Whole-rock Nd-Hf isotopic study of I-type and peraluminous granitic rocks
from the Chinese Altai: constraints on the nature of the lower crust and
tectonic setting

Yang Yu^a, Min Sun^{a*}, Xiaoping Long^b, Pengfei Li^a, Guochun Zhao^a, Alfred Kröner^c,
Arnaud Broussolle^a, Jinhui Yang^d

Abstract

The nature of the lower crust and tectonic setting of the Chinese Altai in the early to middle Paleozoic are still hotly debated. Decoupling between zircon Hf and whole-rock Nd isotopic systems for granites results in different interpretations for the above issues. In order to solve the problem, whole-rock Nd-Hf isotopic analyses were conducted on representative early to middle Paleozoic I-type granite and strongly peraluminous granites and rhyolites from the Chinese Altai. The I-type granites show metaluminous to weakly peraluminous feature and have $\epsilon_{\text{Nd}}(t)$ values ranging from -2.2 to +0.8 and $\epsilon_{\text{Hf}}(t)$ from +3.9 to +12.9, respectively. The strongly peraluminous granites and rhyolites have similar $\epsilon_{\text{Nd}}(t)$ and $\epsilon_{\text{Hf}}(t)$ values ranging from -3.0 to +1.7 and from +2.1 to +10.4, respectively. All samples plot above the Terrestrial Array on Nd-Hf isotopic diagram, indicating significant Nd-Hf isotopic decoupling in the magma sources. These samples show flatten HREE pattern and have Lu/Hf ratios similar to the average crust, suggesting that Nd-Hf isotopic decoupling was not originated from an ancient basement with elevated Lu/Hf ratios. The observed isotopic decoupling is similar to those modern island arcs, such as the Lesser Antilles and Sunda, where Nd selectively enriched over Hf due to metasomatism in the mantle wedge and consequently resulted in decoupling between the Sm-Nd and Lu-Hf isotopic systems. Our results, combined with the available data, show that prolonged subduction and crust-mantle interaction caused the Nd-Hf isotopic decoupling in the lithospheric mantle beneath the Chinese Altai. The crust of the Chinese Altai was extracted from the lithospheric mantle and inherited the Nd-Hf isotopic decoupling feature. Therefore, the Hf, rather than Nd, isotopic data more faithfully reflect

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39 the nature of the lower crust that was quite juvenile in the Paleozoic, and the Chinese
40 Altai represents an early Paleozoic magmatic arc possibly built near western Mongolia.

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42 **Key Words:** Nd-Hf decoupling; Chinese Altai; Central Asian Orogenic Belt;
43 continental growth; accretionary orogeny.

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1. Introduction

The Central Asian Orogenic Belt (CAOB) is the most important site for voluminous continental growth in the Phanerozoic, as evidenced by the widespread positive whole-rock $\epsilon_{\text{Nd}}(t)$ values for granitoid rocks (e.g. Jahn et al., 2000a, 2004; Kröner et al., 2014). In recent years, a large number of in-situ zircon Hf isotopic data have also been obtained for sedimentary rocks and some granites from the CAOB. Despite the large range in initial Hf isotopic compositions of zircon, the results mirror the whole-rock Nd isotopic data in most cases, such as the Lake zone of Mongolia, the Kazakhstan Arc and the Gorny Altai of Russia (Kröner et al., 2014; Kruk et al., 2011; Chen et al., 2015), illustrating significantly continental growth.

However, the inference of crustal growth has become problematic in the Chinese Altai, a key area in the southwestern CAOB (Fig. 1), where decoupling between whole-rock Nd and Hf-in-zircon isotopic data has been observed in early to middle Paleozoic granitoid rocks (478 – 380 Ma) (e.g. Cai et al., 2011a, 2011b, 2011c; Sun et al., 2008; Wang et al., 2009; Yuan et al., 2007). These granitoids occupy about 40% of the entire area, their $\epsilon_{\text{Nd}}(t)$ values range from -4 to +5 (Chen and Jahn, 2002; Liu et al., 2012; Wang et al., 2009), but their zircon $\epsilon_{\text{Hf}}(t)$ values are significantly higher, mostly in the range from +2 to +9 (Cai et al., 2011b, 2011c; Sun et al., 2008;). This isotopic decoupling has led to contrasting interpretations for the tectonic significance of the Chinese Altai, for example, some authors envisage the region as a Precambrian microcontinent derived from eastern Gondwana, mainly based on the negative $\epsilon_{\text{Nd}}(t)$ values for some granitoids (e.g. Wang et

al., 2006, 2009). On the other hand, the Chinese Altai is considered by other researchers as an active continental margin, mainly based on the predominant positive zircon $\epsilon_{\text{Hf}}(t)$ values for Paleozoic granites as well as for the detrital zircons from metasedimentary rocks (Cai et al., 2011b; Long et al., 2010; Sun et al., 2008; Xiao et al., 2008). People may argue that zircon Hf isotope sometimes might not represent whole-rock Hf isotopic image because of possible disequilibrium between zircon and whole-rock (Farina et al., 2014). Otherwise, the decoupling might also stem from metasomatism. This is because $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of whole-rock samples are more easily disturbed by metasomatism than zircon $^{176}\text{Hf}/^{177}\text{Hf}$ ratios since Hf is nearly immobile while Nd is more mobile in a fluid phase (Hermann and Rubatto, 2009; Pearce et al., 1999). Alternatively, isobaric correction for ^{176}Yb may result in the problem as well, because in-situ analysis of zircon does not involve chemical separation of Yb from Hf. As pointed out by Fisher et al. (2014), a 0.1% inaccuracy in the correction of $^{176}\text{Yb}/^{177}\text{Hf}$ will result in a miscalculation of the $^{176}\text{Hf}/^{177}\text{Hf}$ value by about 0.000050 ($\sim 2 \epsilon_{\text{Hf}}$ units) for zircon crystals with $^{176}\text{Yb}/^{177}\text{Hf}$ as low as 0.05 and ~ 0.000150 ($\sim 5 \epsilon_{\text{Hf}}$ units) for zircon crystals with $^{176}\text{Yb}/^{177}\text{Hf}$ of 0.15.).

In order to settle the debate, decoupling between Hf and Nd isotope systems for the Chinese Altai needs to be verified. In this paper, we present whole-rock Nd-Hf isotopic data from representative early to middle Paleozoic granitic rocks, trying to decipher the nature of the lower crust and tectonic setting of the Chinese Altai.

Fig.1 about here

2. Geological background

The Central Asian Orogenic Belt (CAOB) is the largest accretionary orogenic belt on Earth, extending 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; Zonenshain et al., 1990; Mossakovsky et al., 1993; Şengör et al., 1993; Jahn et al., 2000a; Şengör and Natal'in, 2004; Windley et al., 2007). The northwest-trending Altai Orogenic Belt, a major part of the CAO, extends for more than 2500 km from eastern Kazakhstan via Russia, through northwest China to western and southern Mongolia. The Chinese Altai is the southern part of the Altai-Mongolia terrane and is located in the Xinjiang Uygur Autonomous Region of China (Fig. 1). The study area is situated between the Mongolian Lake Zone and Tuva-Mongolia Massif in the northeast, the Russian Gorny Altai in the north and the Kazakhstan-Junggar block in the south (Badarch et al., 2002; Windley et al., 2002; Xiao et al., 2004; 2015). Voluminous granitoid rocks and NW-trending fault system are developed in the Chinese Altai (Fig. 1b). The geology of the Chinese Altai can be broadly divided into four units, based on distinctive lithological assemblages and metamorphism. The North Altai Domain (shown as N in Fig. 1b, unit 1 of Windley et al., 2002) consists of Late Devonian-early Carboniferous neritic clastic sedimentary rocks, roughly coeval island arc volcanic rocks, and limestones, mostly metamorphosed at sub-greenschist-facies. The Central Chinese Altai Domain (shown as C in Fig. 1b, units 2 and 3 of Windley et al., 2002) occupies the major part of this region with widespread sedimentary rocks termed Habahe Group, which is considered to be the oldest sedimentary sequence and mainly composed of a thick sequence (~6000 m) of shale, phyllite and schist (GCRSX, 1981; He et al., 1990; Li et al., 2006; Windley et al., 2002).

Detrital zircons from these rocks yielded $^{206}\text{Pb}/^{238}\text{U}$ ages predominantly between 540 and 460 Ma and mostly gave positive $\varepsilon_{\text{Hf}}(t)$ values, suggesting that their sedimentary provenance was on an active margin in the early Paleozoic (Long et al., 2007, 2010). Zircons from local high-grade banded paragneisses are predominantly 528 to 466 Ma old, comparable to detrital zircon ages of the Habahe Group, and thus they are considered to be the high-grade equivalents of the Habahe Group (Sun et al., 2008). The Qiongkuer Domain (shown as Q in Fig. 1b, Terrane 4 of Windley et al., 2002) consists of late Silurian to Early Devonian island arc-type volcanic and pyroclastic rocks (Kangbutiebao Formation) and an overlying Middle Devonian turbiditic sandstone-shale sequence (Altai Formation), considered as fore-arc basin deposits (Windley et al., 2002). The South Altai Domain (unit 5 of Windley et al., 2002 and “S” in Fig. 1b) includes a sequence of Devonian sediments and late Carboniferous volcanoclastic rocks, with metamorphism at greenschist- to amphibolite-facies. Two ophiolitic mélanges occur in the Kurti and Alegedaiyi areas of the southern Chinese Altai (Wong et al., 2010; Xu et al., 2003). There are also mafic rocks, including gabbros and mafic dikes, exposed near the Keketuohai and Habahe areas (Fig. 1b; Cai et al., 2010, 2012). Further south, the Chinese Altai is separated from the Junggar terrane by the Erqis Fault, one of the large-scale (1000 km) sinistral strike-slip faults in Central Asia (Fig. 1b).

3. Sample description

Samples of this study were collected from two geologic domains: the Hanasi batholith (398±5 Ma), Hemu batholith (412±7 Ma), Jiadengyu rhyolite (445±2 Ma), Kuwei

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4 134 batholith (419 ± 9) in the Central Altai Domain and the Altay batholith (368 ± 13 Ma),
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6 135 Kezgar pluton (411 ± 11 Ma) and Kurmutu pluton (421 ± 6 Ma) in the Qiongkuer Domain
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9 136 (Fig.1b). Zircon ages from these rock bodies were reported by Cai et al. (2011b, 2011c),
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14 138 *3.1. Samples from the Central Altai Domain*

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20 140 The Hanasi batholith covers more than 200 km^2 and intruded into the Habahe Group
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22 141 (Fig. 2a). Our samples show porphyritic texture and are mainly composed of quartz (15 –
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24 142 20%), plagioclase (40 – 45%), biotite (15 – 25%) and amphibole (10 – 15%) with
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27 143 accessory minerals (<5%) such as magnetite and zircon (Fig. 3a).
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30 144 The Hemu batholith also intruded into the Habahe Group rocks (Fig. 2a) and consists
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33 145 of quartz (25 – 30%), K-feldspar (30 – 35%), plagioclase (20 – 30%), biotite (5 – 10%)
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35 146 and muscovite (5 – 10%) with accessory minerals such as tourmaline and zircon (Fig. 3b).
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38 147 The Jiadengyu rhyolite was firstly described by Wang et al. (2011), and samples were
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41 148 collected along the Kanasi-Jiadengyu road (Fig.1). They are porphyritic with euhedral
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44 149 phenocrysts (<15%), predominantly composed of plagioclase, quartz and minor biotite
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46 150 and felsic matrix (>85%).
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49 151 The Kuwei batholith samples (Fig. 2b) are medium-grained biotite granites, mainly
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51 152 composed of plagioclase (20 – 25%), K-feldspar (20 – 30%), quartz (25 – 30%), biotite
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54 153 (5 – 15%) and muscovite (5 – 10%), with minor amounts of accessory minerals such as
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56 154 zircon, Fe-Ti oxides, apatite, tourmaline, and titanite. The samples show porphyritic
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3.2. Samples from the Qiongkuer Domain

The Kurmutu pluton is exposed near Kurmutu Village, 10 km northwest of Altay city (Fig. 2b). The samples are massive biotite granites, composed of plagioclase (~45%), quartz (~20%), biotite (~20%), K-feldspar (~10%) and minor accessory minerals such as zircon and Fe-Ti oxides (Fig. 3d).

The Altay batholith occurs about 8 km south of the Altay city (Fig. 2b) and is weakly foliated. The mineral assemblage consists mainly of plagioclase (~40%), quartz (~35%), biotite (~10%), K-feldspar (~20%), muscovite (~10%) and accessory minerals including magnetite, apatite and zircon (Fig. 3e).

The Kezgar pluton outcrops about 30 km southeast of Altay city (Fig. 2c) and is composed of leucogranites. The samples of this study are weakly foliated and composed of plagioclase (~30%), K-feldspar (~20%), quartz (~35%), biotite (<10%), muscovite (~10%) and accessory minerals such as magnetite, apatite and zircon (Fig. 3f).

Figs. 2 and 3 about here

4. Analytical methods

Major element compositions were determined on fused glass beads by X-ray fluorescence (XRF) spectroscopy. We followed the procedures described by Norrish and Hutton (1969) for matrix correction. The analytical errors are about 1–2% for SiO₂, Al₂O₃, TiO₂, Fe₂O₃ and MgO and ≤ 3% for other oxides. Trace element analyses were

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4 177 carried out in Guangzhou Institute of Geochemistry, Chinese Academy of Sciences, by
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6 178 inductively coupled plasma-mass spectrometry (ICP-MS) after acid digestion of samples
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9 179 in high-pressure Teflon vessels, and the detailed procedures were described by Li et al.
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11 180 (2006). The USGS and Chinese National standards AGV-2, GSR-1, GSR-2, MRG-1,
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14 181 BCR-1, W-2 and G-2 were used for calibrating element concentrations of the analyzed
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16 182 samples. Analytical precisions for REE and other incompatible element analyses are
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22 184 Neodymium and Hf isotopic analyses were conducted in the Institute of Geology and
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24 185 Geophysics, Chinese Academy of Sciences, Beijing. Nd and Hf isotopic ratios were
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27 186 measured on a subset of whole-rock samples using a Micromass Isoprobe multicollector
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29 187 ICP-MS. All chemical preparations were performed on the specially designed class-100
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32 188 work benches inside a class-1000 clean laboratory. Powder of a rock was weighted (about
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34 189 100mg) into a 7mL round bottom SavillexTM Teflon screw-top capsule, mixed
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37 190 concentrate HF-HNO₃-HClO₄ (1.5mL–1.5mL–0.01 mL) was added, and gently heated up
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39 191 on a hotplate for dryness. Then mixed concentrate HF-HNO₃ (2mL–1mL) was added
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42 192 again, the capsule screw-top cap was closed and the capsule was placed into a steel-
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44 193 jacketed bomb for 120 hours in a furnace under 140 °C to ensure resistant mineral phases
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46 194 such as zircon and garnet were totally dissolved. After cooling, the capsule was taken out
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49 195 and the surface was cleaned with purified water. Then, the capsule was opened and gently
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51 196 evaporated. After dryness, 1mL of 6M HCl was added to the residue and dried, this
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54 197 procedure was then repeated. Finally the residue was dissolved in 1mL of 3M HCl. The
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56 198 capsule was again sealed and placed on a hot plate at ~80 °C overnight prior to chemical
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59 199 separation.

Detailed procedures of chemical separation were described by Yang et al. (2010). After centrifuging, the 3M HCl sample solution was loaded onto pre-conditioned 2mL Ln Spec resin column (modified after Münker et al., 2001). Firstly, the Light and Middle Rare Earth Elements (LMREE) were eluted with 3M and 4M HCl sequentially, which was collected in a 30mL PFA beaker for further purification. Lu and Yb were effectively removed by rinsing the column with 6M HCl before collecting the Hf (+Zr) fraction, which can minimize the isobaric interference of ^{176}Lu and ^{176}Yb on ^{176}Hf . Titanium was removed with a 4M HCl + 0.5% H_2O_2 mixture. Finally, Hf (+Zr) was extracted from the column with 5mL 2M HF, collected in a 10mL PFA beaker, and gently evaporated to dryness. Then, this fraction was taken up in trace 2M HF, diluted to 1mL with 2% HNO_3 , and ready for analysis.

The LMREE fraction was gently evaporated to dryness and diluted with 1.5mL 2.5M HCl prior to the second-column purification. The solution was centrifuged and then loaded into a quartz column packed with AG50W-X12, which has been preconditioned with 2mL of 2.5M HCl. The resin was then washed with a further 2mL of 2.5M HCl, followed by 11mL of 5M HCl to remove the unwanted elements. Then, the LMREE fraction was eluted with 6mL of 6M HCl. The LMREE fraction was gently evaporated to dryness, taken up with 0.5mL of 0.25M HCl and loaded onto Ln Spec resin column (modified after Pin and Santos-Zalduegui, 1997). La, Ce and Pr were removed with 6mL of 0.25M HCl wash. Then the fraction containing Nd was eluted with 6mL of 0.25M HCl for further analyses.

Measured $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{177}\text{Hf}/^{176}\text{Hf}$ ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd}=0.7219$ and $^{179}\text{Hf}/^{177}\text{Hf}=0.7325$, respectively, using the exponential law (Yang et al., 2010). Two

reference standards were analyzed along with the studied samples and gave $^{177}\text{Hf}/^{176}\text{Hf} = 0.281954 \pm 5$ (2σ); $^{143}\text{Nd}/^{144}\text{Nd} = 0.511384 \pm 6$ (2σ) for GSP-2, and $^{177}\text{Hf}/^{176}\text{Hf} = 0.282999 \pm 5$ (2σ); $^{143}\text{Nd}/^{144}\text{Nd} = 0.512797 \pm 5$ (2σ) for RGM-1, respectively, which are comparable with the recommended values (Weis et al., 2006) of $^{143}\text{Nd}/^{144}\text{Nd}$ (0.511374 ± 3 (2σ) for GSP-2; 0.512802 ± 15 (2σ) for RGM-1) and $^{177}\text{Hf}/^{176}\text{Hf}$ (0.281949 ± 5 (2σ) for GSP-2; $^{177}\text{Hf}/^{176}\text{Hf} = 0.283017 \pm 13$ (2σ) for RGM-1, Weis et al., 2007).

5. Geochemical Results

Whole-rock major and trace element contents of samples from the Hanasi and Altay batholiths and Kurmutu and Kezgar plutons are presented in Table 1 and plot in Figure S1, together with data of Kuwei and Hemu batholiths and Jiadengyu rhyolites from Cai et al. (2011b) and Wang et al. (2011), respectively.

5.1. Major and trace element geochemistry

5.1.1. Samples from the Central Altai domain

The Hanasi granites possess relatively low SiO_2 and high CaO , Fe_2O_3 and MgO contents. Al_2O_3 , TiO_2 , MgO , Fe_2O_3 , and P_2O_5 decrease with increasing SiO_2 whereas other major oxides such as CaO , Na_2O and K_2O show poor correlation with SiO_2 (Table 1; Fig. S1). All samples are metaluminous ($A/\text{CNK} < 1$) (Fig. S2), with a narrow range of ΣREE (155ppm – 167ppm), and are enriched in LREE ($[\text{La}/\text{Yb}]_N = 5.1 - 5.3$, $[\text{Dy}/\text{Yb}]_N = \sim 1.3$, Fig. S3a). On a primitive mantle-normalized trace element diagram, all samples

show enrichment in LILE, pronounced depletion in Nb, Ta and Ti, and significant positive Pb anomalies (Fig. S3b).

5.1.2. Samples from the Qiongkuer domain

Two samples from the Kurmutu pluton are weakly peraluminous ($1.04 < \text{ACNK} < 1.07$) and display a narrow range in $\sum\text{REE}$ (100ppm – 109ppm) and are enriched in LREE ($[\text{La}/\text{Yb}]_{\text{N}} = 3.7 - 3.8$, $[\text{Dy}/\text{Yb}]_{\text{N}} = 1.1 - 1.2$). Sample ALT67 has low SiO_2 , MgO and Fe_2O_3 and high CaO , Al_2O_3 contents (Table 1; Fig. S1), and shows a weak negative Eu anomaly (0.93). This sample has much higher Sr (210 ppm) compared with the other sample, ALT65, from the same batholith ($\text{Eu}/\text{Eu}^* = 0.66$; $\text{Sr} = 151\text{ppm}$) (Fig. S3a, b), indicating accumulation of plagioclase. The two samples both show enrichment in LILE, pronounced depletion in Nb, Ta and Ti and significant positive Pb anomalies (Fig. S3b).

The Altay and Kezgar granites all have high SiO_2 , Na_2O , K_2O and low Al_2O_3 , CaO , MgO , Fe_2O_3 and P_2O_5 contents (Table 1; Fig. S1). These samples are metaluminous to weakly peraluminous ($0.98 < \text{ACNK} < 1.07$) (Fig. S2) and have various $\sum\text{REE}$ contents (97.5 – 236 ppm for Altay batholith; 47.4 – 112 ppm for Kezgar pluton). They are enriched in LREE ($[\text{La}/\text{Yb}]_{\text{N}} = 1.5 - 2.8$; $[\text{Dy}/\text{Yb}]_{\text{N}} = 0.90 - 0.96$ for Altay batholith; $[\text{La}/\text{Yb}]_{\text{N}} = 0.94 - 2.6$; $[\text{Dy}/\text{Yb}]_{\text{N}} = 0.90 - 1.1$ for Kezgar pluton), with strongly negative Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.29 - 0.65$ for Altay batholith; $0.21 - 0.42$ for Kezgar pluton; Fig. S3c), and are enriched in LILE and depleted in Nb, Ta and Ti (Fig. S3d).

Table 1

5.2. Nd-Hf isotope compositions

Nd-Hf isotopic results for whole-rock samples are presented in Tables 2 and 3. All samples have low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (0.512340 – 0.512559), corresponding to initial ϵ_{Nd} values ranging from -3.0 to +1.7 (Fig. 4), whereas the $^{176}\text{Hf}/^{177}\text{Hf}$ ratios range from 0.282680 to 0.282991 corresponding to positive but highly variable initial ϵ_{Hf} values (+2.1 – +12.9) (Fig. 4). All samples plot above the terrestrial line defined by Vervoort et al. (2011), indicating significant Nd-Hf isotopic decoupling (Fig. 4). One sample from the Hemu batholith (HM-12) has higher $\epsilon_{\text{Nd}}(t)$ and $\epsilon_{\text{Hf}}(t)$ values ($\epsilon_{\text{Nd}}(t) = -2.1$ and $\epsilon_{\text{Hf}}(t) = +9.2$) than the other three samples from the same batholith ($\epsilon_{\text{Nd}}(t) = -3.0$ – -2.9 and $\epsilon_{\text{Hf}}(t) = +2.1$ – $+3.1$) and shows lower $^{147}\text{Sm}/^{143}\text{Nd}$ and higher $^{176}\text{Lu}/^{177}\text{Hf}$ ratios (Table 2) and higher contents of Nd and Hf (Table 1), suggesting that the Sm-Nd and Lu-Hf isotopic systems might be disturbed. One sample from the Kuwei batholith (KW-15) has higher $\epsilon_{\text{Nd}}(t)$ values but similar $\epsilon_{\text{Hf}}(t)$ values ($\epsilon_{\text{Nd}}(t) = 1.7$ and $\epsilon_{\text{Hf}}(t) = +3.3$) compared with the other samples from the same batholith ($\epsilon_{\text{Nd}}(t) = -1.6$ – -1.8) and $\epsilon_{\text{Hf}}(t) = +3.7$ – $+4.2$). The sample also has a lower $^{147}\text{Sm}/^{143}\text{Nd}$ ratio than the other samples from the same batholith (Fig. 4), suggesting that the Sm-Nd isotopic system of this sample (KW-15) was possibly disturbed.

Table 2 and 3

Figs. 4 about here

6. Discussion

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288 **6.1. Effect of post-crystallization alteration/metasomatism on Nd-Hf isotopes**

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290 Both Sm-Nd and the Lu-Hf isotopic systems are generally resistant to secondary
291 alteration and metasomatism (Blichert-Toft and Arndt, 1999; Choi and Mukasa, 2012).
292 Although samples from the Hanasi, Hemu and Kuwei batholiths show slight alteration
293 with sericite on plagioclase and chlorite (Fig. 4), samples from the other locations are still
294 fresh. There are no large variations in the LILE (Rb, Sr, Ba) in each group of samples
295 (Fig. S3), suggesting weak effect of alteration and/or metasomatism. Most samples have
296 uniform Nd-Hf isotopic compositions, demonstrating that the Sm-Nd and Lu-Hf isotopic
297 systems of these samples were not significantly disturbed by post-crystallization
298 alteration and/or metasomatism.

299 **6.2. I-type granites: melting of an igneous lower crust**

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301 I-type granites are generally metaluminous or weakly peraluminous, especially at the
302 more mafic end of the compositional spectrum (Chappell, 1999; Chappell et al., 2012).
303 Our samples from the Hanasi batholith and Kurmutu pluton have low SiO₂ contents (62 –
304 67 wt %. and 65 – 67 wt %., respectively), and compositionally vary from metaluminous
305 to weakly peraluminous (A/CNK < 1.1; Fig. S2). Additionally, these samples show
306 negative correlation between P₂O₅ and SiO₂ (Fig. S1d), demonstrating fractional
307 crystallization of apatite because P has a low solubility in metaluminous melts (Montel

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4 308 and Vielzeuf, 1997; Wolf and London., 1994). Therefore, the primary parental magma of
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6 309 these samples was likely metaluminous and samples from these two intrusions belong to
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12 311 Samples from the Kezgar pluton and Altay batholith show high contents of K₂O and
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15 312 Na₂O (Fig. S1c, d), similar to the A-type granite (e.g. Whalen et al., 1987). However,
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17 313 these samples contain abundant hydrous minerals, such as muscovite and biotite,
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19 314 indicating that their parental magmas were possibly enriched in water, which is in
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22 315 contrast to the typical A-type granite containing high temperature anhydrous pyroxene or
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24 316 fayalite (King et al., 1997). The relatively low Ga/Al and Zr, Nb, Ce and Y contents
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27 317 (Table 1), combined with petrographic characteristics, strongly suggest that these
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29 318 intrusions cannot be A-type granite that is characterized by high Ga/Al ratios and
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32 319 enrichment in high field strength elements (HFSE) and REE (Collins et al., 1982; Whalen,
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34 320 1985; Whalen et al., 1987). Samples from these intrusions have very high SiO₂ (>75
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36 321 wt %), high (Na₂O+K₂O)/CaO and low P₂O₅ (Fig. S1d) and FeO/MgO ratios (Table 1).
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39 322 All these features indicate that their parental magmas have undergone large degree of
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42 323 fractionation of biotite, plagioclase and apatite. Considering that these samples are
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44 324 weakly peraluminous, without cordierite, and that muscovite is subhedral and occurs
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47 325 around biotite, thus possibly a late stage product of alteration, we categorize the Kezgar
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49 326 pluton and Altay batholith as highly fractionated I-type granites.

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52 327 I-type granite is generally derived from dehydration melting of igneous rock or their
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54 328 metamorphic equivalents (Chappell, 1999; Rutter and Wyllie, 1988; Wolf and Wyllie,
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57 329 1994). The metaluminous or weakly peraluminous feature of the studied I-type granites
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60 330 (A/CNK<1.1) also precludes their derivation from melting of a sedimentary source or
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melting of an igneous source under water-saturated condition because both processes would produce strongly peraluminous melts (Beard and Lofgren, 1991; Douce, 1996; Douce and Johnston, 1991; Lebreton and Thompson, 1988). The low SiO₂ contents for the Hanasi and Kurmutu granites suggest an intermediate to mafic source such as amphibolite. Similarly, the primary magma of the Kezgar and Altay granites was probably granodioritic if early fractionation crystallization is considered. Hydrous minerals, such as amphibole, would break down firstly to supply free water for initial dehydration melting with plagioclase, orthopyroxene or garnet as major residual phases (Wolf and Wyllie, 1994). Flatten HREE patterns and high HREE concentrations of the studied I-type granites (Fig. S3a, c) illustrate that garnet was not in the residue because of its extremely high partition coefficients (K_d) for the HREE (Bea et al., 1994), which can result in strong depletion of HREE in resultant melts. Therefore, the studied I-type granites may be derived from dehydration melting of an intermediate-mafic lower crust in the Chinese Altai above the garnet stability field (Fig. 5).

Fig. 5

6.3. Nd-Hf decoupling of crust in the Chinese Altai: from an ancient basement?

Significant Nd-Hf decoupling was shown by all samples of this study (Fig. 4), which may be resulted from disequilibrium melting processes or inherited from the magma source. Melt may show higher ¹⁷⁶Hf/¹⁷⁷Hf ratios than the source if dissolution of zircon was incomplete during anatexis, because the very low Lu/Hf ratio in zircon could result in low unradiogenic ¹⁷⁶Hf/¹⁷⁷Hf ratios in zircon (e.g. Tang et al., 2014). This process will cause correlation between Hf isotopic compositions and Hf contents of the melt (Farina et

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4 354 [al., 2014](#)). However, the poor correlation between Hf and $\varepsilon_{\text{Hf}}(t)$ values shown by our
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6 355 samples does not support disequilibrium melting of the source. Thus, the decoupled Nd-
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9 356 Hf isotopic features of the studied samples should be inherited from the magma source.

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12 357 Above discussion suggested that I-type granites in this study were produced by
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14 358 dehydration melting of an intermediate-mafic lower crust. The peraluminous granitic
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17 359 rocks (granites from Kuwei and Hemu batholiths and Jiadengyu strongly peraluminous
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19 360 rhyolites) also show significantly Nd-Hf isotopic decoupling. Cai et al. ([2011b](#)) suggested
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21 361 that strongly peraluminous felsic rocks in the study area were derived from partial
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23 362 melting of metasediments of the Chinese Altai (e.g. Habahe Group). The age of our
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25 363 samples ranges from 445 to 368 Ma, and the sample locations cover almost the entire
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27 364 Chinese Altai, indicating that decoupling of the Nd-Hf isotopic systems is a common
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29 365 feature in the deep crust of the Chinese Altai during early to middle Paleozoic. Previous
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31 366 studies have suggested that crustal magma sources may display Nd-Hf isotopic
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33 367 decoupling through elevated Lu/Hf ratios (but depressed Sm/Nd) which can result in high
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35 368 radiogenic $^{176}\text{Hf}/^{177}\text{Hf}$ with time evolution ([Chauvel et al., 2009](#); [Schmitz et al., 2004](#);
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37 369 [Vervoort et al., 2011](#)). The studied I-type granites and strongly peraluminous granites and
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39 370 rhyolites were derived from different sources, including sedimentary and igneous sources
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41 371 as mentioned above. The mechanism to elevate the Lu/Hf ratios may be different in
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43 372 different sources as discussed below.

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45 373 The strongly peraluminous granites and rhyolites were possibly derived from an
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47 374 aluminum-enriched sedimentary source. Previous researchers suggested that such magma
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49 375 source may have elevated Lu/Hf ratios through removing of zircon minerals during
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51 376 sedimentary processes, because zircon is a main Zr-Hf bearing mineral and has a very
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low Lu/Hf ratio (Carpentier et al., 2009; Chauvel et al., 2009; Garcon et al., 2013; Vervoort et al., 1999, 2000, 2011). With time evolving, such magma sources would show high radiogenic $^{176}\text{Hf}/^{177}\text{Hf}$. However, the studied strongly peraluminous granites and rhyolites show little depletion in Zr and Hf compared with neighboring elements (Fig. S3b) and have Lu/Hf ratios (0.04 – 0.14) similar to the average crust (Lu/Hf = 0.11, Rudnick and Fountain, 1995), suggesting that zircon might not be significantly removed from the magma source. More importantly, the Habahe Group rocks, a possible source for the studied strongly peraluminous granites and rhyolites, also have low Lu/Hf ratios, ranging from 0.06 to 0.098, similar to average crustal composition (Long et al., 2012; Rudnick and Fountain, 1995). All these indicate that “zircon effect” cannot account for high radiogenic $^{176}\text{Hf}/^{177}\text{Hf}$ ratios for the crustal magma source of the strongly peraluminous granites and rhyolites.

On the other hand, the studied I-type granites were derived from an intermediate to mafic igneous source which may attain high Lu/Hf ratios with enrichment of garnet (Schmitz et al., 2004). Since garnet has very high Lu/Hf ratios, enrichment of garnet may provide very high radiogenic $^{176}\text{Hf}/^{177}\text{Hf}$ as time evolves for such igneous source (Schmitz et al., 2004). Generally, garnet does not break down but behaves as a residual phase during dehydration melting of an intermediate to mafic igneous source (Wolf and Wyllie, 1994). Thus, melts derived from such source would show strong depletion in HREE (Schmitz et al., 2004). This is inconsistent with high content of HREE and flatten HREE pattern shown in the studied I-type granites (Fig. S3a, c) and other I-type granites in the Chinese Altai (Yuan et al., 2007).

Despite of the granitic rocks in the Chinese Altai show juvenile Nd isotopic signatures, an ancient crustal basement was invoked to explain why some granites have lower $\epsilon_{\text{Nd}}(t)$ values (Wang et al., 2006; 2009). However, as discussed above, ancient crustal basement with high Lu/Hf ratios does not exist in the Chinese Altai, thus, an alternative model is necessary to explain the prevailed Nd-Hf isotopic decoupling in the crust of the Chinese Altai.

6.4. Chinese Altai: a juvenile magmatic arc in the Paleozoic

In subduction zones, basaltic magma can be extracted from mantle wedge, forming initial arc crust and then the crust will evolve towards more felsic one as evidenced by Mariana arc. Subduction process can efficiently modify isotopic systems of the mantle through metasomatism by a variety of agents, including fluid and melt (Brenan et al., 1995; Elliott et al., 1997; Hanyu et al., 2006; Todd et al., 2010). These metasomatism agents can effectively fractionate elements from each other and may result in isotopic decoupling in the mantle (Hanyu et al., 2006; Pearce et al., 1999; Spandler et al., 2007; Todd et al., 2010).

Studies on detrital zircons suggest that the metasedimentary rocks in the Chinese Altai predominantly came from a local igneous provenance (Long et al., 2007; 2008; 2010; Sun et al., 2008). I-type granites and strongly peraluminous granites and rhyolites all have similar Nd-Hf isotopic signatures, indicating that their crustal magma sources might be extracted from the lithospheric mantle underneath and inherited Nd-Hf isotopic decoupling feature of the mantle. The impact of subducted sediments on Nd-Hf isotopic systems of the mantle wedge could be mainly related to hydrous melting because Nd and

Hf are almost immobile in fluid (Brenan et al., 1995; Hanyu et al., 2006; Pearce et al., 1999; Spandler et al., 2007; Todd et al., 2010). The behavior of Nd was controlled by monazite/ allanite while Hf is mainly buffered by zircon (Herman and Rubatto, 2009). Since zircon can stay in the residue under high temperature (~ 1000 °C), breakdown of monazite can efficiently fractionate Nd from Hf and result in high Nd/Hf ratios in the melt. Consequently, slab-derived melts possessing high Nd/Hf ratios (Spandler et al., 2007; Todd et al., 2010) may add to the mantle wedge and therefore result in Nd-Hf isotopic decoupling in the mantle (Hanyu et al., 2006; Todd et al., 2010), as shown by the modern Sunda and Lesser Antilles arcs (Fig. 4). To better illustrate the process, we built a model showing melting of subducted sediments and the impact of melt on the Nd-Hf isotopic systems of the Chinese Altai mantle that underwent prolonged subduction from 460 Ma or earlier to 360 Ma in the (Sun et al., 2008; Cai et al., 2011b, 2011c). The mantle end-member is represented by depleted mantle of Salters and Stracke (2004) since the mantle may suffer from extraction of arc magma (Cai et al., 2010; 2012). The average composition of the Habahe Group is employed as subducted sediments because of its similar composition to GLOSS (Long et al., 2012; Plank and Langmuir, 1998) with residual trace phase proportion of monazite: rutile: zircon of 0.0005-0.0015:0.01:0.0005 during melting. Since no whole-rock Hf isotopic composition is available for the Habahe Group, its composition is represented by a vertical bar at $e_{Nd}(t) = -5.3$ (Long et al., 2012). The model shows that the lithospheric mantle metasomatized by subducted sediments-derived melt display significant decoupling between Nd and Hf isotopes due to selective enrichment of Nd over Hf. Such Nd-Hf isotopic decoupling in the lithosphere mantle is clearly evidenced by early Devonian mafic dikes (Fig. 4). Our granitic samples plot along

the mixing line, implying that their crustal magma sources could be derived from underneath lithosphere mantle that was previously metasomatized by melt from subducted components.

The Chinese Altai is part of the Altai-Mongolia terrane which is located in the southwest of the Lake zone and Tuva-Mongolia micro-continent. The metasedimentary rocks in the Chinese Altai also contain minor old continental material as represented by subordinate Precambrian detrital zircons (Jiang et al., 2011; Long et al., 2007, 2008, 2010; Sun et al., 2008). If the Chinese Altai magmatic arc was directly built on a Precambrian basement, such as the Tuva-Mongolia, then the addition of old continental material into the magma source would decrease the $\Delta\epsilon_{\text{Hf}}(t)$ values ($\Delta\epsilon_{\text{Hf}}(t) = \epsilon_{\text{Hf}}(t) - 1.55 * \epsilon_{\text{Nd}}(t) - 1.21$, Vervoort et al., 2011), as well as both $\epsilon_{\text{Nd}}(t)$ and $\epsilon_{\text{Hf}}(t)$ values. This is because old continental crust generally has coupled Nd-Hf isotopic features (Chauvel et al., 2014; Vervoort et al., 2011) and low $\epsilon_{\text{Nd}}(t)$ and $\epsilon_{\text{Hf}}(t)$ values. The poor correlation between $\Delta\epsilon_{\text{Hf}}(t)$ and $\epsilon_{\text{Nd}}(t)$, together with high $\epsilon_{\text{Hf}}(t)$ values, suggesting that the Chinese Altai was not built on an ancient basement. The positive $\epsilon_{\text{Hf}}(t)$ values of I-type granite and strongly peraluminous granites and rhyolites demonstrate juvenile nature of crust in the Chinese Altai in the Paleozoic. We consider that the Chinese Altai represents early Paleozoic magmatic arc mainly made of juvenile materials nearby western Mongolia.

7. Conclusions

The I-type granites were derived from dehydration melting of intermediate to mafic igneous sources. Both I-type and strongly peraluminous granites and rhyolites show significant Nd-Hf isotopic decoupling with relatively high $\epsilon_{\text{Hf}}(t)$ and low $\epsilon_{\text{Nd}}(t)$ values,

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4 468 indicating widespread Nd-Hf isotopic decoupling in deep crust. This decoupling cannot
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6 469 be resulted from melting of an ancient basement with elevated Lu/Hf ratios but can be
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9 470 ultimately resulted from the selective enrichment of Nd over Hf in the mantle due to
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11 471 metasomatism by melts derived from subducted sediments. Our data support that the
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14 472 Chinese Altai represents a Paleozoic magmatic arc made up of relatively juvenile
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16 473 materials and developed nearby western Mongolia.

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807

808 **Figure Captions**

809 Fig. 1. (a) Simplified regional map showing known extent of CAO B (b) Sketch map
810 highlighting the tectonic framework of the Chinese Altai and neighbouring Hovd
811 and Lake zone (revised from [Long et al. \(2010\)](#), [Sun et al. \(2008\)](#) and [Xiao et al.](#)
812 [\(2015\)](#)). The Chinese Altai is bounded by blue dash line in the north and Erqis
813 Fault zone in the south. CAO B extends from Siberia in the north to the North
814 China and Tarim cratons in the south.

815 Fig. 2. Simplified geological map of the (a) Hanasi batholith (HB), Hemu batholith
816 (HMB) and Jiadengyu rhyolite (modified after Baihaba and Qiongkuer Geological
817 Map at 1:200,000 scale and [Wang et al., 2011](#)), (b) Kuwei batholith (KB, modified
818 after Fuyun Geological Map at 1:200,000 scale), showing rock types and sample

locations, (c) Kurmutu pluton (KMP), Altai batholith (AB) and Kezgar pluton (KEP; modified after Altai and Alagarke Geological Map at 1:200,000 scale).

Fig. 3. Petrographic characteristics of (a) Hanasi porphyritic biotite granite; (b) Hemu two-mica granite; (c) Kuwei biotite granite; (d) Kurmutu biotite granite; (e) Altai medium-grained biotite granite; (f) Kezgar leuco-granite. Pl = plagioclase; Bt = biotite; Qz = quartz; Ttn = titanite; Kfs = K-feldspar; Ms = Muscovite.

Fig. 4. The diagram of initial $\epsilon_{Nd}(t)$ vs. Initial $\epsilon_{Hf}(t)$ showing Nd-Hf decoupling of studied samples and metasomatism of lithosphere mantle by melt from subducted sediments. Data of depleted mantle is from Salters and Stracke (2004). The subducted sediment is represented by average composition of the Habahe Group (HG) which 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 orange vertical bar in the diagram at $\epsilon_{Nd}(t) = -5.3$. The subducted sediments were modeled to be melted at eclogite facies (3Gpa) with 5% partial melting degree and residual assemblage of clinopyroxene: garnet: mica (0.4: 0.3: 0.3). The partition coefficients are from Johnson (1998) and Adam and Green (2006). Trace phases of residue is composed of monazite, rutile and zircon with proportion of 0.0005–0.0015:0.001:0.0005. The partition coefficients are from Skora and Blundy (2010), Foley et al. (2000) and Rubatto and Herman (2007). Blue lines refer to depleted mantle metasomatized by melts from subducted sediments with residual 0.0005 monazite and red lines represent mixing between depleted mantle and melts with residual 0.0015 monazite. The data of mafic dikes is from Cai et al. (2010). Terrestrial Array is from Vervoort et al. (2011), $\epsilon_{Hf} = 1.55 * \epsilon_{Nd} + 1.21$. Data of arc

related rocks were compiled using the Georoc database (<http://georoc.mpch-mainz.gwdg.de/georoc/>) for IAV.

Fig. 5. Binary plot $\text{Al}_2\text{O}_3 + \text{FeOt} + \text{MgO} + \text{TiO}_2$ vs. $\text{Al}_2\text{O}_3/(\text{FeOt} + \text{MgO} + \text{TiO}_2)$ for the studied samples; outlined are domains occupied by experimental granitic melts obtained by partial melting of metapelites, metagreywackes and amphibolites (Janoušek et al., 2010 and references therein).

Figure.1

Figure.2

Figure.3

Figure.4

Figure.5

Table.1

Table 1.

Major and trace elements of the Hanasi and Altai batholiths and Kurmutu and Kezgar plutons from the Chinese Altai, NW China

Sample	KN01	KN02	KN04	KN05	KN06	KN07	KN09	KN10	ALT66	ALT67
Pluton Name	HB	HB	HB	HB	HB	HB	HB	HB	KUP	KUP
SiO ₂	67.08	65.28	65.38	63.20	64.07	62.90	62.31	62.58	68.13	65.19
TiO ₂	0.90	1.01	0.97	1.08	1.10	1.11	1.28	1.13	0.52	0.51
Al ₂ O ₃	14.53	14.86	14.77	15.03	15.01	14.92	15.44	15.19	14.97	17.71
Fe ₂ O ₃	5.12	5.70	5.56	6.21	6.17	6.71	6.88	6.40	4.47	3.78
MnO	0.08	0.09	0.09	0.10	0.10	0.10	0.12	0.11	0.09	0.07
MgO	1.36	1.50	1.52	1.69	1.74	2.04	1.90	1.86	1.69	1.07
CaO	3.43	3.86	3.89	4.24	4.60	3.67	4.28	4.68	3.83	5.44
Na ₂ O	3.86	3.42	3.42	3.22	3.28	3.53	3.55	3.50	3.17	3.55
K ₂ O	2.76	2.95	2.68	2.94	2.43	2.89	2.60	2.25	1.71	1.20
P ₂ O ₅	0.20	0.23	0.21	0.26	0.25	0.26	0.30	0.26	0.11	0.10
LOI	0.96	1.20	1.21	1.27	1.30	1.20	1.40	1.44	0.70	0.60
Total	100.29	100.10	99.71	99.24	100.05	99.34	100.05	99.41	99.38	99.22
A/CNK	0.93	0.94	0.95	0.93	0.91	0.96	0.94	0.91	1.07	1.04
Mg [#]	0.23	0.23	0.23	0.23	0.24	0.25	0.23	0.24	0.30	0.24
FeO/MgO	2.87	2.90	2.80	2.81	2.70	2.52	2.77	2.63	2.03	2.70
(Na ₂ O+K ₂ O)/CaO	1.93	1.65	1.57	1.45	1.24	1.75	1.44	1.23	1.27	0.87
Sc	17.5	19.0	18.8	20.4	20.3	21.1	21.8	20.5	12.6	17.3
Ti	5343	5986	5845	6540	6534	6757	7530	6868	3225	3243
V	93	109	113	135	134	148	151	142	82.7	49.2
Cr	7.10	8.48	8.99	10.1	10.3	10.6	10.9	14.5	23.7	11.3
Mn	612	712	713	759	761	800	916	835	638	538
Co	10.6	12.5	12.5	14.3	13.9	17.2	16.3	15.3	11.1	7.21
Ni	10.3	11.7	12.5	15.2	14.5	17.3	16.2	16.5	12.9	7.02
Cu	26.8	43.2	39.2	40.9	36.2	26.9	47.7	41.5	4.54	4.71
Zn	55.8	69.6	63.4	71.0	72.8	82.1	75.8	73.4	74.1	54.7
Ga	16.8	17.6	17.2	17.6	17.8	17.3	18.2	17.9	16.2	20.1
Ge	2.08	2.20	2.11	2.20	2.16	2.22	2.33	2.26	1.77	1.61
Rb	94	110	98	138	74	129	114	81	84	70
Sr	138	141	139	151	151	136	180	152	151	210
Y	38.3	38.7	38.3	37.0	37.2	36.4	36.9	36.7	28.0	31.2
Zr	226	217	219	202	205	196	195	197	156	315
Nb	9.87	10.0	9.94	9.51	9.73	9.23	10.53	9.47	6.16	6.98
Ba	398	419	414	382	418	399	382	375	289	248
Cs	3.24	3.65	3.50	6.23	1.95	6.15	5.11	3.09	10.6	6.95
La	28.2	28.9	28.8	27.6	27.1	26.1	27.0	27.0	15.8	16.4
Ce	59.4	61.9	61.5	58.7	58.5	57.0	58.7	57.9	35.8	38.7

Pr	7.69	8.00	8.01	7.60	7.60	7.40	7.60	7.55	4.78	4.85
Nd	31.6	32.9	32.9	31.9	31.6	30.9	32.0	31.5	19.9	20.9
Sm	6.97	7.23	7.18	7.10	6.98	6.86	7.11	6.97	4.42	5.16
Eu	1.55	1.63	1.63	1.62	1.65	1.60	1.62	1.65	0.95	1.60
Gd	7.06	7.26	7.25	7.12	7.02	6.96	7.12	7.02	4.46	5.33
Tb	1.15	1.17	1.18	1.14	1.14	1.13	1.15	1.13	0.77	0.91
Dy	7.22	7.30	7.32	7.14	7.14	7.00	7.08	7.02	4.93	5.85
Ho	1.56	1.55	1.53	1.49	1.50	1.48	1.50	1.48	1.08	1.26
Er	4.23	4.21	4.22	4.07	4.05	4.02	4.02	4.01	3.02	3.45
Tm	0.63	0.62	0.62	0.60	0.59	0.59	0.59	0.58	0.46	0.50
Yb	3.94	3.90	3.92	3.77	3.74	3.65	3.66	3.71	3.01	3.21
Lu	0.60	0.60	0.60	0.58	0.56	0.56	0.57	0.56	0.46	0.49
Hf	6.32	6.01	6.09	5.64	5.64	5.50	5.43	5.50	4.55	7.97
Ta	0.93	0.91	0.90	0.86	0.87	0.85	0.92	0.86	0.58	0.56
Pb	20.0	18.9	27.4	26.1	20.8	24.5	42.0	20.7	13.6	9.5
Th	10.6	10.1	10.4	9.5	9.5	9.3	9.0	9.2	9.0	6.5
U	3.32	3.16	3.63	3.55	2.86	2.86	2.82	2.85	2.27	1.85
ΣREE	162	167	167	160	159	155	160	158	100	109
10000*Ga/Al	2.19	2.24	2.20	2.21	2.24	2.19	2.22	2.23	2.04	2.14
(La/Yb) _N	5.13	5.33	5.27	5.25	5.21	5.12	5.30	5.22	3.76	3.66
(Dy/Yb) _N	1.23	1.25	1.25	1.27	1.28	1.28	1.30	1.27	1.10	1.22
Eu/Eu [*]	0.68	0.69	0.69	0.70	0.72	0.71	0.70	0.72	0.66	0.93

Sample	ALT03	ALT04	ALT05	ALT06	ALT15	ALT16	ALT17	ALT18	ALT19
Pluton	AB	AB	AB	AB	KEP	KEP	KEP	KEP	KEP
SiO ₂	75.15	76.58	74.69	76.56	76.98	77.68	79.16	77.23	77.94
TiO ₂	0.13	0.09	0.16	0.15	0.06	0.07	0.08	0.07	0.07
Al ₂ O ₃	13.35	12.93	13.58	12.44	12.43	12.04	11.28	12.14	11.95
Fe ₂ O ₃	1.37	0.99	1.67	1.62	0.50	0.53	0.51	0.62	0.55
MnO	0.03	0.02	0.03	0.03	0.01	0.03	0.02	0.03	0.02
MgO	0.26	0.20	0.30	0.30	0.05	0.08	0.10	0.11	0.10
CaO	1.32	1.42	1.48	1.40	0.50	0.50	0.78	0.57	0.54
Na ₂ O	3.69	4.04	4.34	3.76	3.66	3.21	3.22	3.87	3.43
K ₂ O	3.72	3.09	2.83	2.77	4.74	4.80	3.51	4.59	4.61
P ₂ O ₅	0.03	0.02	0.03	0.03	0.01	0.01	0.01	0.01	0.01
LOI	0.27	0.27	0.35	0.35	0.47	0.33	0.47	0.53	0.50
Total	99.31	99.64	99.47	99.41	99.41	99.29	99.14	99.78	99.73
A/CNK	1.07	1.03	1.05	1.06	1.03	1.06	1.07	0.98	1.03
Mg [#]	0.18	0.18	0.17	0.17	0.11	0.14	0.19	0.16	0.17
FeO/MgO	3.98	3.84	4.23	4.08	7.22	5.21	3.74	4.46	4.01
(Na ₂ O+K ₂ O)/CaO	5.62	5.03	4.84	4.65	16.98	15.89	8.66	14.80	14.83
Sc	4.45	4.29	6.30	5.79	7.51	8.44	7.10	8.86	9.27
Ti	753	617	1015	929	398	419	487	442	461

V	9.55	8.04	11.8	11.5	2.89	4.00	3.73	4.75	4.44
Cr	3.15	2.61	3.24	2.81	2.12	2.62	2.33	2.16	2.44
Mn	201	165	261	235	100	179	135	186	173
Co	1.87	1.28	1.65	1.65	0.15	0.25	0.51	0.37	0.33
Ni	1.23	1.19	1.15	1.41	0.65	0.51	0.64	0.72	0.65
Cu	5.50	3.59	2.63	3.30	9.85	6.43	11.5	17.5	6.63
Zn	26.8	22.2	33.8	31.9	16.2	18.8	17.0	25.4	19.8
Ga	13.9	13.7	15.2	13.6	14.3	13.8	12.0	14.1	13.8
Ge	1.47	1.18	1.70	1.54	1.33	1.45	1.08	1.40	1.47
Rb	96	81	86	79	150	170	123	160	164
Sr	85	96	85	84	26.1	24.8	41.2	26.7	24.9
Y	62.6	47.7	130	69.1	42.8	39.9	32.2	43.2	43.7
Zr	99.3	73.0	119	93.9	65.6	59.6	82.0	80.4	61.7
Nb	4.23	3.93	6.00	5.58	15.4	13.8	10.8	21.7	15.6
Ba	366	339	276	291	63.1	107	228	95.1	131
Cs	1.80	1.60	2.22	2.03	2.52	3.26	2.18	3.11	3.20
La	29.5	13.8	31.7	29.4	9.6	13.5	5.1	13.3	17.4
Ce	62.6	30.5	70.0	65.0	22.6	30.7	12.4	30.0	37.5
Pr	7.95	3.93	9.16	8.45	3.00	3.84	1.70	4.05	4.78
Nd	31.8	16.0	37.0	34.4	12.1	15.2	7.1	16.1	19.0
Sm	7.34	4.22	10.19	8.48	3.48	3.97	2.09	4.25	4.89
Eu	1.03	0.98	1.11	1.04	0.27	0.35	0.33	0.34	0.41
Gd	7.78	5.02	13.23	8.97	4.34	4.69	2.76	4.99	5.57
Tb	1.47	1.04	2.88	1.69	0.97	0.97	0.66	1.04	1.10
Dy	10.23	7.61	20.9	11.6	7.21	6.95	5.17	7.62	7.86
Ho	2.38	1.76	4.90	2.62	1.71	1.60	1.28	1.75	1.80
Er	7.10	5.27	14.51	7.69	4.96	4.59	3.85	5.06	5.07
Tm	1.13	0.84	2.31	1.23	0.74	0.69	0.60	0.76	0.76
Yb	7.46	5.69	15.37	8.05	4.66	4.41	3.86	4.79	4.72
Lu	1.15	0.87	2.36	1.23	0.68	0.66	0.59	0.72	0.70
Hf	3.69	3.19	4.60	3.97	3.66	3.21	3.66	4.15	3.27
Ta	0.33	0.28	0.46	0.38	1.50	1.45	0.96	1.54	1.49
Pb	26.7	24.5	23.4	21.7	21.0	23.4	18.9	27.3	23.1
Th	18.0	9.8	24.3	20.2	13.6	11.5	9.2	13.9	13.4
U	3.26	2.30	6.50	3.15	1.32	1.29	1.91	1.56	1.38
ΣREE	179	98	236	190	76	92	47	95	112
10000*Ga/Al	1.97	2.00	2.11	2.06	2.17	2.16	2.01	2.19	2.18
(La/Yb) _N	2.83	1.75	1.48	2.62	1.49	2.20	0.94	1.99	2.64
(Dy/Yb) _N	0.92	0.90	0.91	0.96	1.04	1.06	0.90	1.07	1.11
Eu/Eu*	0.42	0.65	0.29	0.36	0.21	0.25	0.42	0.23	0.24

Assuming $\text{Fe}^{2+}/\text{Fe}^{\text{total}} = 0.85$; LOI = loss of ignition

Al/(Ca+Na+K) are molar ratios

Pluton name abbreviation: HB = Hanasi batholith; KUP = Kurmutu pluton; AB = Altay batholith; KEP = Kezgar pluton.

Table 2.**Sm-Nd isotope of the studied granites from the Chinese Altai, NW China.**

Sample	Age(Ga)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd} (\pm 2\sigma)$	$(^{143}\text{Nd}/^{144}\text{Nd})_i$	$\epsilon_{\text{Nd}}(t)$
Hanasi batholith					
KN01	0.398	0.1333	0.512445 ± 6	0.512080	-0.6
KN02	0.398	0.1328	0.512454 ± 5	0.512090	-0.4
KN04	0.398	0.1321	0.512437 ± 4	0.512076	-0.7
Kurmutu plutonh					
ALT66	0.421	0.1344	0.512518 ± 6	0.512129	0.8
ALT67	0.421	0.1491	0.512538 ± 4	0.512107	0.4
Altay batholith					
ALT03	0.368	0.1398	0.512505 ± 4	0.512152	-0.3
ALT04	0.368	0.1593	0.512533 ± 6	0.512130	-0.2
Kergar pluton					
ALT15	0.411	0.1736	0.512526 ± 11	0.512035	-1.0
ALT16	0.411	0.1581	0.512509 ± 10	0.512063	-0.6
ALT17	0.411	0.1782	0.512481 ± 12	0.511977	-2.2
Hemu batholith					
HM02	0.412	0.1424	0.512344 ± 5	0.511941	-3.0
HM03	0.412	0.1382	0.512342 ± 5	0.511951	-2.9
HM04	0.412	0.1393	0.512341 ± 5	0.511946	-2.9
HM12	0.412	0.1347	0.512373 ± 6	0.511992	-2.1
Kuwei batholith					
KW14	0.419	0.1185	0.512356 ± 4	0.512015	-1.6
KW15	0.419	0.1120	0.512509 ± 10	0.512187	1.7
KW16	0.419	0.1205	0.512350 ± 5	0.512003	-1.8
Jiadengyu rhyolite					
JDY13	0.445	0.1181	0.512403 ± 7	0.512041	-0.7
JDY14	0.445	0.1324	0.512390 ± 5	0.511985	-1.6
JDY15	0.445	0.1282	0.512405 ± 5	0.512013	-1.1

Note

^a $^{147}\text{Sm}/^{144}\text{Nd}$ are calculated using whole-rock Sm and Nd contents in Table 1.

^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, $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} = 0.512638$,

$(^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}} = 0.1967$; $\lambda = 6.54 \times 10^{-6}$. t=crystallization time of zircon.

Table. 3. Lu-Hf isotope of the studied granites from the Chinese Altai, NW China.

Sample	Age(Ga)	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}(\pm 2\sigma)$	$(^{176}\text{Hf}/^{177}\text{Hf})_i$	$\epsilon_{\text{Hf}}(t)$	$\Delta\epsilon_{\text{Hf}}(t)$
Hanasi batholith						
KN01	0.398	0.0134	0.282775 $\pm$ 5	0.282676	5.3	5.0
KN02	0.398	0.0140	0.282775 $\pm$ 5	0.282673	5.2	4.6
KN04	0.398	0.0138	0.282767 $\pm$ 5	0.282666	4.9	4.8
Kurmutu pluton						
ALT66	0.421	0.0143	0.282908 $\pm$ 9	0.282797	10.1	7.7
ALT67	0.421	0.0087	0.282944 $\pm$ 13	0.282876	12.9	11.1
Altay batholith						
ALT03	0.368	0.0441	0.282991 $\pm$ 8	0.282692	5.1	4.4
ALT04	0.368	0.0386	0.282972 $\pm$ 10	0.282710	5.8	4.9
Kezgar pluton						
ALT15	0.411	0.0263	0.282895 $\pm$ 9	0.282696	6.3	6.6
ALT16	0.411	0.0288	0.282880 $\pm$ 9	0.282662	5.0	4.7
ALT17	0.411	0.0226	0.282802 $\pm$ 5	0.282631	3.9	6.1
Hemu batholith						
HM02	0.412	0.0123	0.282698 $\pm$ 10	0.282605	3.1	6.5
HM03	0.412	0.0136	0.282694 $\pm$ 9	0.282590	2.6	5.9
HM04	0.412	0.0137	0.282680 $\pm$ 9	0.282576	2.1	5.4
HM12	0.412	0.0197	0.282927 $\pm$ 10	0.282778	9.2	11.2
Kuwei batholith						
KW14	0.419	0.0112	0.282719 $\pm$ 8	0.282632	4.2	5.5
KW15	0.419	0.0126	0.282704 $\pm$ 7	0.282607	3.3	-0.5
KW16	0.419	0.0113	0.282706 $\pm$ 7	0.282618	3.7	5.3
Jiadengyu rhyolites						
JDY13	0.445	0.0111	0.282883 $\pm$ 8	0.282791	10.4	10.3
JDY14	0.445	0.0179	0.282889 $\pm$ 10	0.282742	8.6	9.9
JDY15	0.445	0.0162	0.282877 $\pm$ 10	0.282744	8.7	9.2

Note

^{a.} ($^{176}\text{Lu}/^{177}\text{Hf}$) are calculated by content of Lu and Hf from Table 1.

^{b.} $\lambda=1.867\times 10^{-11}\text{year}^{-1}$ (Söderlund et al., 2004); ($^{176}\text{Lu}/^{177}\text{Hf}$)_{CHUR}= 0.0332 and ($^{176}\text{Hf}/^{177}\text{Hf}$)_{CHUR}= 0.282772 (Blichert-Toft and Albarède, 1997); t=crystallization time of zircon.

^{c.} $\Delta\epsilon_{\text{Hf}}(t)$ means the deviation of the $\epsilon_{\text{Hf}}(t)$ value of the sample from the Terrestrial Array (Vervoort et al., 2011).

$\Delta\epsilon_{\text{Hf}}(t) = \epsilon_{\text{Hf}} - 1.55 * \epsilon_{\text{Nd}} - 1.21$.