Sr-Nd-Hf-Pb isotopic evidence for modification of the Devonian lithospheric mantle beneath the Chinese Altai

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Abstract

Intensive Devonian felsic magmatism is recorded within the southwestern Mongolian collage system of the Central Asian Orogenic Belt (CAOB). The voluminous magmas have isotopic compositions of juvenile materials from the mantle, thus manifesting significant mantle-crust interaction and continental growth at this time. Here, we present systematic Sr-Nd-Hf-Pb isotopic data for the Devonian mafic intrusions in the Chinese Altai, a key region within the southwestern Mongolian collage system to decipher the evolution of the mantle during this important tectonothermal event. The Keketuohai gabbro (409 ± 5 Ma) and type I mafic dykes (376 ± 5 Ma) within the Habahe complex have high $(^{87}\text{Sr}/^{86}\text{Sr})_i$, $(^{206}\text{Pb}/^{204}\text{Pb})_t$, $(^{207}\text{Pb}/^{204}\text{Pb})_t$ and $(^{208}\text{Pb}/^{204}\text{Pb})_t$ ratios, and decoupled Nd-Hf isotopic compositions; e.g., low $\epsilon_{\text{Nd}}(t)$ values (-2.5 to +5.4) combined with high $\epsilon_{\text{Hf}}(t)$ (+2.6 to +15.1) values. These rocks have low Ba/La and high La/Yb and Th/Yb ratios, and are enriched in Pb, the light rare earth elements (LREE) and Th. They formed from magmas generated from the depleted lithospheric mantle metasomatised by hydrous melts from subducted sediments. In comparison, the gabbroic samples from the Habahe complex (369 ± 3 Ma) are enriched in the LREE, Th and Ba and have high La/Yb, Th/Yb and Ba/La ratios. They do not show significant Pb anomalies, and have depleted isotopic compositions that include low initial $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios and high $\epsilon_{\text{Nd}}(t)$ (+7.4 to +7.8) and $\epsilon_{\text{Hf}}(t)$ (+13.4 to +15.3) values. These rocks are thought to have formed from magmas derived from the lithospheric mantle metasomatised by hydrous melts from subducted oceanic crust. The type II mafic dykes within the Habahe complex are depleted in the LREE and Th, have high Ba/La ratios, and are enriched in Pb, Ba, Sr, and U. They

have positive $\epsilon_{\text{Nd}}(t)$ (+7.6 to +8.1) and $\epsilon_{\text{Hf}}(t)$ (+14.1 to +15.4) values, high initial $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios, and are thought to have formed from a magmas generated by the partial melting of the depleted lithospheric mantle metasomatised by fluids derived from subducted sediments. The melting and dehydration of subducted oceanic crust and sediments indicate the Devonian subduction of a hot slab beneath the Chinese Altai, which could be related to a ridge subduction model previously proposed. The involvement of hydrous melts and fluids promoted the melting of the lithospheric mantle which was responsible for intensive felsic magmatism with abrupt increase of juvenile component and high temperature metamorphism in the region.

Key words: Central Asian Orogenic Belt; Chinese Altai; mantle metasomatism; ridge subduction; continental growth.

1. Introduction

The Central Asian Orogenic Belt (CAOB) is one of the largest accretionary orogenic belts on the Earth and represents a major site of continental growth in the Phanerozoic. It extends from the Urals in the west to the Pacific in the east and from Siberia in the north to the North China and Tarim cratons in the south (Fig. 1a; Jahn et al., 2000a; 2000b; Sengör et al., 1993; Windley et al., 2007). The CAOB was formed by a series of complex subduction and accretion processes that involved numerous allochthonous blocks, including island arcs, ophiolitic complexes, seamounts, oceanic plateaus, accretionary prisms, and microcontinental fragments (Fig. 1a; Windley et al., 2007; Xiao et al., 2015). Extensive Devonian felsic magmatism is recorded within the southwestern Mongolian collage system, including the Lake and Hovd zones in Mongolia, the Altai Mongolia terrane in Mongolia, China and Russia, the Trans Altai and Gobi Altai regions of Mongolia (Fig. 1; Cai et al., 2011a, 2015; Yu et al., in press; Yuan et al., 2007). The Devonian granites within these areas generally have positive $\varepsilon_{\text{Hf}}(t)$ values that are indicative of significant continental growth. As such, understanding the processes that generated both the felsic magmatism in this region and the associated continental growth is critical to furthering our knowledge of the tectonic evolution of the southwestern Mongolian collage system and the CAOB as a whole.

The Chinese Altai, which is a key part of the southwestern Mongolian collage system, contains voluminous early to middle Paleozoic granitic plutons that record a magmatic event that peaked at around 400 Ma (e.g., Cai et al., 2011a, 2012; Sun et al., 2008; Wang et al., 2009; Yu et al., in press; Yuan et al., 2007). The Hf isotopic compositions of zircons

from these granites and metasedimentary rocks in this region record an abrupt increase of $\epsilon_{\text{Hf}}(t)$ values at about 400 Ma (Sun et al., 2009). The U-Pb ages of overgrowth rims on zircons from high-grade metamorphic rocks in this area also record a high temperature metamorphic event at about 390 Ma (Jiang et al., 2010). This demonstrates the importance of the Devonian for the geological evolution of the Chinese Altai, as this period of time records the significant transfer of heat and juvenile material from the mantle to the crust (Sun et al., 2009; Ye et al., 2015). Although some detailed geochemical and isotopic research has have been undertaken on the granites and metasedimentary rocks in this area, Devonian mafic rocks in the region have received little attention. Xu et al. (2003) reported that some mafic rocks in the Kuerti area are light rare earth element (LREE) depleted and have high $\epsilon_{\text{Nd}}(t)$ values, suggesting they were possibly related to the opening of a back-arc basin. Niu et al. (2006) identified adakitic and boninitic units within the southern margin of the Chinese Altai and suggested that the magmas that formed these rocks were generated in response to slab breakoff. Cai et al. (2012) analysed the geochemistry of the Keketuohai gabbro and suggested that this intrusion formed from magmas generated as a result of oceanic ridge subduction. The range of tectonic interpretations for these Devonian mafic rocks is the result of a lack of isotopic data and uncertainty regarding the processes that drove mantle melting and mantle-crust interaction. Here we present integrated geochemical and Sr-Nd-Pb-Hf isotopic data for Devonian mafic rocks in the study area, including the Keketuohai and Habahe gabbros and mafic dykes from the Habahe complex, and use these data to constrain the processes that operated in the mantle during Devonian tectonism within the Chinese Altai.

Fig. 1

2. Geological background and sample descriptions

2.1. Geological background

The Chinese Altai is located in the southwestern part of the Mongolian collage system and is separated from the East and West Junggar terranes by the Erqis Fault (Fig. 1b; also known as the Irtysh Fault, Irtyshi Fault, or Ertix Fault; Li et al., 2015; Zhang et al., 2012), one of the largest (1000 km long) sinistral strike-slip faults in Central Asia (Fig. 1b). The area contains NW-SE trending fault systems and the geology of this region is dominated by four fault-bounded units with distinct lithological assemblages (Fig. 1b). The North Altai Domain (unit 1 of Windley et al., 2002) contains of Late Devonian-early Carboniferous neritic and clastic sedimentary units, roughly coeval volcanic units, and limestones, the vast majority of which have undergone sub-greenschist-facies metamorphism. The Central Chinese Altai Domain (units 2 and 3 of Windley et al., 2002) dominates the Chinese Altai and contains widespread Habahe Group sedimentary rocks that represent the oldest sedimentary sequence in this area and consist of a thick sequence (~6000 m) of shale, phyllite, and schist units (He et al., 1990; Li H.J. et al., 2006; Windley et al., 2002). Detrital zircons from these rocks yield $^{206}\text{Pb}/^{238}\text{U}$ ages that generally range between 540 and 460 Ma, and have positive $\varepsilon_{\text{Hf}}(t)$ values (Long et al., 2012; Sun et al., 2008). The Qiongkuer Domain (unit 4 of Windley et al., 2002) contains late Silurian to Early Devonian volcanic and pyroclastic rocks of the Kangbutiebao Formation and a

Middle Devonian turbiditic sandstone-shale sequence of the Altai Formation. The South Altai Domain (unit 5 of Windley et al., 2002) includes a sequence of Devonian sedimentary rocks and Carboniferous volcanic and clastic rocks, all of which have undergone greenschist- to amphibolite-facies metamorphism. The Chinese Altai also contains voluminous early to middle Paleozoic I- and S-type granites that are enriched in the large ion lithophile elements (LILE) and the LREE, and are depleted in the high field strength elements (HFSE), suggesting these units have arc type affinities (Fig. 1b, Cai et al., 2011a; Yu et al., in press; Yuan et al., 2007). These granites have decoupled Nd-Hf isotopic compositions with low $\epsilon_{\text{Nd}}(t)$ (-3 to +2) and high $\epsilon_{\text{Hf}}(t)$ (+2 to +13) values (Yu et al., in press). The mafic rocks in the Chinese Altai are generally confined to the central and southern marginal parts of this region (Fig. 1b), including the Keketuohai mafic complex, the Habahe complex, the Kuwei complex, and mafic rocks within the Kuerti area.

2.2. Sample descriptions

The samples in this study were collected from the mafic dykes (376 ± 5 Ma; Cai et al., 2010) and gabbroic intrusions within the Habahe complex and from gabbroic intrusions within the Keketuohai mafic complex (409 ± 5 Ma; Cai et al., 2012).

2.2.1. Samples from the Habahe complex

The Habahe complex crops out over an area $>400 \text{ km}^2$ to the northeast of the Habahe town and was emplaced into the Devonian strata and is partly covered by Carboniferous

volcanic rocks (Fig. 2a). The complex is zoned with an outermost fine-grained biotite granite giving way to a central region containing medium- to coarse-grained gabbro, diorite and plagiogranite (389 ± 5 ; Cai et al., 2011a; Fig. 2a). The southern part of the complex has been intruded by mafic dykes (Fig. 2a) that have widths of 0.5–5 m and lengths of 100–500 m. These mafic dykes are black, and are generally aphanitic, and contain very fine-grained amphiboles. The edges of the Habahe gabbro records magma mixing with the adjacent plagiogranite (Fig. 3a). The gabbroic samples contain euhedral-subhedral amphibole (40%–50%), subhedral plagioclase (40%–50%), and accessory zircon and Fe-Ti oxides (Fig. 3c). Some of the plagioclase within the gabbro is partially altered to fine-grained mica-rich aggregates, whereas the amphiboles in this unit are generally fresh and are zoned (Fig. 3c).

2.2.2. Keketuohai mafic complex

The roughly elliptical Keketuohai mafic complex crops out over an area of ca. 4.5 km² near the town of Keketuohai (Fig. 2b). The complex was emplaced into the Habahe Group, has resulted in the formation of a contact aureole within the surrounding units, and contains a central region that is covered by the Quaternary sediments (Fig. 2b). The complex also contains fine-grained plagioclase-rich layers, providing evidence of crystal accumulation processes (Fig. 3b). The gabbroic samples contain plagioclase (35%–40%), clinopyroxene (10%–15%), amphibole (45%–50%), and minor ilmenite (Fig. 3d). The primary amphibole within these samples varies in size from 100 to 500 μm and is generally euhedral, whereas

plagioclase is subhedral-anhedral, and is enclosed in clinopyroxene and amphibole. Ilmenite is generally present interstitial crystals between amphibole and plagioclase.

Fig. 2 and 3

3. Analytical methods

3.1. Zircon U-Pb and Hf isotopic analyses

Zircons were separated using conventional heavy liquid and magnetic techniques before being hand-picked under a binocular microscope and embedded within an epoxy mount. The mount was then polished to expose the zircon centres before being gold coated. Zircon internal structures were examined using cathodoluminescence (CL) imaging prior to U-Pb isotopic analysis.

Zircon U-Pb isotopic analyses were undertaken using a Cameca IMS 1280 ion microprobes at the Guangzhou Institute of Geochemistry, Chinese Academy of Sciences (GIG-CAS), Guangzhou, China. The analytical procedures used were similar to those described by Li et al. (2009a) and employed a primary $O_2-O_2^-$ ion beam with an intensity of ~10 nA that was accelerated at -13 kV. The ellipsoidal spot is about $20 \times 30 \mu m$ in size. Zircon Pb/U ratios were calibrated using a 337.13 Ma Plesovice standard zircon (Slama et al., 2008) that was analysed once every four unknowns, with the resulting calibration based on an observed linear relationship between $\ln (^{206}Pb/^{238}U)$ and $\ln (^{238}U^{16}O_2/^{238}U)$ (Whitehouse et al., 1997). The U and Th contents of unknowns were also calibrated using the standard zircon Plesovice, which has Th and U concentrations of 78 and 755ppm,

respectively (Slama et al., 2008). Measured Pb isotopic ratios were corrected for common Pb using non-radiogenic ^{204}Pb which was very low, and possibly largely derived from laboratory contamination introduced during sample preparation (Ireland and Williams, 2003). An average present-day crustal Pb isotopic composition (Stacey and Kramers, 1975) was used to correct for common Pb. A secondary Qinghu standard zircon (Li et al, 2013) was also analysed to monitor the reproducibility of the entire procedure, yielding a concordant age of 159.4 ± 3.8 Ma ($n = 3$) that is within uncertainty of the recommended value of this standard zircon (159.5 ± 0.2 Ma). The uncertainties on single analyses are reported at 1σ level, with weighted mean ages for pooled U-Pb analyses quoted at a 95% confidence interval. All data reduction was performed using Isoplot/Ex v.3 software (Ludwig, 2003).

Zircon Hf isotope analyses were conducted by employing Nu Plasma HR multi-collector inductively coupled plasma-mass spectrometry (MC-ICP-MS) equipped with a Resonetics Resolution M-50-HR excimer laser ablation (LA) system at the Department of Earth Sciences, The University of Hong Kong, Hong Kong. These analyses were performed with a laser beam diameter of ca. 55 μm and a 6 Hz repetition rate, yielding a signal intensity of 0.04 V at ^{179}Hf during the analysis of the standard zircon 91500. These analyses typically used an ablation time of 40 s, yielding pits of 30–40 μm deep. Masses 172–179 were simultaneously measured in static-collection mode. Standard zircons 91500 and GJ were used as external standards and were analysed twice before and after every 10 analyses. The resulting data were normalised to $^{179}\text{Hf}/^{177}\text{Hf}=0.7325$ using an exponential correction for mass bias. The interference of ^{176}Lu on ^{176}Hf was corrected by measuring

the intensity of the interference-free ^{175}Lu isotope and using the recommended $^{176}\text{Lu}/^{175}\text{Lu}$ ratio of 0.02655 (Machado, 2001). Isotope ratio uncertainties for sample, standards, and reference values are given as $\pm 2\sigma$.

3.2. Whole-rock major and trace elements

Major element compositions were determined using fused glass beads by Philips PW 2400 X-ray fluorescence (XRF) spectroscopy at the Department of Earth Sciences, The University of Hong Kong. These analyses used the procedures described by Norrish and Hutton (1969) for matrix corrections and yielded uncertainties of 1%–2% for SiO_2 , Al_2O_3 , TiO_2 , Fe_2O_3 and MgO , and $\leq 3\%$ for other major elements. Trace element concentrations were determined by ICP-MS employing Perkin-Elmer Sciex ELAN 6000 instrument at GIG-CAS, Guangzhou, China after samples were digested in high-pressure Teflon vessels. Details of the procedures used were described by Li X.H. et al. (2006). Trace element concentrations were calibrated using USGS and Chinese National standards AGV-2, GSR-1, GSR-2, GSR-3, BHVO-2, W-2a and SARM-4 with the resulting REE and other incompatible element concentrations typically having analytical precisions of 1%–5%.

3.3. Whole-rock Sr-Nd-Hf-Pb isotopes

Strontium, Nd, and Hf isotopic analyses were conducted at the Institute of Tibetan Plateau Research, Chinese Academy of Sciences, Beijing, China. Sr, Nd, and Hf isotopic ratios were determined for a subset of whole-rock samples using a Nu Plasma II MC-ICP-MS instrument. All chemical preparations were performed on the specially designed class-

100 work bench inside a class-1000 clean laboratory. Powder of a rock was weighed (about 150mg) into a 7ml round bottom Savillex™ Teflon screw-top capsule. One drop of purified water was added to make the sample wet before a mixture of concentrated HF-HNO₃ (2ml–2ml) was added to the sample. The capsule screw-top cap was then closed and the capsule was placed on a 150°C hotplate for 48 hours. The capsule was then opened and gently evaporated before 0.5 ml HClO₄ was added and the sample evaporated to dryness once again. The resulting dry sample was mixed with concentrated HF-HNO₃ (1 ml–1ml) before the capsule was closed and heat for 120 hours under 150°C. The capsule was then cooled, opened, and gently evaporated before adding 0.5 ml HNO₃ and once again evaporating to dryness. The sample was then mixed with 1ml HNO₃ and 1 ml HBrO₄ (in a 3M HNO₃ solution) before the capsule being sealed and placed on a hot plate overnight at 120°C. The sample was then mixed with 4 ml of purified water prior to chemical separation and was then centrifuged. The 3M HNO₃ sample solution was loaded onto a pre-conditioned 2ml DGA Spec resin column to initially collect Sr with 8 ml of 6M HNO₃. The LREE and the middle REE (MREE) were collected using 12 ml of 2.3 M HCl before Nd was sequentially collected in a 30ml PFA beaker using 8 ml of 1.2 M HCL. Finally, Hf (+Zr) fractions were collected using 15 ml of 0.5M HF before being gently evaporated to dryness. This fraction was taken up using a trace amount of 2M HF diluted to 5ml using 2% HNO₃ to prepare the sample for analysis. Measured ⁸⁷Sr/⁸⁶Sr, ¹⁴³Nd/¹⁴⁴Nd, and ¹⁷⁷Hf/¹⁷⁶Hf ratios were normalised to ⁸⁶Sr/⁸⁸Sr = 0.1194, ¹⁴⁶Nd/¹⁴⁴Nd = 0.7219, and ¹⁷⁹Hf/¹⁷⁷Hf = 0.7325, respectively, using the exponential law. Two reference standards were analysed along with the unknowns, yielding ⁸⁷Sr/⁸⁶Sr ratios of 0.703999±10 (1s), ¹⁴³Nd/¹⁴⁴Nd ratios of 0.512781±12 (1s) and ¹⁷⁷Hf/¹⁷⁶Hf ratios of 0.282947±6 (1s) for AGV-2, and ⁸⁷Sr/⁸⁶Sr ratios

of 0.705011 ± 11 (1s), $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of 0.512629 ± 7 (1s), and $^{177}\text{Hf}/^{176}\text{Hf}$ ratios of 0.282863 ± 7 (1s) for BCR-2. These ratios are similar to the recommended $^{87}\text{Sr}/^{86}\text{Sr}$ (0.703995), $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51278 ± 40 (2 σ)), and $^{177}\text{Hf}/^{176}\text{Hf}$ (0.282954 ± 19 (2 σ)) ratios for AGV-2 (Jochum and Nehring, 2007; Weis et al., 2007) and the recommended $^{87}\text{Sr}/^{86}\text{Sr}$ (0.705015 ± 13), $^{143}\text{Nd}/^{144}\text{Nd}$ (0.512636 ± 4 (2 σ)), and $^{177}\text{Hf}/^{176}\text{Hf}$ (0.282878 ± 14 (2 σ)) ratios for BCR-2 (Balcaen et al., 2005; Jochum and Nehring, 2006).

Lead isotopic analyses were carried out at GIG-CAS, Guangzhou, China, using a Neptune Plus MC-ICP-MS. About 100mg of rock powder was weighed into a Teflon vessel and dissolved in a $\text{HNO}_3 + \text{HF}$ mixture at 140°C for 72 h. The resulting solution was evaporated to dryness before being eluted using 2ml of concentrated HNO_3 and being placed on a hot plate at 140°C for 24 h. The solution was then evaporated to dryness again, subsequently added with 2ml of 6M HCl and being placed on a hot plate at 140°C for another 24 h. The sample was finally dissolved in a 0.8M HBr solution, and Pb was separated and purified by conventional anion exchange technique (AG1X8, 200-400 resin) with dilute HBr as eluant. The total procedural blank was less than 0.4 ng and the samples were doped with Tl, enabling mass discrimination correction using certified $^{203}\text{Tl}/^{205}\text{Tl}$ ratio of 0.418922. The detail analytical procedures are similar to those described by Baker and Waight (2002), with repeated analyses of National Institute of Standards and Technology Standard Reference Material 981 yielding $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of 16.9325 ± 5 (2 σ), $^{207}\text{Pb}/^{204}\text{Pb}$ ratios of 15.4859 ± 6 (2 σ), and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios of 36.6825 ± 18 (2 σ).

4.1. Results

4.1. Zircon U-Pb and Hf isotopes

The results of zircon U-Pb and Hf isotopic analyses of the Habahe gabbro sample ALT14–56 are given in Supplementary Appendices 1 and 2, respectively. The zircons from this sample are generally euhedral and are elongated to stubby and prismatic. These zircons show strongly oscillatory zoning in CL images, indicative of a magmatic origin. They contain variable concentrations of Th and U (183–1607 and 50–643 ppm, respectively) that yield Th/U ratios of 0.27–1.04. Thirteen zircons form a coherent cluster that yields a mean $^{206}\text{Pb}/^{238}\text{U}$ age of 369 ± 3 Ma (MSWD (mean square weighted deviation) = 0.49; Fig. 4a). These zircons have similarly high $^{176}\text{Hf}/^{177}\text{Hf}$ ratios (0.282823–0.282983) that yield high positive $\epsilon_{\text{Hf}}(t)$ values (+9.9 to +15.6; Fig. 4b). Analytical spot 01 contains the lowest concentrations of Th (183 ppm) and U (50 ppm) and yielded the youngest $^{206}\text{Pb}/^{238}\text{U}$ age of 352 ± 5 Ma, with a further five other grains yielding discordant $^{206}\text{Pb}/^{238}\text{U}$ ages that range from 359 ± 5 Ma to 365 ± 7 Ma and are indicative of disturbance of the U-Pb isotopic system within these grains.

Fig. 4

4.2. Major and trace elements

The whole-rock major and trace element compositions of representative the Habahe and Keketuohai gabbro samples are given in Table 1 and are shown in Figures 5 and 6. In addition, all of our new data are combined with previously published data for the Keketuohai gabbro (Cai et al., 2012), the Habahe mafic dykes (Cai et al., 2010), the Kuwei gabbro (Ye et al., 2015) and the Kuerti mafic rocks (Xu et al., 2003) in Supplementary Appendix 3.

4.2.1. Habahe gabbro

The Habahe gabbroic samples are calc-alkaline, have low FeO/MgO ratios (Fig. 5d), contain elevated concentrations of MgO, SiO₂, CaO, and Na₂O, and low concentrations of TiO₂, Al₂O₃, Fe₂O₃, K₂O, and P₂O₅ (Table 1; Fig. 5). These samples have highly fractionated chondrite-normalised REE patterns ($[La/Yb]_N = 2.8\text{--}2.9$) with weakly negative Eu anomalies ($Eu/Eu^* = 0.90\text{--}1.00$; Fig. 6a). On the primitive mantle-normalised multi-element variation diagram, they show enrichments in Rb, Ba, Th and U and depletions in Nb and Ti, with insignificant Pb anomalies (Fig. 6b).

4.2.2. Habahe mafic dykes

The major and trace elements compositions of the mafic dykes within the Habahe complex were reported by Cai et al. (2010). These dykes all have high FeO/MgO and low K₂O/Na₂O ratios, and are therefore classified as low-K tholeiitic series. Based on their trace element compositions, they can be divided into two types. Type I mafic dykes are enriched in Th and the LREE ($[La/Sm]_N = 1.0\text{--}2.6$; $[La/Yb]_N = 0.89\text{--}2.3$) (Fig. 6a,b), whereas the type II mafic dykes are depleted in Th and the LREE ($[La/Sm]_N = 0.77\text{--}0.84$; $[La/Yb]_N = 0.87\text{--}1.1$), and are enriched in Sr, Ba, U, and Pb (Fig. 6a, b).

4.2.3. Keketuohai gabbro

The Keketuohai gabbroic samples are low-K tholeiitic, have high FeO/MgO and low K₂O/Na₂O ratios (Fig. 5), and are characterised by moderate to high TiO₂, Fe₂O₃, Al₂O₃, and CaO concentrations and low concentrations of SiO₂, Na₂O, K₂O, and P₂O₅ (Table 1).

The positive correlation between CaO and Mg[#] and negative correlation between Al₂O₃ and Mg[#] suggest fractional crystallisation/accumulation of clinopyroxene and plagioclase, respectively (Fig. 5). These samples have fractionated REE patterns ([La/Yb]_N = 1.5–5.5) with generally positive Eu anomalies (0.96–1.4; Fig. 6c, d). On the primitive mantle-normalised multi-element variation diagram, they show enrichments in Th, and have positive Pb and Sr anomalies and strongly negative Nb anomalies (Fig. 6d).

Fig. 5 and 6

4.3. *Sr-Nd-Hf-Pb isotope geochemistry*

The Sr-Nd-Hf-Pb isotopic data for gabbroic samples from the Keketuohai mafic complex and the mafic dykes and gabbroic samples from the Habahe complex are given in Table 2, Supplementary Appendices 4 and 5, and Fig. 7.

4.3.1. *Habahe gabbro*

The Habahe gabbroic samples are characterised by low initial ⁸⁷Sr/⁸⁶Sr ratios (0.7034–0.7036) and high positive ε_{Nd}(t) (+7.5 to +7.8) and ε_{Hf}(t) (+13.4 to +15.3) values (Fig. 7a, b). These whole-rock ε_{Hf}(t) values are similar to the zircon ε_{Hf}(t) values (+9.9 to +15.6). These samples also have fairly uniform initial ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb, and ²⁰⁸Pb/²⁰⁴Pb ratios (17.863–18.175, 15.485–15.500, and 37.500–37.940 respectively) (Fig. 7c, d) that plot slightly above the Northern Hemisphere Pb reference line (NHRL; Fig. 7c, d; Hart, 1984).

4.3.2. *Habahe mafic dykes*

The type I Habahe mafic dykes have high and variable initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7049 to 0.7070), and positive $\epsilon_{\text{Nd}}(\text{t})$ (+0.7 to +5.5) and $\epsilon_{\text{Hf}}(\text{t})$ (+12.1 to +17.1) values (Fig. 7a, b), all of which are similar to the Keketuohai gabbro samples. This decoupling between Nd and Hf isotopes is similar to the decoupling identified within mafic rocks from the Sunda and Lesser Antilles arcs (Fig. 7b). The mafic dykes in the study area have high initial $^{206}\text{Pb}/^{204}\text{Pb}$ (17.977–18.169), $^{207}\text{Pb}/^{204}\text{Pb}$ (15.520–15.553), and $^{208}\text{Pb}/^{204}\text{Pb}$ (37.796–38.296) ratios that all plot above the NHRL (Fig. 7c, d; Hart, 1984).

The type II mafic dykes have low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7041–0.7048) and high $\epsilon_{\text{Nd}}(\text{t})$ values (+7.7 to +8.1; Fig. 7a; Cai et al., 2010). They also have high $^{176}\text{Hf}/^{177}\text{Hf}$ ratios (0.283181–0.283292) with positive $\epsilon_{\text{Hf}}(\text{t})$ values (+14.1 to +15.4; Fig. 7), and elevated $(^{207}\text{Pb}/^{204}\text{Pb})_{\text{t}}$ and $(^{208}\text{Pb}/^{204}\text{Pb})_{\text{t}}$ ratios (15.483–15.494 and 37.734–37.794, respectively) related to their $(^{206}\text{Pb}/^{204}\text{Pb})_{\text{t}}$ ratios (17.705–17.834; Fig. 7c, d).

4.3.3. Keketuohai gabbro

Gabbroic samples from the Keketuohai mafic complex have high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7069–0.7082) and decoupled Nd and Hf isotopic compositions that yield negative $\epsilon_{\text{Nd}}(\text{t})$ (-0.6 to -2.5) and positive $\epsilon_{\text{Hf}}(\text{t})$ (+2.6 to +9.8) values (Fig. 7a, b). These samples have initial $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios of 17.928–18.225, 15.522–15.583, and 37.618–38.037 respectively, all of which plot above the NHRL (Fig. 7c, d; Hart, 1984).

Fig. 7

5. Discussion

5.1. Alteration and crustal contamination

Sericites are observed on the plagioclase within the studied samples (Fig. 3), indicative of chemical alteration which may have affected geochemical and isotopic compositions of the mafic rocks. This means that the effect of alteration on the compositions of these samples needs to be determined prior to any discussion of petrogenesis. The low LOI (loss on ignition) values for these samples (0.59–3.3 wt.%) suggest that they have undergone only minimal alteration (Table 1). In addition, the lack of any correlations between Sr-Nd-Hf-Pb isotopic compositions and Zr concentrations suggests that these isotopic systems were not affected by alteration (Fig. S1). The Ba/La, Th/Yb, and Nd/Hf ratios of these samples show weak correlations with Zr concentrations, again indicating weak to negligible alteration. The crustal contamination would significantly increase SiO₂ concentrations and (⁸⁷Sr/⁸⁶Sr)_i ratios, and decrease ε_{Nd}(t) and ε_{Hf}(t) values. These correlations are not present in the new data presented here, suggesting that the samples generally preserve primary geochemical and isotopic compositions.

5.2. Nature and metasomatism of the Devonian lithospheric mantle beneath the Chinese Altai

The mantle wedge generally have depleted geochemical compositions due to extraction of arc magma, but it can be refertilised by the addition of the LILE and the LREE during metasomatism by fluids and/or melts derived from the subducted slab (e.g., Johnson and Plank, 1999). However, the high field strength elements (HFSE; Nb, Ta, Zr, and Hf) are retained by residual rutile and zircon within the subducted slab and are therefore not added to the mantle wedge (Hermann and Rubatto, 2009). This means that the extent of the

depletion of the mantle wedge can be evaluated using HFSE concentrations and Hf isotopic data. All of the mafic samples analysed during this study have positive $\varepsilon_{\text{Hf}}(t)$ values (Fig. 7b) that are indicative of derivation from a depleted mantle source. This is also evidenced by the low Nb/Y and Zr/Y ratios of these samples, all of which plot in the middle ocean ridge basalt (MORB) field of discrimination diagrams (Fig. 8). The negative Nb anomalies within these samples (Fig. 6) indicate that they were not derived from the magmas from the asthenospheric mantle; rather, they were derived from lithospheric mantle source that was modified by interaction with the subduction-derived fluids or melts.

Fig. 8

5.2.1. Melting of subducted sediments

The Keketuohai gabbros and the type I mafic dykes from the Habahe complex have high initial $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{208}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios and negative $\varepsilon_{\text{Nd}}(t)$ values (Fig. 7) that contrast with their positive $\varepsilon_{\text{Hf}}(t)$ values and a model whereby the magmas that formed these units were derived a depleted mantle source. This fractionation between the Sr-Nd-Pb and Hf isotopic systems could be the result of derivation from a mantle source containing a significant portion of garnet as the high Lu/Hf ratio of garnet can generate high time-integrated $^{176}\text{Hf}/^{177}\text{Hf}$ ratios (and high $\varepsilon_{\text{Hf}}(t)$ values). However, these samples have flat heavy REE (HREE) patterns ($\text{Er}/\text{Yb} < 1.1$; Fig. 6a, c) that are indicative of derivation from mantle within the spinel stability field, thereby precluding this garnet effect. Alternatively, this isotopic decoupling could be related to derivation from a metasomatised mantle source (e.g., Choi and Mukasa, 2012 and references therein). Metasomatic agents can modify Sr-Nd-Pb isotopic systems because the LILE and LREE are more mobile in

fluids and/or melts than are the HFSE (e.g., Todd et al., 2010). The Chinese Altai records a long period of early to middle Paleozoic subduction (Cai et al., 2011; Sun et al., 2008; Wang et al., 2009, 2010), indicating that subducted slab material (i.e., oceanic crustal and overlying terrigenous sedimentary materials) was the most plausible source for the metasomatic agents that caused this fractionation. The fact that the oceanic crust is generally derived from depleted mantle means that fluids and/or melts derived from subducted oceanic crust cannot significantly modify the Sr-Nd-Pb isotopic compositions of the depleted mantle wedge. The negative $\epsilon_{\text{Nd}}(t)$ values and significant positive Pb anomalies of the Keketuohai gabbro and type I mafic dykes suggest that the metasomatic agent that affected the mantle wedge in the study area was most likely derived from subducted sediments (Fig. 6). Subducted materials can be recycled back into the mantle wedge as fluids and/or melts (e.g., Pearce et al., 1999), with fluids preferentially mobilizing the LILE (e.g., Ba, Sr) from subducted sediments but not the LREE and Th (Hermann and Rubatto, 2009; Spandler et al., 2007). This in turn indicates that the addition of sediment-derived fluids would fractionate Ba from La and Sr from Nd (Hermann and Rubatto, 2009; Spandler et al., 2007), yielding a depleted mantle source modified by sediment-derived fluid with high Ba/La ratios and with a prominent concave-down $(^{87}\text{Sr}/^{86}\text{Sr})_i - \epsilon_{\text{Nd}}(t)$ curve. However, this is not the case for the Keketuohai gabbro and the type I mafic dykes in the Habahe complex (Fig. 7a, 9), indicating that fluids were not the main agent of the metasomatism of the mantle source for these mafic rocks.

The fact that the primitive magmas that formed these samples were derived from a spinel-bearing mantle source means these melts will most likely have geochemical

compositions that are similar to MORB (Fig. 6; Salters and Stracke, 2004). However, the Keketuohai gabbros and type I mafic dykes are enriched in the LREE and Th (Fig. 6), suggesting that the metasomatism of the mantle source of these units was caused by melts, primarily as such melts can contain large amounts of Th and the LREE (Hermann et al., 2006). Melt enriched in the LREE and Th in melt can be generated by the breakdown of allanite or monazite, both of which represent the major hosts of the LREE and Th within subducted sediments (Fig. 10; Hermann et al., 2006; Todd et al., 2010). These samples have significantly negative Nb anomalies (Fig. 6), suggesting the presence of Nb-rich minerals, such as rutile within residual material after the melting of subducted sediments (Todd et al., 2010). In addition, these mafic rocks have positive $\epsilon_{\text{Hf}}(t)$ values (Fig. 7b) that are indicative of little modification of the Hf isotopic composition of their mantle source, suggesting that zircon may have been a residual phase within the subducted slab (Todd et al., 2010). The low Ba/La (Fig. 9) and high La/Yb and Th/Yb ratios (Fig. 10) of these samples indicate that the metasomatic agent involved in the formation of these units was more LREE enriched than LILE and HREE enriched, with the latter two groups of elements potentially retained by phengite and garnet, respectively (Hermann and Rubatto, 2009; Todd et al., 2010). The presence of residual K-bearing phengite can also account for the low K_2O and $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratios of these mafic rocks.

Here, we present a model for the interaction of the depleted mantle and melt derived from subducted sediments. The subducted sediments were assumed to have undergone melting to yield a residual assemblage of clinopyroxene (55%), garnet (15%), mica (30%), and trace monazite (0.05% to 0.3%), rutile (0.5%), and zircon (0.05%). Simple mixing between typical depleted mantle compositions and these subducted sediments derived

melts can account for the low Ba/La ratios and high Nd/Hf, Th/Yb, Th/La, and La/Yb ratios of the Keketuohai gabbro and type I mafic dykes (Fig. 9, 10, 11). In addition, the incorporation of melts with high Nd/Hf ratios can modify the Nd isotopic composition of the mantle wedge more significantly than the Hf isotopic composition of this mantle source, yielding a decoupling of the Nd and Hf isotopic systematics within the mantle source for these units (Fig. 7b). In conclusion, these mafic rocks were likely derived from the depleted lithospheric mantle wedge that was metasomatised by subducted-sediment-derived melts.

Figs 9, 10, and 11

5.2.2. Melting of subducted basaltic oceanic crust

The Habahe gabbroic samples have low initial $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{208}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios with high positive $\epsilon_{\text{Nd}}(t)$ and $\epsilon_{\text{Hf}}(t)$ values (Fig. 6), all of which are consistent with their formation from magmas derived from the depleted lithospheric mantle. These rocks contain relatively high concentrations of SiO_2 and MgO and have high $\text{Mg}^\#$ values (Fig. 5) that may result from fractional crystallisation and/or crystal accumulation processes. If this is true, the two major mineral phases of the Habahe gabbro, plagioclase and high Mg amphibole, must have played an important role in the fractional crystallisation. The removal of plagioclase will decrease Na_2O and CaO concentrations within the resultant melt, but this contrasts with the similar Na_2O and CaO concentrations of the Habahe gabbro and other mafic rocks in the study area (Fig. 5). In addition, the removal of plagioclase often generates significantly negative Eu anomaly within the resultant melt as a result of Eu preferentially partitioning into plagioclase. However, the Habahe gabbroic samples

have insignificantly negative Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.9\text{--}1.0$) that do not favour the fractional crystallisation of plagioclase. In addition, the MREE are more compatible within amphibole than are the LREE and the HREE (McKenzie and O’Nions, 1991), meaning that the accumulation of amphibole would yield concave downward REE patterns that are not observed for the Habahe gabbro. This means that the relatively high contents of SiO_2 and MgO within these gabbros cannot be the result of fractional crystallization and crystal accumulation. The high SiO_2 contents and the low FeO/MgO ratios of the Habahe gabbro are similar to those of high-Ca boninites (Fig. 5; Crawford et al., 1989) and high-Mg andesites (Fig. 5; Tatsumi et al. 2006). High-Ca boninites are thought to form from magmas generated by the partial melting of lherzolite fluxed by fluids at low pressure (<1 GPa; König et al., 2010). The low mobility of the MREE in fluids yields high-Ca boninites that commonly have U-shaped REE patterns (Fig. 6a; Crawford et al., 1989; König et al., 2010), which is not observed for the Habahe gabbros (Fig. 6a). Alternatively, the high SiO_2 concentrations with the high-Mg andesites of the Setouchi Volcanic Belt of Japan are thought to be related to the addition of silica in the form of subducted-slab-derived hydrous melt migrating to the mantle (Tatsumi, 2001), which may explain the high SiO_2 concentrations in the Habahe gabbros. The insignificant Pb anomalies (Fig. 6) and depleted Sr-Nd-Pb isotopic compositions of the Habahe gabbros (Fig. 7) indicate that subducted-sediment-derived melts did not make a major contribution to the metasomatism of the mantle source of these units. This again supports the involvement of a metasomatic agent in the form of melts derived from subducted oceanic crust with depleted isotopic features. The presence of abundant amphiboles within the Habahe gabbro suggests that the melting of this subducted oceanic crust occurred under water-saturated conditions (Klemme et al.,

2002; Klimm et al., 2008) and, with the resulting hydrous melts potentially adding silica to the mantle wedge, accounting for the high SiO₂ contents of the Habahe gabbros. The high La/Yb and Th/Yb ratios of these gabbroic samples (Fig. 10) also indicate that the LREE and Th were preferentially enriched relative to the HREE in the melt derived from the oceanic crust, suggesting that residual garnet within the subducted oceanic crust retained the latter (Klimm et al., 2008; Skora and Blundy., 2010). The Habahe gabbros also have high Ba/La ratios (Fig. 9), suggesting that Ba was released from the subducted oceanic crust together with the LREE. This indicates that phengite was not present during the melting of the subducted oceanic crust that has low concentrations of potassium (Fig. 9; Hermann and Rubatto, 2009; Kessel et al., 2005; Klimm et al., 2008). The presence of significantly negative Nb anomalies (Fig. 6) within the Habahe gabbros is suggestive of the presence of residual rutile within the subducted oceanic crust (Klimm et al., 2008). We modelled the melting of this subducted oceanic crust to yield a clinopyroxene, garnet, rutile, and allanite residue at a ratio of 50%:50%:0.5%:0.001%. The resulting melts have high Ba/La, Th/La, and La/Yb and low Nd/Hf ratios and are isotopically depleted, all of which are consistent with the geochemical and isotopic characteristics of the Habahe gabbros.

5.2.3. Dehydration of the subducted sediments

Type II mafic dykes within the Habahe complex have high positive $\epsilon_{\text{Nd}}(t)$ and $\epsilon_{\text{Hf}}(t)$ values (Fig. 7b), and are depleted in the LREE and Th (Fig. 6), indicative of derivation from a depleted mantle source. However, these dykes also have high (⁸⁷Sr/⁸⁶Sr)_i ratios (Fig. 7a), are enriched in Ba and U (Fig. 6), and have significantly positive Pb anomalies (Fig. 6), and have elevated (²⁰⁸Pb/²⁰⁴Pb)_t ratios relative to (²⁰⁶Pb/²⁰⁴Pb)_t (Fig. 7d). These features

are indicative of derivation from a mantle source that was metasomatised by agents from subducted sediments rather than the oceanic crust, primarily as the latter has low concentrations of Pb and has low $^{208}\text{Pb}/^{204}\text{Pb}$ ratios (Fig. 6; Elliott et al., 1997; Hickey-Vargas et al., 2008). The high $\epsilon_{\text{Nd}}(t)$ values and coupled Nd-Hf isotopic systematics of these dykes (Fig. 7b) indicate that the metasomatic agent was not Nd enriched, and their low La/Yb and Th/Yb ratios (Fig. 10) indicate that this agent did not mobilise the LREE and Th. Therefore, the metasomatic agent involved in the formation of these dykes was not a melt phase because of the high mobilities for the LREE and Th in melt. However, fluids derived from subducted sediments can efficiently mobilise the LILE at relatively low temperatures ($<600^\circ\text{C}$; Kessel et al., 2005; Spandler et al., 2007) and the LREEs can be retained by residual allanite and/or monazite (Hermann and Rubatto, 2009; Spandler et al., 2007). This suggests that these mafic dykes were derived from a depleted mantle source that was metasomatised by fluids released from subducted sediments.

5.3. Mechanism of melting and dehydration of the subducted slab

The ages of our mafic samples range from 409 ± 5 Ma (the Keketuohai gabbro; Cai et al., 2012) to 369 ± 3 Ma (the Habahe gabbro, this study), suggesting that the dehydration and melting of the subducted slab beneath the study area prevailed during the Devonian. The melting or dehydration of subducted sediments and/or oceanic crust is dependent on the surface temperature of the subducted slab; if the surface temperature is high enough then the slab can undergo melting, as is the case within the Sunda and Lesser Antilles arcs (Fig. 7; e.g., Woodhead et al., 2001). However, surface temperatures that are too cold to

melt the subducted slab mean that dehydration reactions will prevail (Pearce et al., 1999). This occurs in island arcs along the southwestern Pacific margin, such as the Mariana arc, the New Zealand arc and the Kermadec arc, in which the long-lived and cold Pacific oceanic crust is consumed. The high Th/Yb and La/Yb ratios of the studied mafic rocks suggest that garnet, instead of amphibole, was a major residual phase during the melting of subducted slab in the study area, primarily because garnet preferentially incorporates the HREE relative to amphibole (Rollinson, 1993 and references therein). This suggests that the subducted slab might have been melted at high pressure (≥ 2.5 GPa). The presence of abundant amphiboles within the Keketuohai and Habahe gabbros suggest that the melts from the subducted slab were volatiles enriched, indicating that the subducted basaltic oceanic crust and sediments may be melted under water-saturated condition. Solidus line for wet pelite and basalt is about 750 °C at 2.5 GPa and the temperature increases with pressure (Hermann and Spandler, 2008). All of this indicates that the melting of subducted sediments and basaltic oceanic crust was caused by the Devonian subduction of a hot slab beneath the Chinese Altai.

Melting of subducted slab can be caused by the breakoff of the subducted slab or by the subduction of an oceanic ridge. Slab breakoff generally occurs during the early stage of continent-continent or arc-continent collision, as recorded in the continental collision zones between India and Eurasia and between Arabia and Eurasia. A slab window could form as a result of tensile stress from the pull of the heavy oceanic lithosphere connected to the buoyant continental part of the plate (Davies and von Blanckenburg, 1995). Then, the asthenospheric mantle could upwell through the slab window, facilitating melting of the subducted slab and lithospheric mantle. However, no Paleozoic collisional event has been

identified in the Chinese Altai, and the high-T/low-P metamorphism at about 390 Ma strongly contradicts with a collisional event at this time. In addition, high-K magmatism would develop after the collision as occurred in the Tibetan plateau (Chung et al., 2005). This characteristics is not observed in the study area, as all of the Devonian mafic rocks in the Chinese Altai have low K_2O/Na_2O ratios (<0.6 ; Fig. 5f). This led to previous researchers to suggest that the extensive magmatism in the study area could be explained by the subduction of an oceanic ridge beneath the Chinese Altai during the Devonian (Cai et al., 2012; Sun et al., 2009; Windley et al., 2002). Melting of a subducted slab is a typical process during ridge subduction as exemplified by the subduction of the Kula-Farallon Ridge in southern Alaska (DeLong et al., 1978), the Cocos Ridge in Central America (Johnston and Thorkelson, 1997), the Nazca Ridge in Peru, and the Juan Fernandez Ridge in Chile (Bourdon et al., 2003). The melting of subducted slab in these cases can be attributed to two reasons. Firstly, the subducted slab was young and hot as it was located near the oceanic ridge (Cai et al., 2014; Defant and Drummond, 1990). Secondly, a slab window was developed because diverging oceanic slabs ceased to grow and the ridge progressively unzipped to form a gap (Thorkelson and Breitsprecher, 2005). This allow hot sub-slab asthenospheric mantle to ascend and fill the gap, promoting melting of the subducted slab (Thorkelson and Breitsprecher, 2005). The data present in this study reveal prevailed melting of subducted slab in the Chinese Altai during Devonian, which is consistent with such a ridge subduction model.

The tectonic setting of the Chinese Altai during the early to middle Paleozoic (pre-Devonian) was also hotly debated, primarily as a result of the Nd-Hf isotopic decoupling for the early to middle Paleozoic granites in this region. Some authors have suggested that

this region represents a Precambrian microcontinent, as evidenced by the negative $\epsilon_{\text{Nd}}(t)$ values for some granites in this area (e.g., Wang et al., 2009). In contrast, other research suggests that the Chinese Altai was an active continental margin or a magmatic arc, as evidenced by the predominantly positive $\epsilon_{\text{Hf}}(t)$ values for zircons from the Paleozoic granites in this area as well as for detrital zircons from metasedimentary rocks (Cai et al., 2011b; Sun et al., 2008; Xiao et al., 2008; Yu et al., in press). Yu et al. (in press) demonstrated that Nd-Hf isotopic decoupling was prevailed for the whole crust of the Chinese Altai, which cannot be explained by the “zircon effect” or “garnet effect”. This study also demonstrate that this Nd-Hf isotopic decoupling was possibly inherited from a metasomatised mantle source. Devonian melting of subducted slab prevailed in the Chinese Altai, meaning that the extensive metasomatism associated with this melting could have resulted in widespread Nd-Hf isotopic decoupling (Fig. 7b; Xie et al., 2016). The addition of volatile-rich agents from the subducted slab into the lithospheric mantle would have promoted melting of the lithospheric mantle (Pearce et al., 1999), causing significant volumes of mafic magma with decoupled Nd-Hf isotopic characteristics to be emplaced into the crust. The Devonian granites in the Chinese Altai may have inherited the decoupled Nd-Hf isotopic characteristics as a result of remelting of solidified mafic magma under a high geothermal gradient. This suggests that the Chinese Altai represents a magmatic arc that records the Devonian subduction of an oceanic ridge. The addition of hydrous melts and fluids from the subducted slab may have significantly modified the geochemical composition of the lithospheric mantle as well as enhancing partial melting of the lithospheric mantle. Ridge subduction was possibly responsible for the peak of magmatism

and significant continental growth in the Chinese Altai and southwestern Mongolian collage as a whole in the Devonian.

6. Conclusions

1. The Habahe gabbro has an age of 369 Ma and was formed from magmas generated by the partial melting of the depleted lithospheric mantle that was metasomatised by hydrous melts derived from subducted oceanic crust.

2. The primary magmas of the Devonian Keketuohai gabbro and type I mafic dykes within the Habahe complex were derived from the depleted lithospheric mantle that was metasomatised by melts derived from subducted sediments. In comparison, the mantle source for the type II mafic dykes was metasomatised by fluids derived from subducted sediments.

3. The melting of subducted slab in this region indicates that the Chinese Altai records the Devonian subduction of an oceanic ridge. Hydrous melts and fluids promoted the melting of the lithospheric mantle that was responsible for the intense magmatism, significant continental growth and high temperature metamorphism.

7. Acknowledgements

We are grateful for the comments from Dr. Andrew Kerr and two anonymous reviewers, which significantly helped us to improve the quality of this paper. H. Y. Geng, J. Wong, L. Zhang, Q.S. Zhao, Y.Q. Wang, X. Fu and X.L. Tu are thanked for their analytical assistance. This research was supported by Hong Kong RGC grants (HKU704013P, 17303415), National Science Foundation of China (41273048), the Major Basic Research Project of the Ministry of Science and Technology of China (Grant: 2014CB440801), Hong Kong–Germany Joint Research Scheme (G-HKU706115), and a HKU CRCG grant (201511159199). This is a contribution to IGCP 592 from the Joint Laboratory of Chemical Geodynamics between HKU and CAS (Guangzhou Institute of Geochemistry).

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Figure captions

Fig. 1. (a) Simplified regional map showing the known extent of the CAO. (b) Geological map of the tectonic framework of the southwestern Mongolia collage system (revised from Sun et al., 2008 and Xiao et al., 2015).

Fig. 2. Simplified geological maps of the (a) Habahe complex, (b) Keketuohai complex (modified after Cai et al., 2012), showing rock types and sample locations.

Fig. 3. Field photographs showing (a) magma mixing between the Habahe gabbro and the surrounding plagiogranite, and (b) the accumulation of plagioclase and mafic minerals within the Keketuohai gabbro. Petrographic characteristics of (c) Habahe gabbro and (d) Keketuohai gabbro. Amp = amphibole, Pl = plagioclase, Ilm = ilmenite.

Fig. 4. (a) Concordia diagram showing the weighted mean ages of zircons from the Habahe gabbro analysed by ion microprobe during this study with uncertainties at the 1 σ ; (b) Evolution of Hf isotopic compositions of zircons from the Habahe gabbro.

Fig. 5. (a) SiO₂ vs. Mg[#], (b) Al₂O₃ vs. Mg[#], (c) CaO vs. Mg[#], (d) FeO^T/MgO vs. SiO₂ diagrams with dashed boundaries for tholeiitic and calc-alkaline fields from Miyashiro (1974). (e) K₂O vs. SiO₂ and (f) K₂O vs. Na₂O. Rock suit classifications are according to Rickwood (1989). The concentrations of peridotite melts are from Stern and Kilian (1996), the MORB compositions are from Salters and Stracke (2004). Data of high-Ca boninites and high-Mg andesites are from the Troodos complex on Cyprus (König et al., 2010) and south western

Janpan (Tatsumi et al., 2006) These diagrams also include previously published data for the Keketuohai gabbro and the Habahe mafic dykes from Cai et al. (2012) and Cai et al. (2010), as well as for the Kuerti mafic rocks and Kuwei gabbro from Xu et al. (2003) and Ye et al. (2015). Pl = plagioclase, I = type I mafic dyke; II = type II mafic dyke.

Fig. 6. Chondrite-normalised REE patterns and primitive mantle-normalised trace elements variation diagrams for the Keketuohai gabbro, Habahe gabbro and Habahe mafic dyke. Chondrite and PM normalisation factors from Sun and McDonough (1989). The dashed line represents the average composition of MORB from Salters and Stracke (2004). The previously published data of the Keketuohai gabbro and Habahe mafic dykes are from Cai et al. (2012) and Cai et al. (2010). Data of high-Ca boninites are from the Troodos complex on Cyprus (König et al., 2010)

Fig. 7. (a) $\epsilon_{\text{Nd}}(t)$ vs. $(^{87}\text{Sr}/^{86}\text{Sr})_i$ diagram for samples for the study area, including the depleted mantle (DM) composition of Salters and Stracke (2004). (b) $\epsilon_{\text{Hf}}(t)$ vs. $\epsilon_{\text{Nd}}(t)$ diagram for samples from the study area along with the compositions of subducted sediments represented by the average composition of the Habahe Group (HG) that is similar to GLOSS (Long et al. 2012; Plank and Langmuir, 1998). Since no whole-rock Hf isotope is available, the Habahe Group is represented by a grey vertical bar at $\epsilon_{\text{Nd}}(t) = -4.3$. The subducted sediments were melted at eclogite facies (3Gpa) conditions with 5% partial melting degree. This diagram uses partition coefficients from Johnson (1998), Adam and Green (2006), Skora and Blundy (2010), Foley et al. (2000), and Rubatto and Herman (2007)

along with trace residual phases that consist of monazite, rutile, and zircon at a 0.05%:0.5%:0.05% ratio. The terrestrial array is from Vervoort et al. (2011) ($\epsilon_{\text{Hf}} = 1.55 * \epsilon_{\text{Nd}} + 1.21$) (c) $(^{207}\text{Pb}/^{204}\text{Pb})_{\text{t}}$ vs. $(^{206}\text{Pb}/^{204}\text{Pb})_{\text{t}}$ diagram and (d) $(^{208}\text{Pb}/^{204}\text{Pb})_{\text{t}}$ vs. $(^{206}\text{Pb}/^{204}\text{Pb})_{\text{t}}$ diagram. Both (c) and (d) include the NHRL of Hart (1984) along with MORB (DMM) and OIB data from Salters and Stracke (2004) and the PetDB database (<http://www.petdb.org>). “Cold arc” includes data from the Mariana, Kermadec and New Zealand arcs. “Hot arc” is represented by the Sunda, Banda, Lesser Antilles and Luzon arcs. The source of data for the Habahe mafic dykes, the Keketuohai gabbros, and the Kuerti mafic rocks are as in Fig. 5, and Sr-Nd-Hf isotopic data for granites within the Chinese Altai are from Cai et al. (2011b), Wang et al. (2009), Yu et al. (in press), and Yuan et al. (2007). CA = Chinese Altai.

Fig. 8. Nb/Y vs. Zr/Y diagram. OIB = ocean island basalt except for data from Iceland. Data of OIB, Indian Ocean MORB and Pacific Ocean MORB are from the PetDB database (<http://www.petdb.org>). DM= depleted mantle composition of Salters and Stracke (2004), PM = primitive mantle composition of Sun and McDonough (1989), EN = enriched mantle components of Condie (2003) and all other data source are identical to those in Fig. 7.

Fig. 9. (a) $\epsilon_{\text{Nd}}(\text{t})$ vs. Ba/La and (b) Th/La vs. Ba/La diagrams. Melting conditions of the subducted sediments and basaltic oceanic crust and data source are the same with Fig. 7. The compositions of fluid are calculated using the partition coefficients of Spandler et al. (2007) at 600 °C and 2.2 GPa.

Fig. 10. (a) $\epsilon_{\text{Nd}}(t)$ vs. La/Yb and (b) $\epsilon_{\text{Nd}}(t)$ vs. Th/Yb diagrams. Melting conditions of the subducted sediments are identical to those shown in Fig. 6 but with different portions of residual monazite (0.05% and 0.3%). The N-MORB composition of Sun and McDonough (1989) was adopted for the melting of subducted basaltic oceanic crust that is melted with a residue containing clinopyroxene, garnet, rutile, allanite at a ratio of 50%:50%:0.5%:0.001% during 1% partial melting. This diagram uses partition coefficients from Johnson (1998), Foley et al. (2000), Klemme et al. (2002), and Klimm et al. (2008) with data source as in Fig. 7.

Fig. 11 (a) $\epsilon_{\text{Nd}}(t)$ vs. Nd/Hf and (b) $\epsilon_{\text{Hf}}(t)$ vs. Nd/Hf diagrams for samples in this study and the melting conditions of the subducted sediments and basaltic oceanic crust and data source as in Fig. 7.

Table 1, Representative major and trace elements of the gabbros from the Habahe complex and the Keketuohai complex

Sample	ALT14-	ALT14-	ALT14-	ALT14-	ALT14-	ALT14-	ALT14-	ALT14-	ALT14-	ALT14-	ALT14-	ALT14-	ALT14-
No	56	57	58	59	60	125	126	127	128	130	131	134	135
Type	HBG	HBG	HBG	HBG	HBG	KKG	KKG	KKG	KKG	KKG	KKG	KKG	KKG
SiO ₂	53.5	52.7	52.7	53.8	52.1	47.4	46.1	46.9	46.4	47.9	45.4	43.6	43.9
TiO ₂	0.48	0.48	0.48	0.48	0.55	0.98	1.69	1.28	1.09	0.94	1.32	1.30	1.50
Al ₂ O ₃	9.99	11.3	10.2	10.6	10.5	16.2	16.2	16.6	13.3	19.2	12.5	20.7	18.2
Fe ₂ O ₃ ^T	7.68	7.98	7.87	7.72	7.82	12.7	15.2	13.5	15.1	10.7	16.5	13.1	15.7
MnO	0.16	0.15	0.15	0.16	0.16	0.20	0.17	0.18	0.20	0.14	0.22	0.15	0.20
MgO	12.3	11.9	12.6	11.7	11.5	9.31	7.24	8.16	11.0	7.05	11.5	7.03	6.60
CaO	12.9	11.9	12.0	12.6	13.2	11.1	11.4	10.6	10.6	12.3	10.3	11.4	11.6
Na ₂ O	1.36	1.71	1.50	1.37	1.39	0.79	0.89	1.00	0.75	0.94	0.70	0.68	0.63

K ₂ O	0.52	0.74	0.75	0.53	0.62	0.17	0.15	0.20	0.16	0.16	0.17	0.10	0.11
P ₂ O ₅	0.04	0.05	0.05	0.04	0.05	0.03	0.15	0.11	0.02	0.05	0.08	0.06	0.38
LOI	1.01	1.13	1.26	1.06	1.59	0.94	0.59	0.88	0.80	0.65	0.86	1.44	0.78
Total	100	100	99.6	100	99.4	99.8	99.9	99.3	99.4	100.0	99.6	99.5	99.6
Mg [#]	0.79	0.78	0.79	0.78	0.78	0.63	0.53	0.59	0.63	0.61	0.62	0.56	0.50
Sc	74.6	62.6	66.2	69.5	62.0	38.2	43.7	38.3	48.1	34.6	49.1	24.2	41.2
Ti	2721	2648	2658	2627	3050	5462	9930	7358	6324	5301	7375	7658	8494
Cr	965	676	834	723	603	283	18.1	43.1	460	69.2	536	250	108.7
Mn	1274	1154	1174	1219	1185	1399	1318	1335	1540	1009	1652	1134	1546
Co	75.5	66.1	64.7	67.7	74.6	95.1	99.9	107	96.6	86.0	87.6	91.8	84.6
Ni	90.7	96.5	106	77.5	85.5	75.6	39.6	54.5	123	41.6	97.9	157	41.6
Cu	16.5	12.0	9.31	11.8	15.0	59.3	47.3	37.3	37.5	38.6	45.4	75.4	53.9
Zn	79.0	74.5	71.1	71.1	63.1	84.7	92.1	93.6	101	69.9	94.0	96.5	105
Ga	10.4	10.3	10.0	10.9	10.7	16.5	18.3	17.0	15.5	18.2	15.9	22.6	21.1
Ge	2.38	2.28	2.25	2.16	2.25	2.88	3.31	3.12	3.55	2.53	3.93	2.81	3.45
Rb	6.30	9.71	9.88	6.06	8.09	2.59	1.02	2.56	1.40	3.03	1.87	0.80	0.83
Sr	184	169	152	198	182	197	184	177	112	217	68.4	237	198
Y	10.7	10.6	10.5	10.1	12.0	5.89	10.1	8.46	6.94	7.82	8.48	2.477	10.6
Zr	59.9	25.7	29.3	22.2	35.4	18.8	21.1	24.6	17.7	27.7	24.0	25.2	11.8
Nb	0.79	0.58	0.58	0.67	0.77	1.13	3.01	2.60	0.90	1.45	1.39	2.93	1.97
Ba	104	123	138	129	102	20.8	45.5	44.7	9.65	31.9	21.6	13.7	28.0
La	4.34	3.97	4.11	4.19	4.60	2.32	4.37	4.04	2.08	4.13	3.14	3.10	5.23
Ce	10.4	10.2	10.8	10.2	12.5	5.43	10.5	9.34	5.30	9.99	7.96	5.42	12.6
Pr	1.48	1.51	1.58	1.37	1.85	0.76	1.54	1.31	0.78	1.36	1.12	0.61	1.79
Nd	6.44	7.52	7.65	6.17	8.85	3.43	7.28	5.77	3.59	5.83	5.01	2.33	8.38
Sm	1.66	2.07	2.04	1.67	2.37	0.90	1.86	1.38	1.03	1.33	1.33	0.40	2.06
Eu	0.58	0.65	0.63	0.55	0.71	0.42	0.70	0.56	0.39	0.56	0.46	0.43	0.67

Gd	1.87	2.14	2.11	1.77	2.44	0.97	1.98	1.47	1.15	1.35	1.48	0.40	2.18
Tb	0.32	0.35	0.35	0.32	0.40	0.18	0.33	0.25	0.21	0.24	0.26	0.06	0.35
Dy	2.02	2.08	2.08	1.97	2.38	1.12	2.01	1.60	1.35	1.48	1.65	0.41	2.07
Ho	0.44	0.42	0.43	0.41	0.49	0.25	0.42	0.34	0.29	0.31	0.35	0.10	0.43
Er	1.19	1.14	1.13	1.14	1.30	0.70	1.14	0.96	0.80	0.89	0.96	0.31	1.13
Tm	0.17	0.16	0.16	0.16	0.19	0.11	0.17	0.15	0.12	0.13	0.14	0.06	0.15
Yb	1.09	1.02	1.03	1.05	1.17	0.69	1.06	0.97	0.77	0.87	0.92	0.41	0.96
Lu	0.17	0.15	0.15	0.16	0.18	0.11	0.17	0.15	0.12	0.13	0.14	0.07	0.15
Hf	1.74	0.93	1.05	0.92	1.24	0.57	0.66	0.73	0.59	0.93	0.80	0.72	0.42
Pb	1.63	1.23	1.11	3.57	1.54	1.86	2.44	2.38	1.29	2.00	0.94	1.95	2.07
Th	0.94	0.79	0.82	0.91	1.22	0.59	0.69	0.81	0.46	0.88	1.08	0.14	0.55
U	0.24	0.17	0.18	0.24	0.31	0.17	0.32	0.31	0.23	0.27	0.46	0.07	0.25

Rock type abbreviation: HBG = Habahe gabbro; KKG = Keketuohai gabbro; $Mg^{\#} = 100 * Mg / (Mg + Fe^{2+})$, assuming

$Fe^{2+} / Fe^{total} = 0.85$; LOI = loss of ignition; $Fe_2O_3^T$ = total Fe oxides as Fe_2O_3

Table 2. Whole-rock Sr-Nd isotopes of the mafic rocks of the Chinese Altai

Sample	Type	age(Ma)	$^{87}\text{Rb}/^{86}\text{Sr}^a$	$^{87}\text{Sr}/^{86}\text{Sr} (\pm 1s)$	$(^{87}\text{Sr}/^{86}\text{Sr})_i$	$^{147}\text{Sm}/^{144}\text{Nd}^a$	$^{143}\text{Nd}/^{144}\text{Nd}(\pm 1s)$	$(^{143}\text{Nd}/^{144}\text{Nd})_i$	ϵ_{Nd}^b
Habahe									
complex									
ALT14-56	HBG	369	0.099	0.703946±10	0.7034	0.1562	0.512937±9	0.512538	7.8
ALT14-57	HBG	369	0.166	0.704312±9	0.7034	0.1667	0.512954±9	0.512528	7.6
ALT14-58	HBG	369	0.188	0.704424±12	0.7034	0.1610	0.512938±11	0.512527	7.5
ALT14-59	HBG	369	0.089	0.704021±13	0.7036	0.1634	0.512937±9	0.512520	7.4
ALT14-60	HBG	369	0.129	0.704201±9	0.7035	0.1618	0.512943±8	0.512530	7.6
HB12	MDI	376	0.019	0.706939±14	0.7068	0.1910	0.512677±13	0.512208	1.0
HB14	MDI	376	0.026	0.706975±14	0.7068	0.1960	0.512672±8	0.512191	0.7
HB15	MDI	376	0.021	0.707093±11	0.7070	0.1940	0.512679±10	0.512203	0.9
HB16	MDI	376	0.023	0.706726±9	0.7066	0.1724	0.512664±10	0.512224	1.7
HB17	MDI	376	0.026	0.706728±13	0.7066	0.1833	0.512644±10	0.512176	0.8
BH5-1	MDI	376	0.003	0.705698±22	0.7057	0.1899	0.512715±6	0.512249	1.8
BH5-3*	MDI	376	0.005	0.705683±12	0.7057	0.1774	0.512804±12	0.512371	4.2
BH5-4*	MDI	376	0.002	0.704896±12	0.7049	0.1843	0.512885±12	0.512435	5.4
BH24*	MDI	376	0.046	0.705439±13	0.7052	0.1586	0.512716±11	0.512327	3.4
BH1-7*	MDII	376	0.044	0.705040±14	0.7048	0.2012	0.513062±11	0.512571	8.1
BH1-9*	MDII	376	0.036	0.704249±14	0.7040	0.1994	0.513035±13	0.512548	7.6
Keketuohai									
complex									
ALT14-134	KKG	409	0.010	0.707749±11	0.7077	0.1037	0.512270±9	0.511992	-2.3
ALT14-135	KKG	409	0.012	0.7082693±12	0.7082	0.1490	0.512385±11	0.511986	-2.4
ALT14-136	KKG	409	0.017	0.707574±11	0.7075	0.1795	0.512466±9	0.511985	-2.5
ALT14-137	KKG	409	0.038	0.707689±11	0.7075	0.1649	0.512439±7	0.511997	-2.2
ALT14-125	KKG	409	0.038	0.707325±10	0.7071	0.1589	0.512464±12	0.512038	-1.4
ALT14-126	KKG	409	0.016	0.70788±11	0.7078	0.1540	0.512432±9	0.512020	-1.8

ALT14-128	KKG	409	0.036	0.707485±14	0.7073	0.1737	0.512505±9	0.512040	-1.4
ALT14-129	KKG	409	0.026	0.707204±11	0.7071	0.1630	0.512519±14	0.512082	-0.6
ALT14-131	KKG	409	0.079	0.707451±11	0.7070	0.1603	0.512463±10	0.512034	-1.5
ALT14-132	KKG	409	0.045	0.707133±12	0.7069	0.1602	0.512491±10	0.512062	-1.0

Note

^a $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ are calculated using whole-rock Rb, Sr, Sm and Nd contents in supplementary appendix 3 and Cai et al. (2010).

^b $\epsilon_{\text{Nd}}(t) = [(^{143}\text{Nd}/^{144}\text{Nd})_{\text{S}} / (^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} - 1] \times 10000$.

In the calculation, $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} = 0.512638$, $(^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}} = 0.1967$, $\lambda = 6.54 \times 10^{-6}$,

Rock type abbreviation: HBG = Habahe gabbro; MDI = type I mafic dyke; MDII = type II mafic dyke; KKG = Keketuohai gabbro. Ages of the mafic dykes and Keketuohai gabbros were from Cai et al. (2010, 2012).

^{*}The Sr-Nd data of type II mafic dykes and some type I mafic dykes of the Habahe complex is from Cai et al. (2010).