.1±0.5	10.7±0.4	2403	38	-1.0	2343	62
8.1	0.004106	0.000163	0.281383	0.000022	1901	-9.3±0.4	6.7±0.4	2556	30	-1.0	2591	49
9.2	0.029982	0.001278	0.282075	0.000033	176	14.0±0.6	29.4±0.5	1670	46	-0.96	1190	72
10.1	0.005631	0.000210	0.281376	0.000023	1695	-9.6±0.4	6.3±0.4	2569	30	-0.99	2612	49
Sample 07JG30 (t=390Ma)												
No.	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm(2\sigma)$	Age (Ma)	$\epsilon_{\text{Hf}(t)}$ $\pm(2\sigma)$	T_{DM1} (Ma)	$\pm(2\sigma)$	$f_{\text{Lu/Hf}}$			

1.1	0.029215	0.001231	0.282631	0.000029	393	3.3	0.5	886	42	-0.96
3.1	0.031191	0.001344	0.282678	0.000026	384	4.9	0.5	821	37	-0.96
5.1	0.041135	0.001808	0.282732	0.000035	397	6.7	0.6	754	50	-0.95
7.1	0.023913	0.001121	0.282702	0.000028	410	5.8	0.5	782	40	-0.97
8.1	0.036162	0.001620	0.282609	0.000029	389	2.4	0.5	926	42	-0.95
9.1	0.027403	0.001182	0.282710	0.000027	391	6.1	0.5	772	38	-0.96
10.1	0.033455	0.001502	0.282553	0.000044	425	0.5	0.8	1003	62	-0.95

Sample 07JG32 (t1 = 2100Ma, t2=1800Ma)

No.	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm(2\sigma)$	Age (Ma)	$\epsilon_{\text{Hf}(t2)}$ $\pm(2\sigma)$	$\epsilon_{\text{Hf}(t1)}$ $\pm(2\sigma)$	T_{DM1} (Ma)	$\pm(2\sigma)$	$f_{\text{Lu/Hf}}$	T_{DM2} (Ma)	$\pm(2\sigma)$
2.1	0.052867	0.002158	0.281424	0.000043	2082	-10.2±0.7	-3.9±0.8	2636	61	-0.94	2927	94
5.1	0.000469	0.000017	0.281523	0.000021	1908	-4.1±0.4	2.7±0.4	2359	28	-1.00	2523	45
6.1	0.000973	0.000036	0.281531	0.000030	1937	-3.9±0.5	3.0±0.6	2350	41	-1.00	2508	66
7.1	0.000270	0.000009	0.281495	0.000031	1778	-5.1±0.6	1.7±0.5	2396	42	-1.00	2584	68
9.1	0.000478	0.000018	0.281542	0.000031	1795	-3.5±0.5	3.4±0.5	2335	41	-1.00	2484	67
10.1	0.001333	0.000049	0.281505	0.000036	1761	-4.8±0.6	2.0±0.5	2386	49	-1.00	2567	79
12.1	0.026872	0.001162	0.281350	0.000038	2145	-11.7±0.7	2.0±0.6	2668	52	-0.97	3001	83
13.1	0.042025	0.001588	0.281486	0.000031	2124	-7.3±0.6	-0.8±0.5	2510	43	-0.95	2742	67
15.1	0.026188	0.001231	0.281421	0.000038	2093	-9.2±0.7	-2.6±0.7	2575	53	-0.96	2852	84
17.1	0.000632	0.000020	0.281523	0.000035	1690	-4.1±0.7	2.7±0.6	2360	46	-1.00	2524	76
18.1	0.003476	0.000163	0.281460	0.000033	2166	-6.6±0.6	0.3±0.5	2454	44	-1.00	2675	72
19.1	0.000516	0.000022	0.281554	0.000029	1941	-3.0±0.5	3.8±0.5	2319	38	-1.00	2457	62

Sample 07JG34 (t1 = 2500Ma, t2=1800Ma)

No.	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm(2\sigma)$	Age (Ma)	$\epsilon_{\text{Hf}(t2)}$ $\pm(2\sigma)$	$\epsilon_{\text{Hf}(t1)}$ $\pm(2\sigma)$	T_{DM1} (Ma)	$\pm(2\sigma)$	$f_{\text{Lu/Hf}}$	T_{DM2} (Ma)	$\pm(2\sigma)$
1.1	0.006343	0.000234	0.281261	0.000021	2458	-13.7±0.4	2.2±0.4	2724	28	-0.99	2864	45
3.1	0.012047	0.000446	0.281328	0.000019	2485	-11.6±0.4	4.2±0.4	2649	26	-0.99	2740	42
4.1	0.006986	0.000263	0.281326	0.000022	2203	-11.4±0.4	4.5±0.4	2640	29	-0.99	2727	47
5.1	0.008959	0.000333	0.281288	0.000018	2567	-12.8±0.3	3.0±0.3	2695	25	-0.99	2815	40
7.1	0.004794	0.000180	0.281332	0.000021	2389	-11.1±0.4	4.8±0.4	2626	28	-0.99	2704	45
11.1	0.007860	0.000297	0.281321	0.000019	2540	-11.7±0.4	4.2±0.4	2649	26	-0.99	2741	42
12.1	0.003820	0.000146	0.281282	0.000019	2466	-12.8±0.4	3.1±0.3	2690	25	-1.00	2808	41
14.1	0.005750	0.000223	0.281330	0.000024	2683	-11.2±0.5	4.7±0.5	2631	33	-0.99	2713	53
15.1	0.049588	0.001917	0.281391	0.000023	2287	-11.1±0.4	4.0±0.4	2665	32	-0.94	2756	50
17.1	0.024919	0.000946	0.281368	0.000018	2161	-10.7±0.3	4.8±0.3	2628	25	-0.97	2704	39
18.1	0.004660	0.000206	0.281234	0.000018	2403	-14.60±4	1.3±0.3	2758	25	-0.99	2919	40
19.2	0.014326	0.000637	0.281290	0.000025	2148	-13.1±0.5	2.6±0.5	2713	34	-0.98	2842	55
20.1	0.005367	0.000214	0.281312	0.000027	2759	-11.9±0.5	4.1±0.5	2655	36	-0.99	2752	58
22.1	0.006858	0.000275	0.281317	0.000019	2147	-11.8±0.4	4.0±0.4	2653	26	-0.99	2747	42

Figure 1

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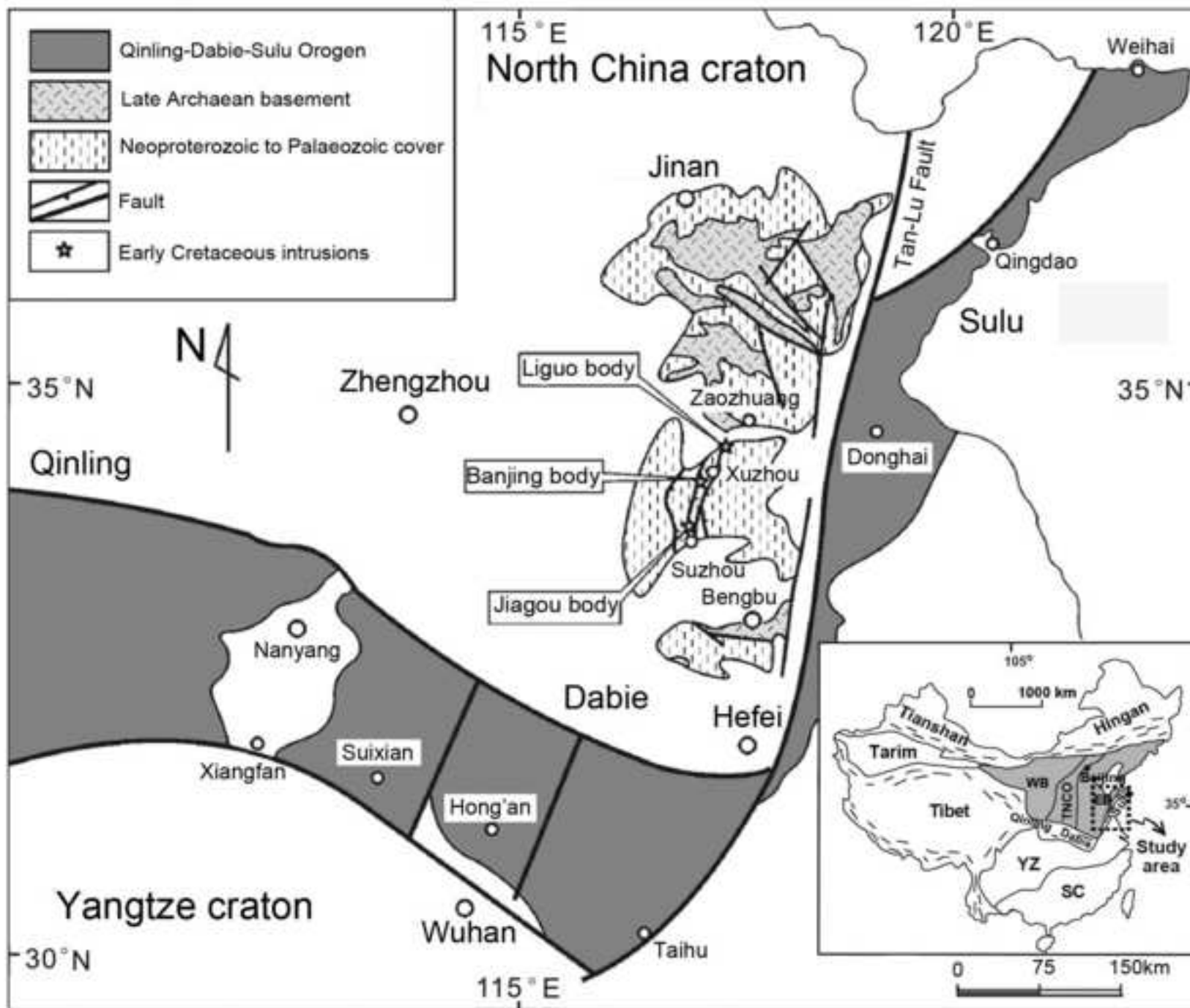


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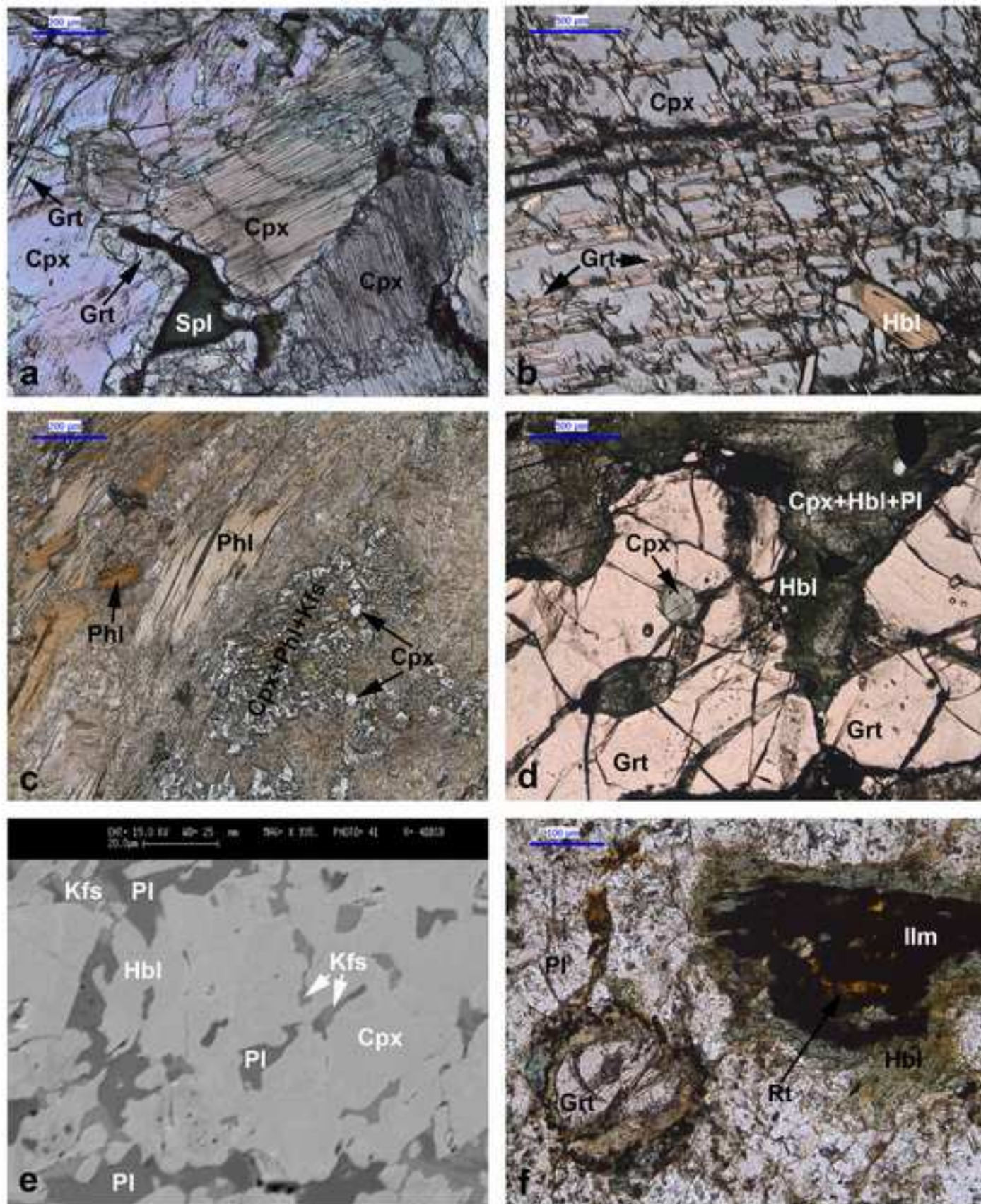


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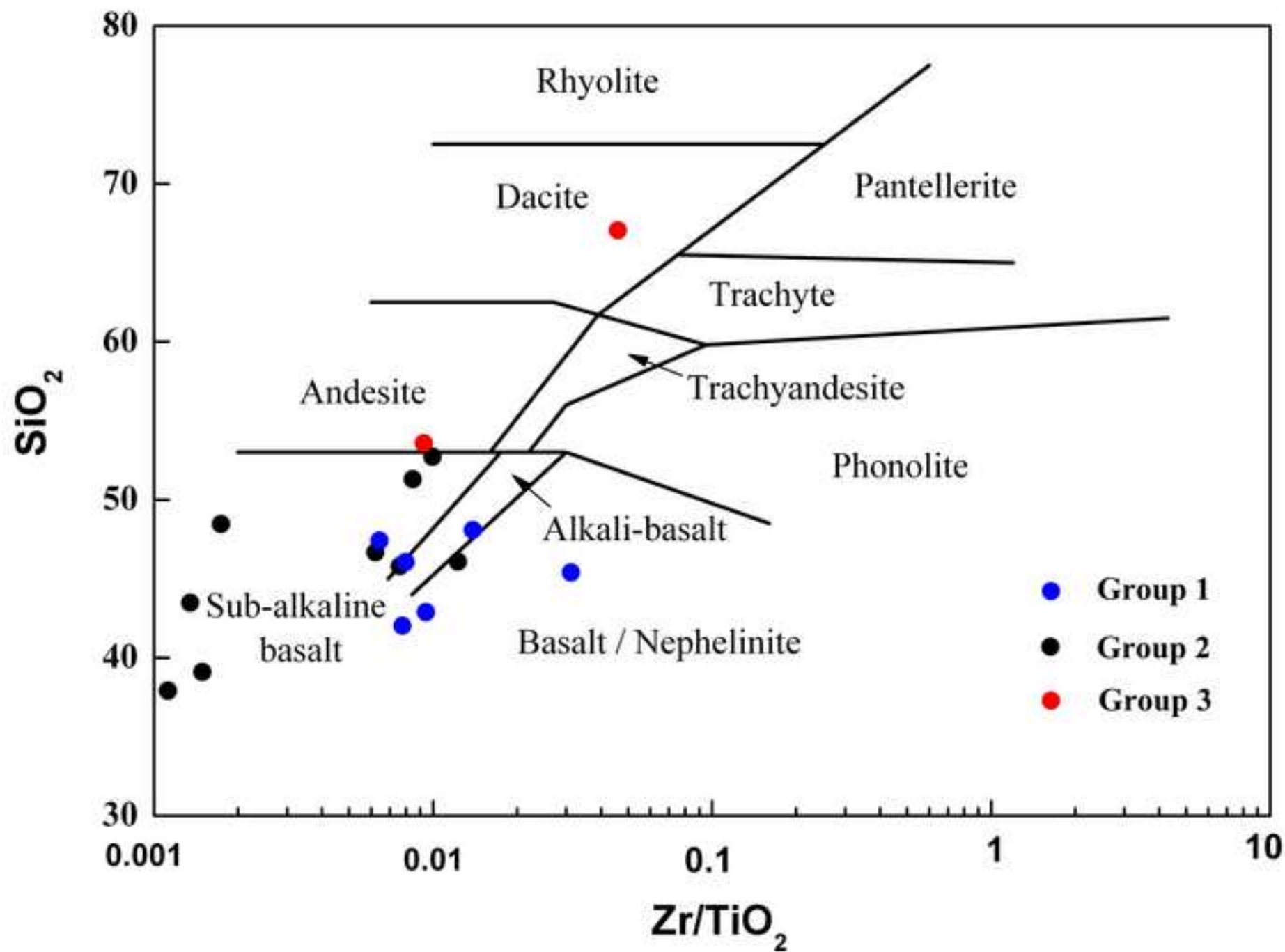


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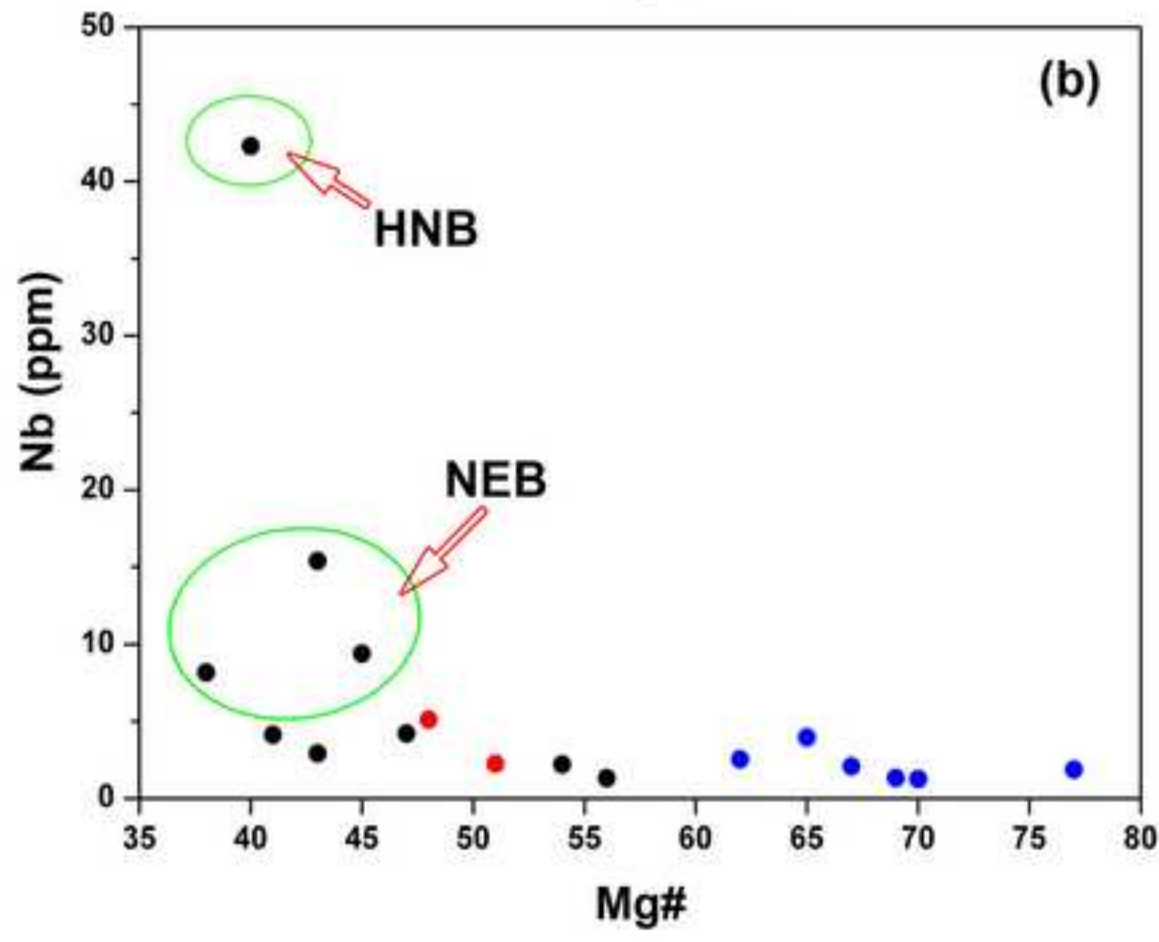
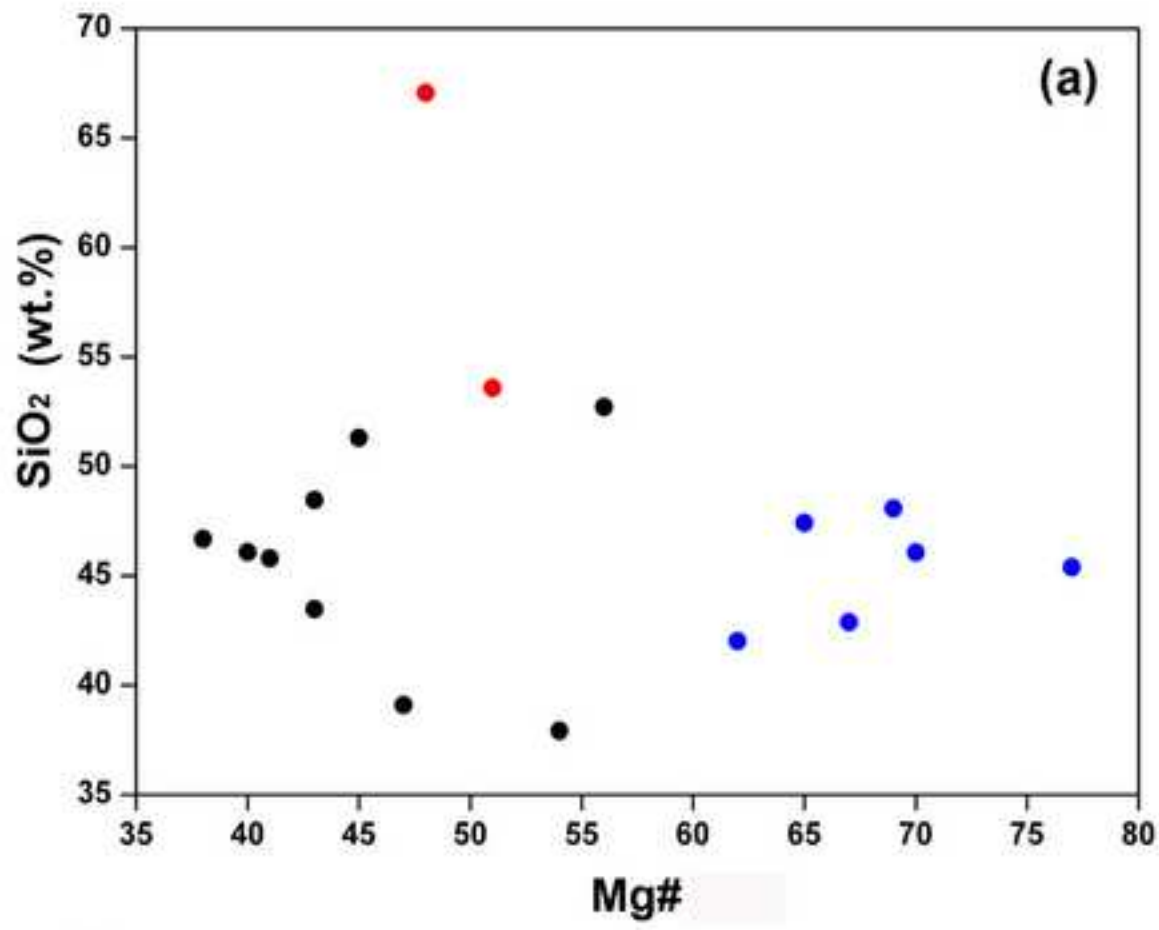


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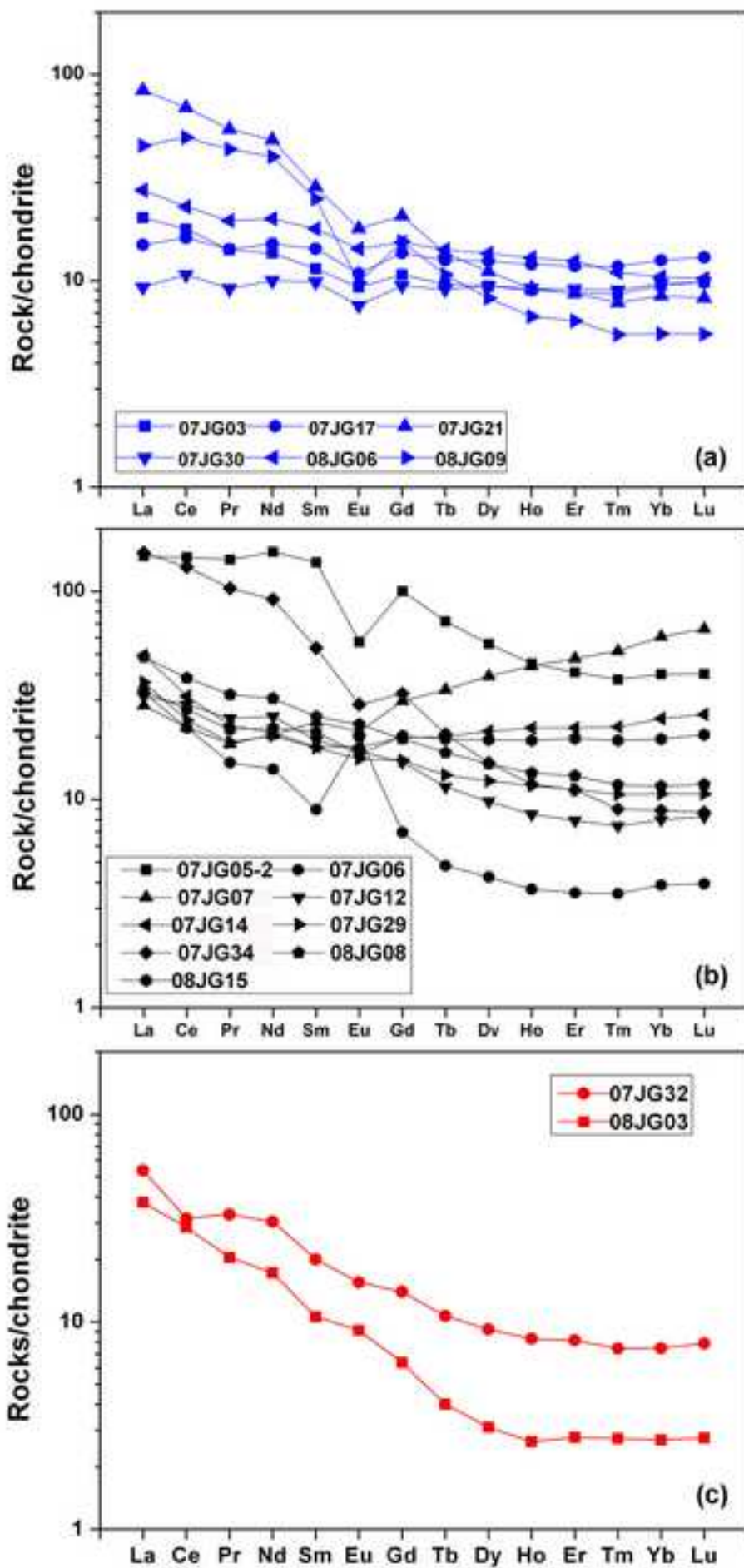


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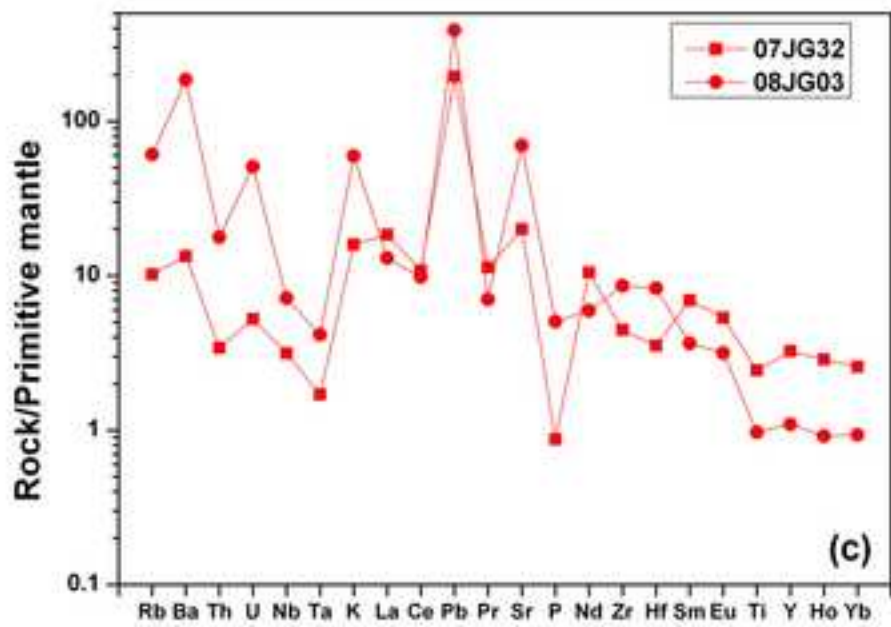
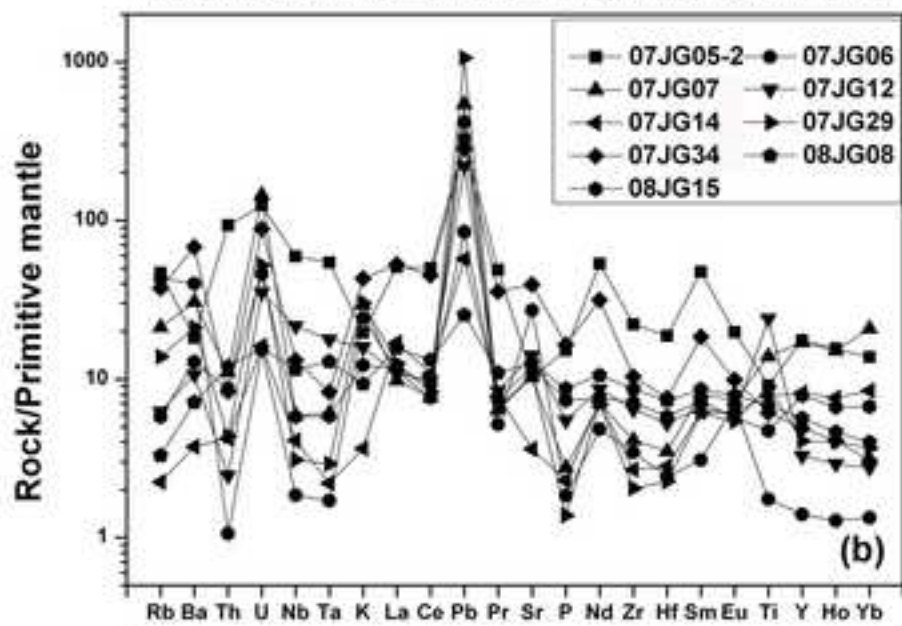
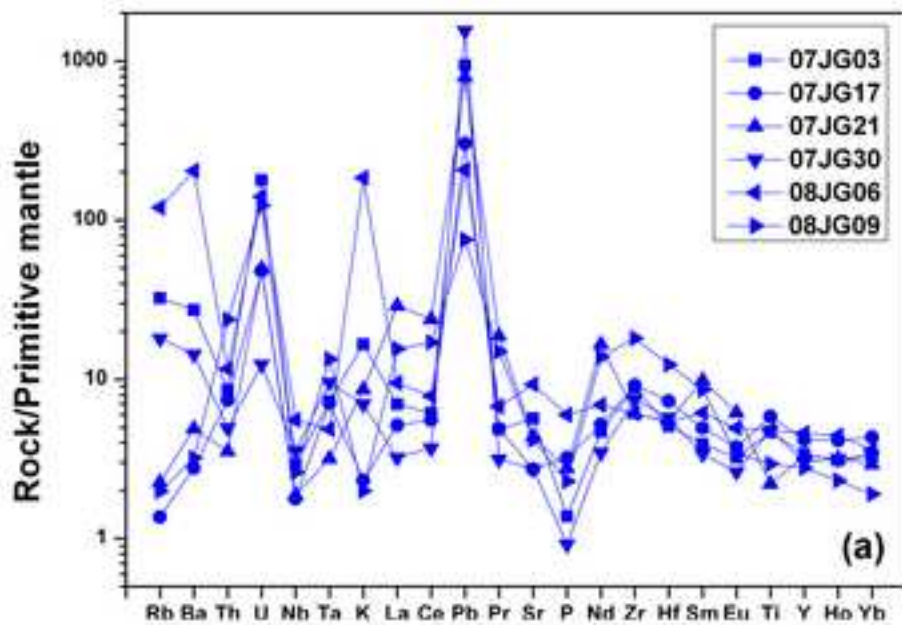


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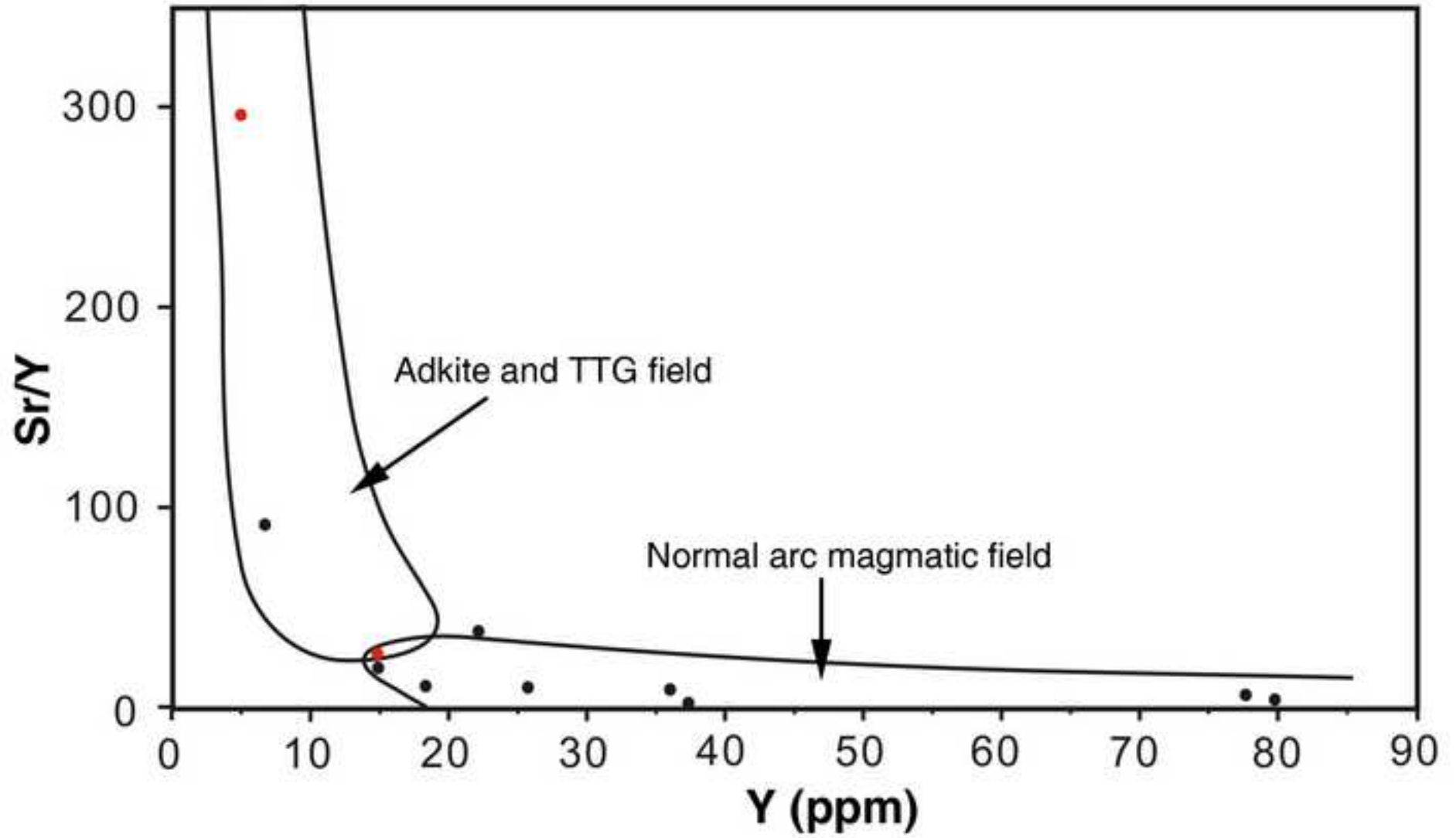


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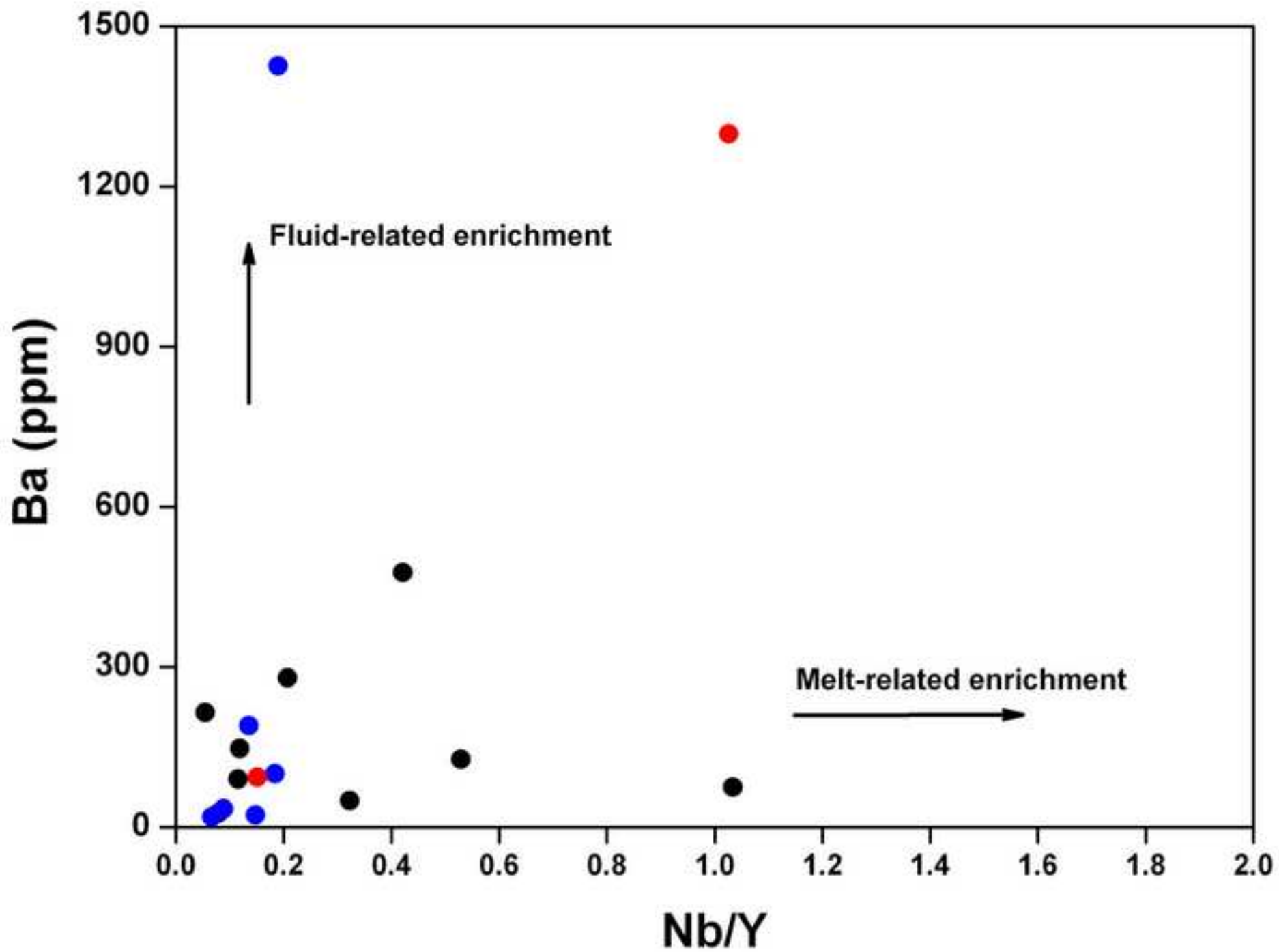


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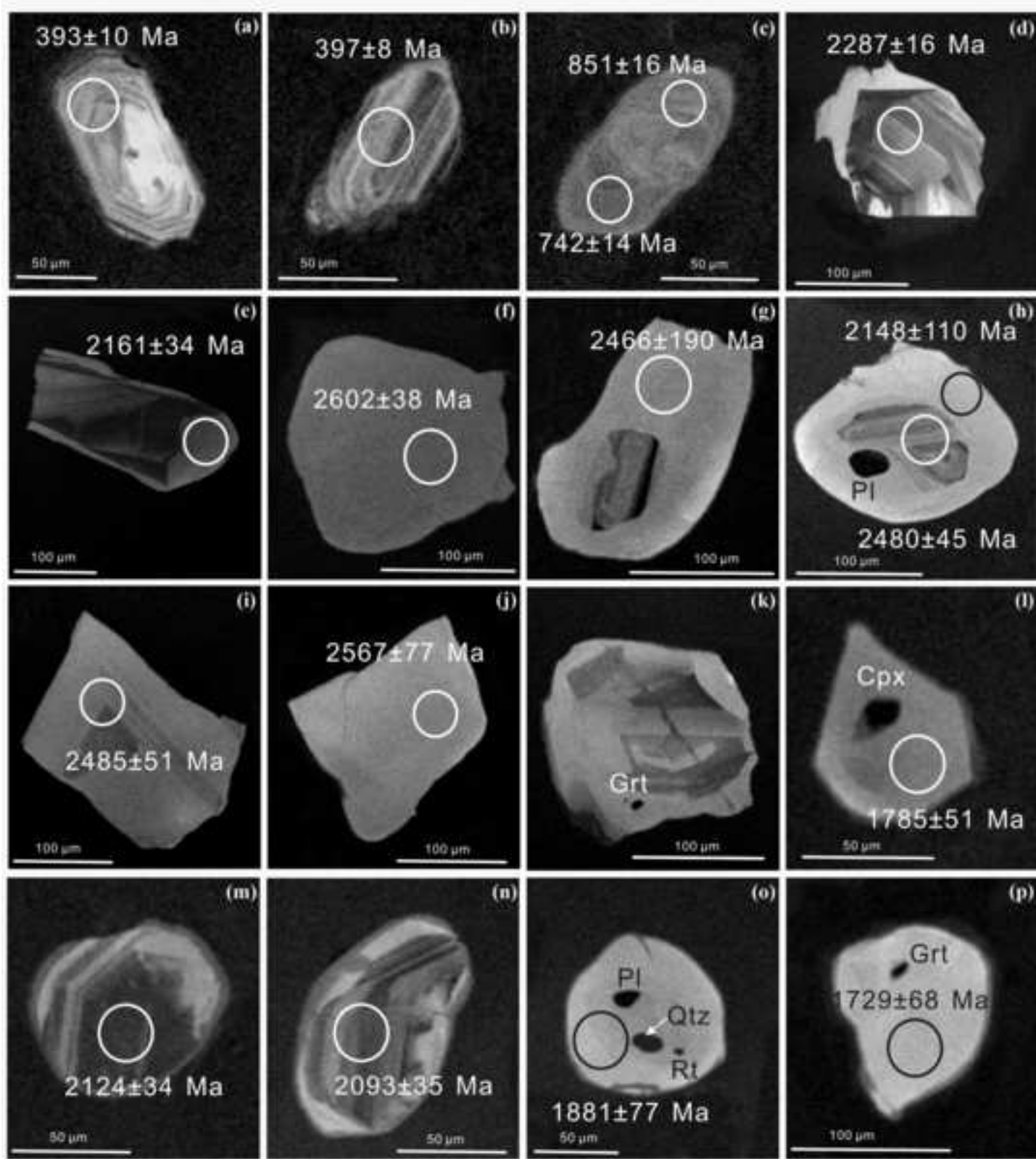


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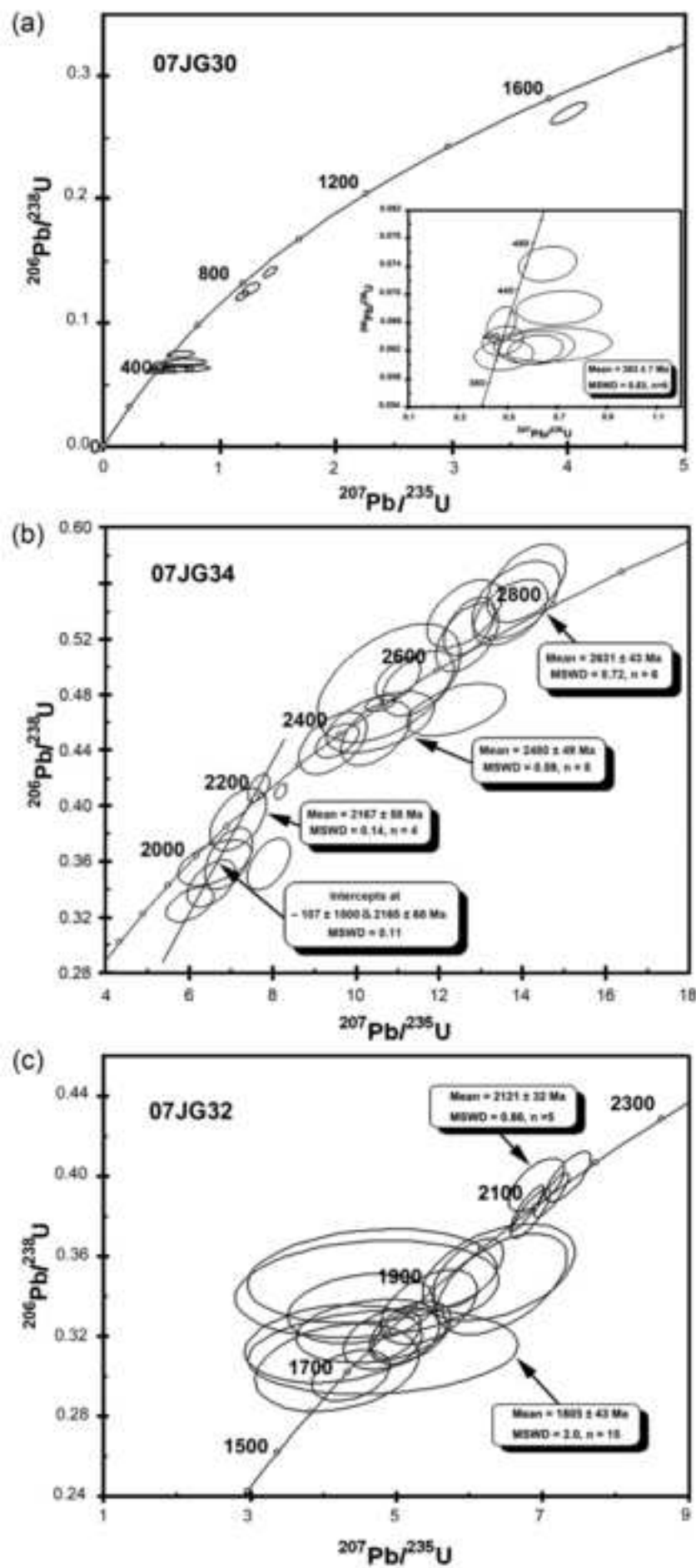


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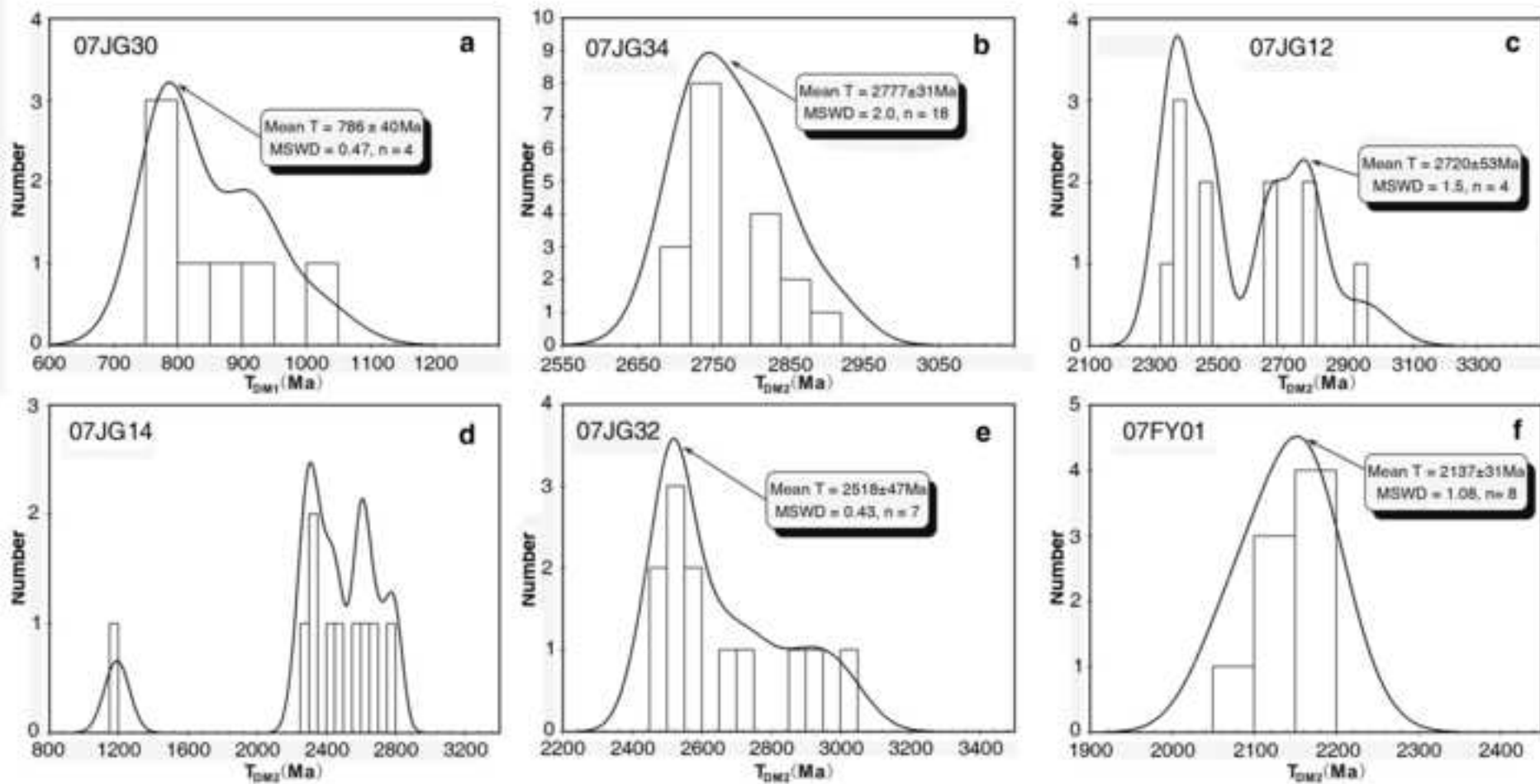


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