Updating the Geologic Barcodes for South China: Discovery of Late Archean Banded Iron Formations in the Yangtze Craton

Hui Ye¹, Chang-Zhi Wu¹, Tao Yang¹, M. Santosh²,³, Xi-Zhu Yao¹, Bing-Fei Gao¹, Xiao-Lei Wang¹ & Weiqiang Li¹

Banded iron formations (BIFs) in Archean cratons provide important “geologic barcodes” for the global correlation of Precambrian sedimentary records. Here we report the first finding of late Archean BIFs from the Yangtze Craton, one of largest Precambrian blocks in East Asia with an evolutionary history of over 3.3 Ga. The Yingshan iron deposit at the northeastern margin of the Yangtze Craton, displays typical features of BIF, including: (i) alternating Si-rich and Fe-rich bands at sub-mm to meter scales; (ii) high SiO₂ + Fe₂O₃ total contents (average 90.6 wt.% and Fe/Ti ratios (average 489); (iii) relative enrichment of heavy rare earth elements and positive Eu anomalies (average 1.42); (iv) and sedimentary Fe isotope compositions (δ⁵⁶Fe IRMM-014 as low as −0.36‰). The depositional age of the BIF is constrained at ~2464 ± 24 Ma based on U-Pb dating of zircon grains from a migmatite sample of a volcanic protolith that conformably overlaid the Yingshan BIF. The BIF was intruded by Neoproterozoic (805.9 ± 4.7 Ma) granitoids that are unique in the Yangtze Craton but absent in the North China Craton to the north. The discovery of the Yingshan BIF provides new constraints for the tectonic evolution of the Yangtze Craton and has important implications in the reconstruction of Pre-Nuna/Columbia supercontinent configurations.

The Archean-Paleoproterozoic boundary was a critical period in Earth’s history with a series of significant changes in atmosphere, lithosphere and hydrosphere.¹² The formation of BIFs also reached its climax during this period.³ The BIFs worldwide are important repositories of the early Earth’s major environmental transitions and biological innovations.³⁻⁴. They are considered to have formed in distinct tectonic settings. For example, the Superior-type BIFs are thought to have developed on continental shelf below storm wave base, and granular iron-formations such as the Gunflint-type BIFs developed in shallow-water, high-energy environments.³ In contrast, the Neoproterozoic iron formations are commonly associated with continental rift-basins.⁵. Records of BIFs therefore provide important constraints on the tectonic histories of Earth’s ancient crustal blocks. The Yangtze Craton (Fig. 1a) in South China is one of the largest Precambrian blocks in East Asia, with an Archean-Paleoproterozoic basement that dates back to 3.3 Ga.⁶ Although a number of Neoproterozoic iron formations occur around the southeastern margin of the Yangtze Craton (Fig. 1a), so far there has been no report on Archean-Paleoproterozoic BIFs from this craton, which is in sharp contrast with the North China Craton to the north where Archean-Paleoproterozoic BIFs are abundant.¹¹⁻¹²

Whether the absence of BIFs in the Yangtze Craton is a preservational issue or it reflects the lack of a favorable tectonic environment for Archean-Paleoproterozoic BIF deposition is a crucial question, particularly in understanding the evolution of this craton and its position with respect to pre-Nuna/Columbia supercontinents.⁶⁻⁷ Occurrence of magnetite quartzite has been mentioned from the Archean-Paleoproterozoic Feidong Group¹³ and the Archean Yudongzi group¹⁴ around the boundary between the North China Craton and the Yangtze Craton (Fig. 1a), but their protolith, age and tectonic affinity remain elusive. In this contribution, we place geochronological and geochemical constraints on the magnetite quartzite from the Paleoproterozoic Feidong Group,⁶⁻¹⁷. Our data provide unequivocal evidence for ca. 2.46 Ga BIFs from the Yangtze Craton, confirming the first discovery of late Archean BIFs in this craton.

¹State key laboratory for Mineral Deposits Research, School of Earth Sciences and Engineering, Nanjing University, Nanjing, 210093, China. ²School of Earth Sciences and Resources, China University of Geosciences, Beijing, 100083, China. ³Department of Earth Sciences, University of Adelaide, Adelaide, SA 5005, Australia. Correspondence and requests for materials should be addressed to C.-Z.W. (email: wucz@nju.edu.cn) or W.L. (email: liweiqiang@nju.edu.cn)
Geological setting, samples, and analyses

The Yangtze Craton, separated from the North China Craton by the Qinling-Sulu-Dabie orogen to the north, contains a widespread Archean basement. Much of the basement is covered by weakly metamorphosed Neoproterozoic and Phanerozoic strata, with limited Archean-Paleoproterozoic outcrops restricted to the northern part of the craton (e.g., Kongling complex, ~3.3 Ga; Fig. 1a).1–8

The Tan-Lu fault, the largest fault system in East Asia, defines the eastern boundary between the North China Craton and the Yangtze Craton (Fig. 1a,b). The fault sinistrally offsets the Sulu-Dabie orogen by a maximum apparent displacement of ~400 km, exposing a NEE-trending belt of the basement of the Yangtze Craton, locally termed as the Zhangbaling metamorphic belt.17,18 This belt is composed of the greenschist-facies Neoproterozoic Zhangbaling Group in the north and the amphibolite-facies Paleoproterozoic Feidong Group in the south.17 The Feidong Group contains the Fuchashan, the Tongshan and the Qiaotouji formations from bottom to top. The Tongshan Formation is a metamorphosed sedimentary-volcanic succession and contains several thin-bedded magnetite quartzite layers that extend over 30 km (Fig. 1c).

The Yingshan iron deposit is hosted in the Tongshan Formation of the Feidong Group. The ore bodies occur as Fe-rich layers that typically extend for over 1 km with an average thickness of over 10 m and Fe grade of 27 wt% (Supplementary Fig. S1). The ores consist of banded quartz-magnetite and garnet-amphibolite-quartz-magnetite (Supplementary Fig. S1). The layered ore bodies are inter-bedded with amphibolite, biotite schist and migmatite, and have been intruded by granitoids (Supplementary Fig. S1). The rocks locally underwent amphibolite-grade metamorphism, structural deformation and hydrothermal alteration associated with the Tan-Lu fault system (Supplementary Fig. S1) during the Triassic.17

Figure 1. Location (a) and geological map (b) of the Yingshan iron deposit. Stratigraphic column of the Yingshan iron deposit (c) showing sampling point (arrow) and zircon U-Pb age (star). Representative photos showing macroband (d), mesoband (e), and microband (f) textures of the iron ores from Yingshan, that are similar to those of typical BIF bands elsewhere in the world such as the Dales Gorge Member, Hamersley basin. Acronyms in Fig. 1a: NCC-North China Craton; YDZ-Yudongzi group; DL-Douling complex; KL-Kongling complex; HTL-Huangtuling granulites; XY-Xinyu iron formation (Neoproterozoic); FL-Fulu iron formation (Neoproterozoic). The geological maps were generated using CorelDRAW Graphcs Suite 2017, http://www.coreldraw.com/cn/free-trials/?topNav=cn.
Twenty-six samples were collected from the Yingshan deposit (Fig. 1c), including iron ores from three layers of orebodies, as well as the host rocks including migmatite, biotite schist and amphibolites. Bulk rocks were pulverized and analyzed for major and trace elements using XRF and ICP-MS, respectively. Iron isotope compositions of the bulk rocks were measured by solution-nebulization MC-ICP-MS after ion-exchange purification. Zircon grains were separated from a leucosome (YS-10) sample from the upper wall of the ore bodies, and a granitoid (YS-01) that intruded into the iron ore bodies (Fig. 1c). The grains were analyzed using LA-ICP-MS for U-Pb dating. Details of the results are provided in the Supplementary Information.

Results and Discussion

Protolith of the iron ores. In spite of the deformation and metamorphism, the iron ores from Yingshan show characteristic banded texture with alternating silica-rich and iron-rich layers. The banded textures are obvious in meter scale within the layered ore bodies (Fig. 1d), at centimeter scale in hand specimens (Fig. 1e), and at sub-millimeter scale within iron-rich bands under the microscope (Fig. 1f). These are consistent with the classic macroband, mesoband, and microband textures of typical BIFs as for example in the case of the Dales Gorge Member in the Archean-Proterozoic Hamersley basin, western Australia.

Major elements of the banded iron ores (Supplementary Table S1) are dominated by SiO₂ and Fe₂O₃Total (79.4–95.8 wt%), a feature characteristic for oceanic chemical deposits, and consistent with the average composition of BIFs as summarized from 214 BIFs worldwide. The variable contents of Al₂O₃ (1.66–9.81 wt%) and TiO₂ (0.04–0.86 wt%) likely reflect syndepositional volcanic inputs. Additionally, ore samples with low Al₂O₃ and TiO₂ contents display rare earth element plus Y (REE + Y) patterns of relative enrichment in heavy REE (average HREE/LREE = 2.28) and positive Eu anomalies (average ΔEu = 1.42) (Supplementary Table S4). The REE + Y patterns of the iron ores are similar to those of the classic BIFs formed during the Archean-Proterozoic transition (Fig. 2a).

Iron isotope analyses for all the 23 samples (Fig. 2b, Supplementary Table S1) revealed a variation of 0.83‰ in δ⁵⁶Fe. The wall-rock samples, including leucosome of migmatite, amphibolite and granitoid intrusion, have positive δ⁵⁶Fe values (0.00–0.47‰). In contrast, ore samples are generally enriched in light Fe isotopes, with δ⁵⁶Fe ranging from −0.36‰ to 0.04‰ with an exception of one sample that has δ⁵⁶Fe of +0.32‰ (Fig. 2b). Magmatic rocks in general possess a homogeneous Fe isotope composition, except for highly evolved (SiO₂ > 70 wt%) granitoids and leucosome of migmatites that have high δ⁵⁶Fe values. The variable and low δ⁵⁶Fe values of the iron ores therefore exclude an igneous parentage, and instead reflect Fe isotope variation in the protolith.

The complexities in bulk rock Fe isotope compositions caused by mixing with detrital components as well as metamorphism and hydrothermal alteration can be assessed using a plot of δ⁵⁶Fe versus Fe/Ti. This approach...
has proven to be effective in resolving the protolith of the earliest BIF from the highly metamorphosed 3.8 billion-year-old rocks of Isua, Greenland (Fig. 2b). There is a general negative correlation between the Fe/Ti atomic ratio and δ56Fe values for the iron ores, reflecting mixing between an igneous Fe end member and an end member that is characterized by a high (>1200) Fe/Ti ratio and low δ56Fe (Fig. 2b). Low δ56Fe values are one of the hallmarks of BIFs that are absent in other bulk geological samples, and are considered to reflect redox processes during deposition of Fe in the water column and subsequent diagenesis in soft sediments.

Depositional age and tectonic affinity of the Yingshan BIF. Multiple lines of evidence from texture, chemical compositions and Fe isotope compositions presented above collectively indicate that the protolith for the iron ores from the Yingshan deposit was a banded iron formation, which is referred to as the Yingshan BIF hereafter. The depositional age of the Yingshan BIF is constrained by U-Pb geochronology of zircon grains from the leucosome (YS-10) of a migmatite. The protolith of the migmatite is thought to be a volcanic rock interbedded within the Tongshan volcano-sedimentary sequence and conformably overlying the Yingshan BIF mineralization. Association with volcanic rocks is a common feature for Archean BIFs, particularly for Algoma-type BIFs, and the age of zircons from the volcanic layers interbedded in BIFs have been widely used to constrain the depositional ages of the BIFs. The zircons from the migmatite leucosome are euhedral with a size of 40–170 μm, and cathodoluminescence (CL) imaging reveals common core-rim textures (Fig. 3). The zircon cores show bright oscillatory CL zoning, and have high Th/U ratio (Th/U = 0.45–1.16), as well as HREE-enriched patterns with positive Ce and Sm anomalies (Fig. 3; Supplementary Fig. S2), all of which are typical of magmatic zircons. The rims of the zircon in contrast are dark-gray in CL, have low Th/U values (Th/U = 0.04–0.07) and flat REE patterns without significant Ce and Sm anomalies (Fig. 3; Supplementary Fig. S2), suggesting a metamorphic origin. The concordant U-Pb ages for the magmatic zircon cores tightly cluster at 2464 ± 24 Ma (207Pb/206Pb age, n = 14, MSWD = 0.04). The tight distribution of U-Pb ages as well as the irregular shapes of the zircon cores supports the idea that the protolith of the migmatite is a volcanic rock rather than a detrital sediment. Except for an inherited zircon with a concordant 207Pb/206Pb age of 2544 Ma (Fig. 3, Supplementary Table S5), all the U-Pb analyses of cores and rims of zircon grains from YS-10 define a Discordia with an upper intercept age of 2465 ± 11 Ma and a lower intercept age of 265 ± 27 Ma (n = 67, MSWD = 7.1; Fig. 3). The upper intercept age of the Discordia is consistent with the concordant age (2464 ± 24 Ma) from the zircon cores, representing the age of the volcanic protolith of the migmatite, whereas the lower intercept age reflects an early Mesozoic thermal event that produced the metamorphic rims of the zircon cores. Because of the conformable relationship between the volcanic protolith of the migmatite and the iron formation (Fig. 1c; Supplementary Fig. S1), the Yingshan BIF is interpreted to be approximately coeval with the volcanic protolith of the migmatite corresponding to late Archean-early Paleoproterozoic (2464 ± 24 Ma).
Because the Yingshan iron deposit is located within a major fault zone between the Yangtze Craton and the North China Craton, it is crucial to ascertain the tectonic setting of the BIF protolith. The Yangtze Craton and the North China Craton did not collide until the Triassic, and the two cratons have distinct Precambrian evolution histories (Fig. 4). The North China Craton is characterized by widespread Archean magmatism that peaked at ~2.5 Ga and very minor Neoproterozoic magmatic activities in the form of mafic dykes. In contrast, the Yangtze Craton is characterized by widespread Neoproterozoic magmatism and ca. 2.0 Ga magmatic events and metamorphism. The Yingshan iron deposit was intruded by a granitoid pluton (Fig. 1c). Zircon grains from the granitoid (YS-01) are dark gray to gray in color, euhedral to subhedral with a size of 25–150 μm and aspect ratio of 0.5–6.0 (Fig. 3). These zircons show bright oscillatory zoning in CL imaging revealing their magmatic origin. U-Pb ages of these magmatic zircons from YS-01 are concordant and yield a weighted mean 206Pb/238U age of 805.9 ± 4.7 Ma (n = 18, MSWD = 0.36). The age of the granitoid intrusion is consistent with the magmatism along the northern margin and elsewhere within the Yangtze Craton but is conspicuously absent in the southern margin of the North China Craton. The age data suggest that the Yingshan iron ore bodies were intruded by Neoproterozoic (805.9 ± 4.7 Ma) granitoid prior to the collision between the Yangtze Craton and the North China Craton, thus confirming that the Yingshan BIF belongs to the Yangtze Craton (Fig. 4).

**Geological Implications**

Based on the geological, geochemical and geochronological evidence presented above, the Yingshan iron deposit is identified as a Neoarchean-Paleoproterozoic banded iron formation and provides a first case of BIF mineralization of such age in the Yangtze Craton. The discovery of the Yingshan BIF indicates that the northern part of the Yangtze Craton was in a shallow marine to continental shelf environment during the late Archean. Occurrence of the late Archean BIF on the northern margin of the Yangtze Craton is in contrast with the linear distribution of Neoproterozoic BIFs in the southern margin of the Yangtze Craton that are indicative of rifting during the Neoproterozoic (Fig. 1a). Such contrast in BIF occurrence seems to reflect a tectonic polarity for the Yangtze Craton, which is interestingly perpendicular to the northern boundary with the North China Craton and the southern boundary with the Cathaysia Block. This feature might have important implications on the patterns of amalgamation and breakup of continents and direction of craton drifting, at least for the case of the Yangtze Craton. The discovery of the Yingshan BIF also places further constraints for understanding the tectono-magmatic event in the Yangtze Craton which also correlates with the global peak in magmatism during the Archean to Paleoproterozoic transition (ca. 2.45 Ga).

Figure 4. Comparison of Precambrian detrital zircon U-Pb age spectra between the Yangtze Craton and the North China Craton (modified after 30). The U-Pb ages of zircons from the Yingshan iron deposit in this study are represented by stars.
Banded iron formations are considered as important “geologic barcodes” for the reconstruction of supercontinents in Earth’s deep time. For example, the similarity between Archean-Paleoproterozoic BIF records in the Pilbara Craton in Western Australia and the Kaapvaal Craton in South Africa lays the foundation for the idea that these two cratons were once part of a supercontinent (the Vaalbara) or the same sedimentary basin. We note remarkable similarity in the tectonic-sedimentary history between the Yangtze Craton and the Sao Francisco Craton of South America, including a 3.2–2.9 Ga basement of TTG, ~2.7 Ga high-K granitoid magmatic episode, ~2.5 Ga unconformity, ~2.46 Ga BIF (Yingshan and Caue BIFs), 2.4–2.0 Ga sedimentary cover (Fig. 5). Discovery of the Yingshan BIF therefore, provides additional geological constraints from the Yangtze Craton for further evaluation of the Pre-Nuna/Columbia supercontinents.

**Methods**

**Sample preparation.** All 23 wall/ore samples collected from the Yingshan iron deposit were cut to remove weathered surfaces. These relatively fresh samples were cleaned, dried and pulverized for compositional analysis. Zircon grains from the leucosome (YS-10) and granitoid (YS-01) sample were separated using standard crushing, heavy liquid, magnetite separation, and hand-picking techniques, then mounted in epoxy resin and polished for U-Pb isotope analysis. All analyses were done in the state key laboratory for mineral deposits research, Nanjing University.

**Whole-rock major element and trace element analyses.** Whole-rock major elements were analyzed using an ARL9800XP and X-ray fluorescence spectrometer (XRF), which gives the analytical precision better than 2% for all major elements. Whole-rock REEs abundances were measured using a Finnigan Element II ICP-MS and gives precision better than 10% for most REEs. Major element and selected trace element (REE + Y) results are provided in Supplementary Tables S1 and S4.

**Whole-rock iron isotope analysis.** Approximately 10 to 150 mg bulk-rock powder for each sample was digested in a 2:1:1 mixture of concentrated HCl-HNO₃-HF in 7 mL Teflon beaker on hot-plate at ~130°C for 2 days. After evaporation, the samples were completely dissolved in a 3:1 mixture of concentrated HCl-HNO₃ and dried again.
The fully dissolved samples were converted to chloride form by repeated redissolution in 1 mL concentrated HCl and subsequent evaporation to dryness. The samples were finally dissolved in 5 mL 7 M HCl and stored in a Teflon beaker as sample stock solution. Based on measured Fe concentrations, an aliquot of the sample stock solution that contained 100 μg Fe was extracted and evaporation to dryness and then dissolved in 100 μL 7 M HCl for chemical purification.

Iron was separated from matrix elements by anion exchange chromatography using a 100 μL Bio-Rad AG MP-1 resin in a custom-made shrinkable Teflon column (4mm ID, 26mm height). Before anion exchange, the resin was cleaned with 1000 μL 2% (volume ratio) HNO3 and 1000 μL Milli-Q H2O2, then conditioned with 2000 μL 7 M HCl. After loading 100 μL sample solution in 7 M HCl onto the resin, the matrix elements were eluted off the column using 3 mL of 7 M HCl in 0.5 mL increments. Iron was subsequently eluted from the resin using 3 mL of 2% HNO3. The Fe cut was evaporated to dryness, redissolved in 100 μL sample in 7 M HCl, and was purified for a second time by repeating the anion exchange procedure as described above. Purified Fe was dried and treated with three drops of 30% H2O2 and 2 mL concentrated HNO3 to decompose organic matters. Then Fe was dissolved in 1 mL 2% HNO3 and ready for mass spectrometry analysis. Recovery of Fe for the column procedure was routinely monitored for each sample by measuring the Fe contents in solutions before and after the iron exchange chromatography using photo spectroscopy (the Ferrozine method), and the Fe recovery was >95%.

Iron isotope ratios were measured using a Thermo Fisher Scientific Neptune Plus MC-ICP-MS at State Key Laboratory for Mineral Deposit Research, Nanjing University. The instrument was running at ‘wet-plasma’ mode using a 100 μL/min self-aspirating nebulizer tip and a glass spray chamber. Molecular interferences of 48Ar/44N and 44Ar/40O on 56Fe and 54Fe were fully resolved using high mass resolution setting of the instrument. Isobaric interference of 56Cr on 54Fe was monitored by simultaneous measurement of 53Cr signals and was corrected offline. Instrument sensitivity was 4–6 V/ppm on 56Fe. δ56Fe is better than ±0.06‰, n = 2. SD). The long-term external reproducibility (2 standard deviation or 2 SD) of Fe isotope ratio measurement is better than ±0.06‰ in 56Fe/54Fe and ±0.16‰ in 57Fe/54Fe over six months, based on repeat analysis of multiple Fe isotope standard solutions against in-house stock solutions.

Accuracy of Fe isotope measurements was confirmed by repeated measurements of reference samples and geostandards that were treated as unknowns with the rhyolite samples. δ56Fe of two ultrapure Fe solutions from University of Wisconsin-Madison, J-M Fe and HPS Fe, are 0.37 ± 0.06‰ (n = 10, 2SD) and 0.58 ± 0.06‰ (n = 7, 2SD), respectively, which are in excellent agreements with the recommended values23,24. In addition, the measured Fe isotope compositions of the international whole-rock standards, DNC-1a (δ56Fe = 0.02 ± 0.06‰, n = 3), BCR-2 (δ56Fe = 0.11 ± 0.08‰, n = 9), BHVO-2 (δ56Fe = 0.13 ± 0.03‰, n = 9), BIR-1a (δ56Fe = 0.08 ± 0.06‰, n = 3) and DTS-2b (δ56Fe = 0.06 ± 0.08‰, n = 3), are all consistent with the recommended values23,24 within analytical uncertainties. For igneous rocks investigated in this study, each sample was measured at least three times and analytical uncertainties of Fe isotope ratios were given as ±0.2 ppm to match the concentration of an in-house standard that was constant at 2.0 ppm. A 40 s on-peak acid blank was measured before each analysis. Each Fe isotope ratio measurement consisted of fifty 4-s integrations, and the typical internal precision (2 standard error or 2SE) was better than ±0.03‰ for 56Fe/54Fe and ±0.05‰ for 57Fe/54Fe. The long-term external reproducibility (2 standard deviation or 2 SD) of Fe isotope analysis is better than ±0.06‰ in 56Fe/54Fe and ±0.16‰ in 57Fe/54Fe over six months, based on repeat analysis of multiple Fe isotope standard solutions against in-house stock solutions.

Zircons U-Pb dating and trace element. All grains were imaged using a Zeiss Supra 55 scanning electron microscope (SEM) equipped with a GeolasPro193nm ArF Excimer laser ablation system combined with an Element XR high resolution inductively coupled plasma mass-spectrometer (ThermoFisher, USA). Laser ablation was conducted in a helium atmosphere, running with an energy density of 6 J/cm2, a pulse repetition rate of 8 Hz, and a spot size of 10 μm. Each time-resolved laser ablation analysis took about 90 s, including 30 s of gas blank measurement (i.e. on-peak zeros), followed by 40 s of laser ablation and 20 s of washout time to allow the signals to drop back to background levels. Zircon standards 9150025 and GJ-1116 were used for calibration and data quality control. Raw data from mass spectrometer were reduced using a Glitter (ver 4.0) software and the U-Pb ages were calculated using Isoplot8 (ver 4.15). Common-Pb corrections were carried out prior to U-Pb age calculation using a well established routine of ComPbCorr3–15 by Andersen (2002)26. The U-Pb isotope and trace element composition of zircons is provided in Supplementary Tables S5 and S6.

Statement of informed consent. Hui Ye appears in Fig. 1d as a scale bar of the orebody and he grants permission on his appearance in this figure.

References
1. Barley, M. E., Bekker, A. & Krapež, B. Late Archean to Early Palaeoproterozoic global tectonics, environmental change and the rise of atmospheric oxygen. Earth and Planetary Science Letters 238, 156–171, https://doi.org/10.1016/j.epsl.2005.03.062 (2005).
2. Bekker, A. et al. Iron formation: The sedimentary product of a complex interplay among mantle, tectonic, oceanic, and biospheric processes. Economic Geology and the Bulletin of the Society of Economic Geologists 105, 467–508, https://doi.org/10.2113/gsecongeo.105.3.467 (2010).
3. Klein, C. Some Precambrian banded iron-formations (BIFs) from around the world: Their age, geologic setting, mineralogy, metamorphism, geochemistry and origins. American Mineralogist 90, 1473–1499, https://doi.org/10.2138/am.2005.1871 (2005).
4. Li, W., Beard, B. L. & Johnson, C. M. Biologically recycled continental iron is a major component in banded iron formations. *Proceeding of the National Academy of Sciences of the United States of America* **112**, 8193–8198, https://doi.org/10.1073/pnas.1505131112 (2015).
5. Cox, G. M. et al. Neoproterozoic iron formation: An evaluation of its temporal, environmental and tectonic significance. *Chemical Geology* **362**, 232–249, https://doi.org/10.1016/j.chemgeo.2013.08.002 (2013).
6. Zheng, J. et al. Widespread Archean basement beneath the Yangtze craton. *Geology* **34**, 417–420, https://doi.org/10.1130/G22282.1 (2006).
7. Zhao, G. C. & Cawood, P. A. Precambrian geology of China. *Precambrian Research* **222–223**, 13–54, https://doi.org/10.1016/j.precamres.2012.09.017 (2012).
8. Qu, Y. M., Gao, S., McNaughton, N. J., Groves, D. I. & Ling, W. L. First evidence of ~3.2 Ga continental crust in the Yangtze craton of south China and its implications for Archean crustal evolution and Phanerozoic tectonics. *Geology* **28**, 11–14, https://doi.org/10.1130/0091-7613(2000)028<0011:FEAOCT>2.3.CO;2 (2000).
9. Gao, S. et al. Age and growth of the Archean KONGLing terrain, South China, with emphasis on 3.3 Ga granitoid gneisses. *American Journal of Science* **311**, 153–182, https://doi.org/10.2475/02.2011.03.03 (2011).
10. Shen, B. F. Geological Characters and Resource Prospect of the BIF Type Iron Ore Deposits in China. *Acta Geologica Sinica* **86**, 1376–1395 (2012).
11. Li, H. M. et al. Types and general characteristics of the BIF-related iron deposits in China. *Ore geology reviews* **57**, 264–287, https://doi.org/10.1016/j.oregeorev.2013.09.014 (2014).
12. Wang, C., Konhauser, K. O. & Zhang, L. Depositional environment of the Paleoproterozoic Yunnajacun banded iron formation in Shaxi Province, China. *Economic Geology* **110**, 1515–1539, https://doi.org/10.2113/econgeo.110.6.1515 (2015).
13. Meert, J. G. & Santosh, M. The Columbia supercontinent revisited. *Gondwana Research* **6**, 22–53 (1985).
14. Zhang, G. W., Yu, Z. P., Dong, Y. F. & Yao, A. P. On Precambrian framework and evolution of the Qinling belt. *Acta Petrologica Sinica* **16**, 11–21 (2000).
15. Meng, Q. R., Li, S. Y. & Li, R. W. Mesozoic evolution of the Hefei basin in eastern China: Sedimentary response to deformations in the adjacent Dabieshan and along the Tanlu fault. *Geological Society of America Bulletin* **119**, 897–916, https://doi.org/10.1130/B25931.1 (2007).
16. Zhao, T., Zhu, G., Lin, S. & Wang, H. Indentation-induced tearing of a subducting continent: Evidence from the Tan-Lu fault zone, East China. *Earth Science Reviews* **152**, 14–36, https://doi.org/10.1016/j.earscirev.2015.11.003 (2016).
17. Xu, J. W., Zhu, G., Tong, W. X., Cui, K. R. & Liu, Q. Formation and evolution of the Tancheng-Lijiang wrench fault system: a major shear system to the northwest of the Pacific Ocean. *Tectonophysics* **134**, 273–310, https://doi.org/10.1006/jtph.1997.0229 (1997).
18. Ewers, W. E. & Morris, R. C. Studies of the Dales Gorge member of the Brockman iron formation, Western Australia. *Geology* **1987**.
19. Li, W., Beard, B. L. & Johnson, C. M. Application of Fe isotopes to tracing the geochemical and biological cycling of Fe. *Chemical Geology* **195**, 87–117, https://doi.org/10.1016/S0009-2541(02)00390-X (2007).
20. Teng, F. Z., Dauphas, N., Huang, S. & Marty, B. Iron isotopic systematics of oceanic basalt. *Geochimica et Cosmochimica Acta* **107**, 12–26, https://doi.org/10.1016/j.gca.2012.12.027 (2013).
21. Heimann, A., Beard, B. L. & Johnson, C. M. The role of volatile exsolution and sub-solidus fluid/rock interactions in producing high Sr/86 Fe/54 Fe ratios in siliceous igneous rocks. *Geochimica et Cosmochimica Acta* **72**, 4379–4396, https://doi.org/10.1016/j.gca.2008.06.009 (2008).
22. Telus, M. et al. Iron, Zinc, Magnesium and Uranium Isotopic Fractionation During Continental Crust Differentiation: The tale from migmatites, granitoids, and pegmatites. *Geochimica et Cosmochimica Acta* **97**, 247–265, https://doi.org/10.1016/j.gca.2012.08.024 (2012).
23. Dauphas, N. et al. Clues from Fe isotope variations on the origin of early Archean BIFs from Greenland. *Science* **306**, 2077–2080, https://doi.org/10.1126/science.1104639 (2004).
24. Johnson, C. M., Beard, B. L. & Roden, E. E. The iron isotope fingerprints of redox and biogeochemical cycling in modern and ancient Earth. *Annual Review of Earth Planetary Science* **36**, 457–493, https://doi.org/10.1146/annurev.earth.36.031207.124139 (2008).
25. Trendall, A. F., Compston, W., Nelson, D. R., De Laeter, J. R. & Bennett, V. C. SHRIMP zircon ages constraining the depositional chronology of the Hamersley Group, Western Australia. *Australian Journal of Earth Sciences* **51**, 621–644, https://doi.org/10.1111/j.1440-0929.2004.01082.x (2004).
26. Chennaik, D. J. & Watson, E. B. Diffusion in zircon. *Reviews in Mineralogy and Geochemistry* **53**, 113–143, https://doi.org/10.2113/0530113 (2003).
27. Hoskin, P. W. O. & Black, L. P. Metamorphic zircon formation by solid-state recrystallization of protolith igneous zircon. *Journal of Metamorphic Geology* **18**, 423–439, https://doi.org/10.1046/j.1525-1314.2000.00266.x (2000).
28. Heaman, L. M. Global mafic magmatism at 2.45 Ga: Remnants of an ancient large igneous province? *Geology* **25**, 299–302, https://doi.org/10.1130/0091-7613(1997).
29. Bleeker, W. The late Archean record: a puzzle in 35 pieces. *Lithos* **71**, 99–134, 10.1016/j.lithos.2003.07.003 (2003).
30. Condé, K. C., Belousova, E., Griffin, W. L. & Sircombe, K. N. Granitoid events in space and time: constraints from igneous and detrital zircon age spectra. *Gondwana Research* **15**, 228–242, https://doi.org/10.1016/j.gr.2008.06.001 (2009).
39. Pehrsson, S. J., Berman, R. G., Eglington, B. & Rainbird, R. Two Neoarchean supercontinents revisited: the case for a Rae family of cratons. *Precambrian Research* **232**, 27–43, https://doi.org/10.1016/j.precamres.2013.02.005 (2013).
40. Han, P. Y. *et al*. Widespread Neoarchean (~2.7–2.6 Ga) magmatism of the Yangtze craton, South China, as revealed by modern river detrital zircons. *Gondwana Research* **42**, 1–12, https://doi.org/10.1016/j.gr.2016.09.006 (2017).
41. Marechéal, C. N., Télouk, P. & Albarelle, F. Precise analysis of copper and zinc isotopic compositions by plasma-source mass spectrometry. *Chemical Geology* **156**, 251–273, https://doi.org/10.1016/S0009-2541(98)00191-0 (1999).
42. Graddock, P. R. & Dauphas, N. Iron isotopic compositions of geological reference materials and chondrites. *Geostandards and Geoanalytical Research* **35**, 101–123, https://doi.org/10.1111/j.1751-908X.2010.00085.X (2011).
43. He, Y. *et al*. High-precision iron isotope analysis of geological reference materials by high-resolution mc-icp-ms. *Geostandards and Geoanalytical Research* **39**, 341–356, https://doi.org/10.1111/j.1751-908X.2014.00304.x (2015).
44. Wiedenbeck, M. A. P. *et al*. Three natural zircon standards for U-Th-Pb, Lu-Hf, trace element and REE analyses. *Geostandards newsletter* **19**, 1–23, https://doi.org/10.1111/j.1751-908X.1995.tb00147.x (1995).
45. Jackson, S. E., Pearson, N. J., Griffin, W. L. & Belousova, E. A. The application of laser ablation-inductively coupled plasma-mass spectrometry to in situ U-Pb zircon geochronology. *Chemical Geology* **211**, 47–69, https://doi.org/10.1016/j.chemgeo.2004.06.017 (2004).
46. Andersen, T. Correction of common lead in U-Pb analyses that do not report 204 Pb. *Chemical Geology* **192**, 59–79, https://doi.org/10.1016/S0009-2541(02)00195-X (2002).

**Acknowledgements**

This manuscript benefits from constructive reviews from F.-X. d’Abzac, and Franco Pirajno. We also thank Dr. Massimo Chiaradia for his editorial handling and constructive comments. This study is supported by Natural Science Foundation of China (No. U1603114 to CZW and No. 41622301 to WL). We thank Xiaopeng Bian, Dehong Du, Xiaoming Wang, Shugao Zhao, Chuan Liu for assistance in field and laboratory.

**Author Contributions**

C.Z.W., W.L., and T.Y. designed the project. H.Y. and X.Z.Y. performed the geochemical and geochronological analyses. C.Z.W., W.L., T.Y. and H.Y. wrote the manuscript. X.L.W., M. Santosh, and B.F.G. contributed to the scientific discussion.

**Additional Information**

Supplementary information accompanies this paper at https://doi.org/10.1038/s41598-017-15013-4.

**Competing Interests:** The authors declare that they have no competing interests.

**Publisher’s note:** Springer Nature remains neutral with regard to jurisdictional claims in published maps and institutional affiliations.

© The Author(s) 2017