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Triassic depleted lithospheric mantle underneath the Paleozoic Chinese Altai orogen: evidence from MORB-like basalts

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Abstract

Whole-rock geochemical and Sr–Nd–Pb–Hf isotopic analyses were conducted on the Triassic Ashele post-orogenic basalts from the Chinese Altai. The tholeiitic Ashele basalts possess moderate Na_2O (2.46–4.52 wt%) but extremely low K_2O (≤0.06 wt%) contents, and are depleted in light rare earth elements (LREE). Their $(La/Sm)_N$ and $(La/Yb)_N$ ratios range from 0.78 to 0.88 and from 0.72 to 0.84, respectively, which,

together with their low initial $^{87}Sr/^{86}Sr$ (0.7048–0.7052), high $\varepsilon Nd(t)$ (+4.9 to +5.6) and $\varepsilon Hf(t)$ (+14.3 to +16.3) values, indicate that the magmas were derived from a depleted mantle source. Their $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ ratios (17.919–18.125 and 15.507–15.522, respectively) also support this interpretation. On the other hand, the Ashele basalts are enriched in Cs, Th and U, but depleted in Rb, Ba, Nb and Ta, possibly due to slight mantle metasomatism by subduction-related fluids. Because there was no record for large-scale thermal activity in the deep mantle during the Triassic, and geological evidence demonstrates reactivation of the Erqis fault, we propose that such a tectonic reactivation affected the lithospheric mantle and resulted in its partial melting and subsequent generation of the Ashele basalts. This study shows that a depleted lithospheric mantle, not an ancient continental lithosphere, was underneath the Chinese Altai in the Triassic, which was possibly accreted during the Paleozoic orogeny. Therefore, the lithospheric mantle in this area today was possibly evolved from such a depleted mantle modified by consecutive partial melting after the Paleozoic orogeny.

Keywords: Altai orogen; Triassic MORB-like basalt; Intraplate magmatism; Depleted lithospheric mantle

1. Introduction

Subcontinental lithospheric mantle plays a crucial role in sustaining the stability of continental blocks and preventing them from destruction by supercontinent cycle. Studies of mantle xenoliths, crustal geology and geochemical investigation as well as

experimental and numerical modeling have revealed that the subcontinental lithospheric mantle, especially that beneath cratons, is thick (150–250 km), cold, refractory and buoyant(Jordan, 1988; Griffin et al., 1999; Lee, 2006; O'Reilly and Griffin, 2010; Aulbach, 2012). The highly depleted compositions (i.e. high in Mg and low in Ca and Al) of cratonic mantle xenoliths suggest that the subcontinental lithospheric mantle must have experienced extensive melt extraction, during which basaltic melts and other fusible components were mostly removed from the lithospheric mantle (Pollack, 1986; Bédard, 2006). Although there is a consensus that the subcontinental lithospheric mantle is the residue of high-degree partial melting of mantle peridotite, how the lithospheric mantle formed is still a matter of hot debates. There are two categories of opinions, namely "mantle plume" and "lithospheric stacking" models. The former ascribed the refractory nature to polybaric melting of an upwelling mantle plume (Bédard, 2006; Aulbach, 2012), while the latter proposed the formation of the lithospheric mantle first at a hot ocean ridge and further conversion of the oceanic lithospheric mantle into subcontinental lithosphere (de Wit et al., 1992; Carlson et al., 2005; Lee, 2006; Pearson and Wittig, 2008; Rollinson, 2010). In spite of the big difference between the models, they may not be mutually exclusive and both need further test. To probe the nature of a subcontinental lithospheric mantle, a direct approach is to investigate the mantle xenoliths entrained in mantle-derived lavas. However, most mantle xenoliths underwent metasomatism before their arrival at the surface, and compositional estimates based on studies of mantle xenoliths are strongly biased towards metasomatized materials (Griffin et al., 1999; Shaw et al., 2007).

Another approach to mirror the nature of the subcontinental lithospheric mantle is to study the lithospheric mantle-derived magmas that may retain the information of their mantle source, and thus their petrological and geochemical characteristics would help to decipher the composition and nature of the cryptic lithospheric mantle (Turner and Hawkesworth, 1995; Lucassen et al., 2008).

The Central Asian Orogenic Belt (CAOB), also called Altaids, is located between the Kazakhstan continent and Tarim, Siberian and North China cratons (Fig. 1a), and represents the most important continental growth in the Phanerozoic (Sengör et al., 1993; Jahn et al., 2000; Jahn, 2004; Kovalenko et al., 2004). It was formed by amalgamation of juvenile terranes (e.g. island arc, oceanic plateau, backarc basin and ophiolite) with a few microcontinents and thus is an ideal site to understand the newly formed continental lithosphere (Jahn, 2004; Kröner et al., 2007; Windley et al., 2007; Zhang et al., 2018). A Seismic refraction profile across NW China has revealed a significant difference on crustal compositions, with the northern segment (Junggar Basin and Altai Mountains) being compositionally less evolved than the southern segment (Tianshan Mountains and Eastern margin of Tarim) (Wang et al., 2003). The Altai orogen is located in the northwestern area of the CAOB (Fig. 1b), which was formed by long-lasting accretion of accretionary wedges, island arcs, seamounts and ophiolites towards the southern margin of the Siberian Craton (Windley et al., 2002; Xiao et al., 2004; Utsunomiya et al., 2009). During the last decade, a large number of research work has been conducted on the Chinese Altai, and studies on the clastic, granitoid and gneissic rocks revealed juvenile material-dominated crustal

characteristics, demonstrating that the Chinese Altai represents an Early Paleozoic accretionary prism (Yuan et al., 2007; Long et al., 2008; Sun et al., 2008; Wang et al., 2009). Recently, a post-collisional tholeiitic basalt with Triassic age was reported in the SW Chinese Altai, and a preliminary study shows their affinities to mid-ocean ridge basalt (MORB) (Yuan et al., 2011). In this paper, we present new whole-rock major and trace elements and Sr–Nd–Pb–Hf isotopic data, to elucidate the origin of the basalt and to discuss the nature of the lithosphere underneath the Chinese Altai after the Paleozoic orogeny.

2. Geological background and sample description

The Altai is a NW–SE trending mountain range situated between the Sayan belt to the north and Junggar basin to the south (Fig. 1b), and is considered as the most typical region in the CAOB (Sengör and Natal'in, 1996). As a part of the mountain range, the Chinese Altai has received much attention (e.g. Windley et al., 2002; Xiao et al., 2004; Wang et al., 2009), not only for its plentiful mineral resources (Gao et al., 2018; Yang et al., 2018) but also for understanding the crustal growth in the Phanerozoic. The Chinese Altai experienced a long-lasting subduction accretionary process, and most of the tectono-stratigraphy were interpreted to be accretionary complexes formed in a forearc environment (Windley et al., 2002; Xiao et al., 2004; Zhang C. et al., 2018). Based on lithological association and stratigraphy, the Chinese Altai can be divided into several terranes or units that are separated by a series of subparallel NW–SE trending faults (Fig. 1b). According to the division scheme by Windley et al. (2002), the Chinese

Altai consists of five terranes, i.e. the Altaishan, NW Altaishan, Central Altaishan, Qiongkuer–Abagong and Erqis, which are briefly introduced below with synthesis of recent geochronological data.

The Altaishan terrane is made up of Middle to Late Devonian andesites and dacites that are conformably overlain by thick Late Devonian to Early Carboniferous metagreywackes and metasandstones. These rock associations suggest a volcanic arc with clastic detritus of a fore-arc origin (Xiao et al., 2004).

The NW Altaishan terrane mainly consists of thick metagreywackes of the Habahe Group. Timing of the sedimentation was a long-debating issue and it was widely considered as Neoproterozoic to middle Ordovician (BGMRX, 1993; Windley et al., 2002; Chen and Jahn, 2002). Recent LA-ICP-MS U–Pb analyses for detrital zircons constrained that the Habahe Group was deposited not prior to the middle Ordovician (Long et al., 2007). Overlying the Habahe Group is the Baihaba Formation dominated by andesitic rocks and volcaniclastic sediments, which were considered to be formed in a fore-arc environment (Windley et al., 2002).

The Central Altaishan terrane is mainly composed of variably metamorphic rocks and granitic intrusions, with Ordovician to Silurian sedimentary cover. The metamorphic turbiditic rocks contain both low- and high-pressure associations. The low-pressure association is dominated by chlorite–biotite–garnet–staurolite–andalusite–sillimanite–cordierite, while the high-pressure association is characterized by garnet–staurolite–kyanite–sillimanite. Based on isotopic model ages and upper intercept ages of some gneissic rocks, this terrane was considered to represent an

Archean or Paleoproterozoic microcontinent (e.g. Chen and Jahn, 2002), but in-situ zircon U–Pb analyses revealed that the paragneiss and orthogneiss are actually highly metamorphosed early Paleozoic sediments and granitic intrusions, respectively (Sun et al., 2009). Therefore, the metasediments are considered to be metamorphosed equivalent of the Habahe Group.

The stratigraphy in the Qiongkuer–Abagong terrane mainly comprises Devonian assemblages, including the lower Devonian Kangbutiebao Formation, the middle Devonian Altai Formation and the upper Devonian Qiye Formation. The Kangbutiebao Formation is composed of arc-related volcanic and pyroclastic rocks (Windley et al., 2002; Liu et al., 2010). The Altai Formation unconformably rests on the Kangbutiebao Formation, and is dominated by tuffaceous turbidites interlayered with Mg-rich dacite, and has been interpreted to be formed in an intra-oceanic island arc (Niu et al., 1999; Windley et al., 2002). This terrane experienced metamorphism of variable degrees, and an Early Permian ultra-high temperature metamorphic event has been reported in the middle of this terrane (L.X. Tong et al., 2014). Based on Nd isotope mapping of granites in the terrane, Wang et al. (2009) further divided the terrane into two parts: an younger arc assemblage (Nd model ages of 0.8–1.2 Ga) and a "continental fragment" characterized by older Nd model ages (1.1–1.6 Ga).

The Erqis terrane is a tectonic sliver along the northern side of the Erqis fault, with width less than 10 km. It is characterized by mylonitic deformation relate to the early Permian sinistral displacement (Laurent-Charvet et al., 2003). Feldspar isotopic data for the granitic gneisses and mylonites from this unit revealed contributions to the

magma sources from both mantle and upper crustal materials (Qu and Chong, 1991). MORB-type pillow basalts and island arc tholeiites imbricated with Ordovician–Silurian radiolarian chert and Silurian–Devonian turbidites, consistent with an intra-oceanic island arc setting (Xiao et al., 2004).

Our study area, Ashele, is located in the Qiongkuer–Abagong terrane. Strata in this area is dominated by the upper Devonian Qiye Formation, comprised of tuffaceous breccia, breccia andesite, spilite, quartz keratophyre and pyroclastic rocks. This Formation is heavily folded. To the north of the Ashele basalt outcrop, the strata dips over 75° to the north-east direction, while to the south, a reversed dipping was identified by Li (1998). The Ashele basalt overlies the Qiye Formation with an angular unconformity (Fig. 2) and outcrops for less than 0.1 km^2. The Ashele basaltic rocks are un-deformed and show nearly-vertical columnar joints, which are in sharp contrast to the Qiye Formation. The basalt was previously considered as a volcanic edifice within the upper Devonian Qiye Formation (Li et al., 2011), but the Ar–Ar dating indicates a late Triassic age (225 ± 6 Ma) (Yuan et al., 2011). The basalt is fine grained, fresh and homogeneous in lithology. The rocks consist of plagioclase, clinopyroxene and minor iron–titanium oxides, with an intergranular texture (Fig. 2c).

3. Analytical methods

After removing the weathered surface with an electric saw, samples were crushed into small fragments, which were further cleaned with deionized water in an ultrasonic bath and powdered in a corundum mill. Major oxides of whole-rock samples were

measured using a Rigaku ZSX100e X-ray fluorescence spectrometer on fused glass disks, while trace elements were analyzed by a Perkin-Elmer Sciex ELAN 6000 ICP-MS. Both major and trace element analyses were carried out at the State Key Laboratory of Isotope Geochemistry, Guangzhou Institute of Geochemistry, Chinese Academy of Sciences (SKLIG GIG CAS). The detailed analytical procedures were described by Li et al. (2006), and analytical uncertainties for major elements, rare earth element (REE) and other trace elements are typically 1–5%.

Bulk-rock Sr, Nd, Pb and Hf isotope analyses were performed on a Micromass Isoprobe multi-collector (MC) ICP-MS instrument at the SKLIG GIG CAS. The detailed analytical procedures followed Zhu et al. (2001), Wei et al. (2002) and Li et al. (2006). Sr and REEs were separated using DGA Spec resin column, followed by REE separation using HDEHP-coated Kef columns to obtain Nd. Measured ^{143}Nd/^{144}Nd and ^{87}Sr/^{86}Sr ratios were normalized to ^{146}Nd/^{144}Nd = 0.7219 and ^{86}Sr/^{88}Sr = 0.1194, respectively. During the analysis session for our samples, the measured values for the NBS987 Sr standard and JNdi-1 Nd standard were 0.710220 ± 10 (1σ) for ^{87}Sr/^{86}Sr and 0.512099 ± 7 (1σ) for ^{143}Nd/^{144}Nd, respectively. Pb was separated and purified with conventional anion exchange techniques (AG1X8, 200–400 resin), as described by Baker and Waight (2002). The total procedural blank was less than 0.4 ng of Pb, and the samples were doped with Tl, enabling mass discrimination correction using certified ^{203}Tl/^{205}Tl ratio of 0.418922. Repeated analyses of National Institute of Standards and Technology Standard Reference Material 981 yielded ^{206}Pb/^{204}Pb = 16.929 ± 2 (2σ), ^{207}Pb/^{204}Pb = 15.482 ± 2 (2σ), and ^{208}Pb/^{204}Pb = 36.669 ± 4 (2σ). Hf was separated and

purified using LN resin. The measured $^{176}Hf/^{177}Hf$ ratios were normalized to $^{179}Hf/^{177}Hf$ = 0.7325, and the reported $^{176}Hf/^{177}Hf$ ratios were further adjusted relative to the reference standard JMC-14374 that give $^{176}Hf/^{177}Hf$ = 0.282186 ± 8 (2σ).

4. Analytical results

The analytical results of whole-rock major and trace elements and Sr–Nd–Pb–Hf isotopes for the Ashele basalts are presented in Tables 1–3.

4.1. Major and trace elements

The Ashele basalts have variable LOI, most of which range from 1.4 to 3.9. Only two samples have relatively high LOI (5.0 and 5.9), and are not used in the following discussion. Samples of the Ashele basalt show relatively uniform compositions (Table 1), as reflected by their limited ranges of SiO_2 (48.2–53.2 wt%) and Al_2O_3 (14.4–16.3 wt%). Although the rocks have intermediate TiO_2 (0.95–1.08 wt%), their MgO (4.27–5.01 wt%) and CaO (5.68–8.79 wt%) contents are significantly lower than those of primitive oceanic basalts (Sun et al., 1979; Flower, 1991), suggesting that the Ashele rocks represent evolved melts. In comparison with typical MORB-type basalts, the Ashele basalts possess high Na_2O (2.46–4.52 wt%) but extremely low K_2O (≤0.06 wt%), indicating a tholeiitic series. In the nomenclature diagram based on immobile elements, all the samples plot into the field of sub-alkaline basalt (Fig. 3). Among the oxides, TiO_2, $Fe_2O_3^T$ and Al_2O_3 display decreasing trends with decreasing MgO (Fig. 4a–c), while CaO, Na_2O and K_2O do not show significant trend with either MgO or SiO_2 (not

shown). However, negative trends can be observed between $Fe_2O_3^T$ and TiO_2, CaO and Na_2O, and Al_2O_3 and Na_2O (Fig. 4d–f).

The rock samples have low Cr (9.47–20.7 ppm) and Ni (5.41–14.1 ppm), which, together with their low Mg# (≤ 42), suggest that they are not primary magmas. The rocks display LREE-depleted patterns (both the $(La/Yb)_N$ and $(La/Sm)_N$ ratios <1) and slightly fractionated middle REE (MREE) relative to heavy REE (HREE) ($[Gd/Yb]_N =$ 0.88–0.94) (Fig. 5a), a typical feature of MORB-type basalts (Arevalo and McDonough, 2010). In comparison with the mean global N-MORB (MgO $\approx$ 8 wt%) (Arevalo and McDonough, 2010), the Ashele rocks have relatively high Ba, Th, U, Sr and Pb, but distinctively low REE and high field strength element (HFSE) concentrations. In particular, their low Zr contents (16.6–20.1 ppm), together with LREE-depleted characteristics, make them akin to low-Zr basalts in the Izu–Bonin arc (Tamura et al., 2005, 2007). On the primitive mantle normalized spider diagram, all the rocks show relatively enrichments of Cs, Th, U and Sr, with remarkable troughs at Nb, Ta, Zr and Hf (Fig. 5b), which are typically subduction-related signatures. In contrast to their relatively high Cs and Pb contents, the rocks nevertheless display prominent troughs at Rb and K, which are characteristics for MORB-type basalts (Arevalo and McDonough, 2010). There are two more distinctive features for the rocks. Firstly, all the rocks exhibit positive Eu anomalies, with Eu/Eu* ratios between 1.14 and 1.26. Secondly, their relatively high V (365–427 ppm) and correspondingly high V/Sc ratios (8.8–9.3) are notably higher than those of MORB (V/Sc = 6.7, Lee et al., 2005). Their relatively high V contents cause remarkably low Ti/V ratios ($Ti \times 10^{-4}/V$ = 14.7–17.1), which, along

with their low Zr/Y ratios (1.2–1.3), are common for arc-related rocks (Fig. 6a). Besides, the rocks possess considerably low Nb/Ta (11.0–13.7) and Zr/Hf (24.6–26.3) ratios, which are comparable to those of the Papua new Guinea and the Cyprus boninites and basalts and displaced from the MORB field (Fig. 6b) (König et al., 2010).

4.2 Isotopic compositions

The Ashele basalts possess relatively consistent Nd and Hf isotope compositions, with $\varepsilon Nd(t)$ and $\varepsilon Hf(t)$ values ranging from +4.9 to +5.6 and from +14.3 to +16.3, respectively. In contrast with their Nd and Hf isotopes, the rock samples display more radiogenic Sr isotope compositions (initial $^{87}Sr/^{86}Sr$ ratios = 0.7048–0.7052) (Table 2), which make plots of the samples deviate from the mantle array (Fig. 7a). However, their $^{87}Sr/^{86}Sr$ ratios remain constant with LOI values (not shown), suggesting that post-magmatic alteration did not play a significant role in their Sr isotopic compositions. In the $\varepsilon Hf(t)$ vs. $\varepsilon Nd(t)$ diagram, all the samples plot into the field of Indian mantle domain (Fig. 7b). For the Pb isotopes, the rock samples show relatively homogeneous $^{206}Pb/^{204}Pb$ (17.919–18.125), $^{207}Pb/^{204}Pb$ (15.507–15.522) and $^{208}Pb/^{204}Pb$ (37.740–37.910) ratios (Table 3), which are consistent with the Pb isotope compositions of various mafic sequences and pegmatitic rocks in the Chinese Altai (Chiaradia et al., 2006). In the $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram (Fig. 7c, d), all the rock samples plot in the range of the Australian–Antarctic Discordance (AAD) and well within the Indian mantle domain (Kempton et al., 2002; Pearce et al., 2007).

5. Discussion

5.1. Effects of alteration and crustal contamination

The rocks are homogeneous without visible vesicles, amygdloids or veins. Their fine-grained texture suggests a low permeability, which may prevent broad hydrothermal alteration. Most samples exhibit low loss on ignition (LOI < 3 wt%), and there is no evident correlations between LOI and highly mobile incompatible element ratios (e.g. Rb/Ba and Rb/Cs), implying a limited effect of hydrothermal alteration on the trace element compositions of the rocks.

Because crustal rocks are relatively enriched in large ion lithophile elements (LILEs, e.g. K, Rb, Cs, Ba), assimilation of crustal materials would lead to distinct enrichment of LILEs relative to HFSEs. All the basaltic samples have low K_2O (≤0.06 wt%) contents (Table 1), which are incompatible with remarkable involvement of crustal materials. On the primitive mantle normalized diagrams (Fig. 5b), the Ashele basalts display coherent patterns and all show troughs at K and Rb, and their Ba concentrations are roughly equivalent to, or slightly lower than the level of HFSE and HREE, demonstrating a negligible contribution of crustal materials. More importantly, the samples show LREE-depleted patterns and have positive εHf(t) and εNd(t) values, which are typical features of the depleted mantle and do not support significant assimilation of crustal materials. We therefore consider that the trace element characteristics most likely reflect the mantle source signature and magmatic process during their ascent.

5.2. Nature of the mantle source

The Ashele basalts show LREE-depleted patterns (Fig. 5a), a typical feature of MORB, and their subchondritic HREE distribution ($(Gd/Yb)_N = 0.88–0.94$) indicates a limited fractionation of hornblende in the spinel stability field. In addition, their Nb–Ta troughs and Cs–U–Th spikes suggest involvement of slab-derived fluids in their mantle source. As fluid-mobile elements, LILEs can be readily transferred to the overlying lithosphere during slab dehydration. In spite of their high mobility, LILEs in the Ashele basalts behave differently. The Ashele basalts show enrichment of Cs, Th and U relative to REEs and HFSEs, while K and Rb display significant depletion (Fig. 5b). The prominent depletion of Rb and K relative to Cs in the Ashele basalts may reflect the dehydration process in the presence of phengite (Melzer and Wunder, 2000; Wunder and Melzer, 2003) and imply a relatively shallow dehydration depth, probably in a forearc setting. A phengite–fluid partitioning experiment simulating the dehydration process beneath the volcanic front in island arcs (2–4 GPa) demonstrates that Rb preferentially fractionates into phengite, whereas Cs tends to partition into fluids (Melzer and Wunder, 2000; Wunder and Melzer, 2003). In comparison with arc basalts with equivalent MgO, the Ashele basalts have relatively high V/Sc ratios (8.8–9.3), probably indicating an oxygen fugacity above the FMQ and reflecting the joint contribution of slab-derived fluids (Brandon and Draper, 1996; Bénard et al., 2018) and fractional crystallization of clinopyroxene (Lee et al., 2005).

The εHf(t) values (+14.3 to +16.3) of the basaltic samples are well within the range of MORB (+12 to +17) (Chauvel and Blichert-Toft, 2001), affirming a highly depleted

mantle source. The basalts also show depleted mantle-like Nd isotopic compositions, but their $\varepsilon Nd(t)$ values (+4.9 to +5.6) are lower than those of Pacific and Atlantic MORB (+8 to +12) (Chauvel and Blichert-Toft, 2001 and references therein), reflecting a decoupling between the Nd and Hf isotopes (Fig. 7b). Such decoupled Nd–Hf isotopes have been reported for Devonian mafic rocks in the Chinese Altai, and interpreted to be resulted from partial melting of a depleted mantle metasomatized by slab fluids from subducted sediments (Yu et al., 2017). Slab fluids are enriched in Nd relative to Hf, so change Nd isotopic compositions of the mantle to a much higher degree. The relatively variable $^{87}Sr/^{86}Sr$ ratios make the plots of the basaltic samples deviate away from the mantle array in Nd–Sr isotope space and lead to a trend nearly parallel to the X-axis (Fig. 7a), probably suggesting the incorporation of Sr from seawater in the source. In the $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram (Fig. 7c, d), all the basaltic samples plot in the field of Indian MORB, consistent with the Nd–Hf isotope characteristics.

Based on the mobility of elements during slab dehydration, trace elements can be classified into 'conservative' and 'non-conservative' (Pearce and Peate, 1995). The LILEs and LREE to MREE are highly non-conservative and slightly non-conservative, respectively, regardless of aqueous fluids or siliceous melts (Pearce and Peate, 1995). Commonly, Rb–Sr and U–Th–Pb are highly non-conservative because of their high ionic radius, while Sm–Nd are slightly non-conservative. The Elements Lu and Hf, however, behave differently in different medias, namely Lu and Hf generally behave conservatively in aqueous fluids, whereas in siliceous melts Hf can be moderately–slightly non-conservative (Pearce and Peate, 1995). The different behaviors of these

elements during slab dehydration can significantly affect their corresponding isotope systems in arc magmas. Altered oceanic lithosphere is the main source of slab-derived aqueous fluids. Recent investigations on oceanic basalts and serpentinized abyssal peridotites have revealed that Hf isotopes and Lu/Hf ratios are relatively unaffected, while Nd and Sr isotopic ratios have been significantly changed during the interaction of ocean floor with seawater (Thompson et al., 2008; Frisby et al., 2016), suggesting significant uptake of Nd–Sr from seawater to altered rocks. This is exactly the case for the Ashele basalts, and their Sr–Nd–Hf isotopic variations are consistent with the order of their mobility in aqueous fluids, namely Sr > Nd > Hf.

Like Sr, Pb is also a highly non-conservative element and strongly partitioned into aqueous fluids (Pearce and Peate, 1995). Therefore, Sr and Pb isotopic ratios are expected to display coupling change in mantle melting fluxed by slab-derived fluids (Ishizuka et al., 2003). In comparison with Sr isotope ratios that display prominent drift away from the mantle array, however, Pb isotopic ratios of the Ashele basalts seem to be less affected and plot in the MORB field (Fig. 7c, d). The inconsistency between Sr and Pb isotopic compositions has been observed in tholeiites of the Scotia and Marina arcs (Meijer, 1976; Hawkesworth et al., 1977), and could be ascribed to their different contents in deep seawater (Sr = 7.5 μg/g; Pb = 1 pg/g, Frank, 2002), which give rise to a limited impact on Pb isotopic compositions of arc magmas. All the above evidence suggests that the Ashele basalts were derived from a highly depleted mantle source that had been metasomatized by slab-derived aqueous fluids, i.e. the Ashele basalts originated from an ancient forearc mantle wedge.

5.3. Genesis of positive Eu anomalies

Positive Eu anomalies are widely reported for basalts from mid-ocean ridge, intraplate and arc-related tectonic settings (Hawkesworth et al., 1977; Vukadinovic, 1993), and high-precision trace element analyses on glass samples of oceanic floor basalts have revealed that positive Eu anomalies occur only in depleted and primitive samples and are commonly associated with positive Sr anomalies (Niu and O'Hara, 2009; Arevalo and McDonough, 2010; Jenner and O'Neill, 2012; Tang et al., 2017). For intraplate mafic–ultramafic intrusions, since the magma passes through the lower continental crust characterized by Eu excess (Taylor and Mclennan, 1995), positive Eu anomalies in some layered mafic to ultramafic intrusions have been attributed to contamination of granulitic rocks in the lower crust (e.g. Mazzucchelli et al., 1992; Hermann et al., 2001). Given that the coupled Eu–Sr anomalies are only present in primitive oceanic floor basalts (MgO > 9.5 wt.%), it has been suggested that the coupled Eu–Sr anomalies in ocean floor basalts were inherited from a depleted MORB mantle (DMM) with excess Sr and Eu (Niu and O'Hara, 2009; Jenner and O'Neill, 2012), or recycled lower continental crust (Tang et al., 2015). The coupled positive Eu–Sr anomalies in arc and MORB basalts commonly reflect the process of plagioclase accumulation and resorption in ponded magma chamber (Vukadinovic, 1993; Hermann et al., 2001). With the fractional crystallization of plagioclase, the positive Eu anomalies in basalts will fade away (Hawkesworth et al., 1977; Arevalo and McDonough, 2010), confirming the role of plagioclase in the occurrence of coupled

Sr–Eu anomalies. Woodhead et al. (1998) noticed that some arc basalts with positive Eu–Sr anomalies do not contain high Al_2O_3 and excessive plagioclase phenocrysts, which makes the plagioclase accumulation regime questionable. Considering that Eu^{2+} has ion radii and coordination similar to those of Sr^{2+} and Pb^{2+}, they proposed that Eu^{2+} could readily be transported with Sr^{2+} and Pb^{2+} in slab-derived aqueous fluids and give rise to the positive anomalies of these elements in primitive arc magma (Woodhead et al., 1998).

Although the Ashele basalts formed in a terrestrial environment, their extremely depleted and uniform geochemical compositions, LREE-depleted patterns and radiogenic Nd–Hf isotope characteristics exclude the possibility of contamination by the lower crust. In the Eu/Eu* versus La/Yb diagram, the Ashele basalts show consistently low Eu/Eu* and La/Yb ratios, in contrast to the lower-crust contaminated basalts that possess variable and positive correlation of Eu/Eu* with La/Yb ratios (Fig. 8a), suggesting negligible crustal contamination. Although the Ashele basalts display positive Eu–Sr anomalies, their MgO contents (< 5.0 wt%) are remarkably lower than those of primitive arc and ocean floor basalts. Meanwhile, with decreasing MgO contents (evolution of magma), the Eu/Eu* ratios of the Ashele basalts increase (Fig. 8b), which is significantly different from those of primitive arc basalts that commonly show opposite trend (e.g. Hawkesworth et al., 1977; Woodhead et al., 1998). This may imply a distinct regime for the positive Eu anomalies in the Ashele basalts.

Based on some experimental results showing faster diffusion of divalent Eu and Sr than trivalent Pr, Nd, Sm and Gd in clinopyroxene, Tang et al. (2017) invoked the

kinetic effect of disequilibrium melting of spinel peridotite in the MORB mantle source to account for the excess of Eu and Sr in the primitive melt. According to their model, melt fractions produced during early- and late-stage melting may carry positive and negative Eu–Sr anomalies, respectively. This is apparently contradictory to what is shown in the low-MgO Ashele basalts, and hence cannot be the cause either. Metasomatism by slab-derived fluids is an alternative process to generate positive Eu–Sr anomalies in primitive arc basalts, especially for those lack of plagioclase (Woodhead et al., 1998). The Ashele basalts were derived from a depleted mantle source that experienced fluid metasomatism, however, the basalts exhibit pronounced signature of plagioclase. In the diagram of Al_2O_3 versus MgO (Fig. 4b), the Ashele basalts show an increasing trend of Al_2O_3 with the decreasing MgO, which is distinctively different from those of MORB and primitive arc basalts (Hawkesworth et al., 1977; Woodhead et al., 1998; Jenner and O'Neill, 2012). Instead, the negative Al_2O_3–MgO relationship most likely reflects a retarded crystallization of plagioclase under high water contents (Müntener et al., 2001), consistent with the mantle source hydrated by slab fluids.

5.4. Implications for the growth of continental lithosphere

In the interior of continents, LREE-depleted basalts are sporadic except for a few cases for some peculiar plume-related basalts that have been attributed to the involvement of a recycled oceanic lithosphere component (e.g. Barrat et al., 2003). LREE-depleted basalts can also be found in the volcanic fronts of modern subduction

zones (e.g. Izu Bonin arc), where slab-derived fluids would effectively decrease the solidus of the overlying lithosphere and give rise to N-MORB basalts (Taylor and Nesbitt, 1988; Duggen et al., 2004). The Ashele basalts formed in the late Triassic (225 ± 6 Ma), posterior to the closure time (i.e. early Permian) of the Altai branch of the Paleo-Asian Ocean (Li et al., 2014; Y. Tong et al., 2014), suggesting that the Ashele basalts formed in an intraplate setting.

Mantle melting in intraplate settings is usually caused by anomalously high temperature (e.g. the arrival of mantle plume) (Hofmann 1997). The arrival of a hot, ascending mantle plume will cause extensive mantle and crustal melting as well as voluminous metal mineralization over a short timescale (generally 1–5 Ma) (Bryan and Ernst 2008; Pirajno et al. 2009). However, there is absence of coevally plume-related flood basalts, metallogenesis or high-temperature magmas, indicating that a Triassic deep mantle plume was not evident beneath the Altai area. A possible effect of Siberian mantle plume can be precluded as the Ashele basalts erupted significantly later than the volcanism (ca. 250 Ma) in the Siberian large igneous province (Ivanov et al., 2013). Small-volume intraplate basalts could have originated from gravitational collapse of lower lithosphere, and this mechanism has been advocated for the genesis of Cretaceous basalts in the Gobi Altai with increasing Nb/La ratios (Sheldrick et al., 2018). The abrupt increase in $(Nb/La)_{PM}$ ratios (from 0.33–0.82 to 0.93–1.64) at about 100 Ma reflects the change in magma source from a lithospheric to an asthenospheric mantle as a response to the lithospheric delamination (Sheldrick et al., 2018). However, the Ashele basalts have low $(Nb/La)_{PM}$ (0.17–0.19) ratios, distinctively different from those

for the basalts formed by lithospheric delamination (Fig. 9).

Since the Altai orogen is situated between the rigid Junggar and Hangai blocks, the generation of the Ashele N-MORB-like basalts could be related to the tectonic reactivation. Actually, the Ashele basalts erupted in an area close to the Erqis fault, which is one of the largest strike-slip faults in the CAOB and underwent multiple reactivation during the late Permian and early Mesozoic (Briggs et al., 2007; Li et al., 2015). Briggs et al. (2009) conducted ^{40}Ar–^{39}Ar thermochronological studies on muscovite, biotite, plagioclase and K-feldspar as well as Th/Pb ion-microprobe dating on monazite using samples collected along this fault. Their data demonstrate that, from Permian to Jurassic, the Erqis fault experienced at least 3 peaks of activity (Briggs et al., 2009). Thereinto, the second peak of activity is recorded by muscovite, biotite and K-feldspar Ar–Ar systems, which took place during late Triassic (230–209 Ma) and was coeval with the eruption of the Ashele basalt. Mantle melting can be induced by strike-slip fault activity via different mechanisms such as decompressional melting and frictional heating. Strike-slip faults could create localized zones of extensional stress, and may introduce decompressional melting (Latin and White, 1990). This scenario has been documented in the Tazheran–Olkhon area within the eastern Baikal Rift, where coeval low-K tholeiitic and calc-alkaline mafic magmatic activities occurred in a staggered order following a series of faults separating sigmoid-shaped tectonic slivers (Fedorovsky et al., 2010). On the other hand, the intense friction accumulated during the movement of fault surfaces leads to heating, which may account for a significant proportion of total fault energy release budget (Brown, 1998; Kanamori, 2001). The

existence of fault fractures also provides channels of hydrous flow which reduces the solidus of silicate minerals, facilitating partial melting (Otsuki et al., 2003). High-pressure friction experiments (Del Gaudio et al., 2009) and studies of glass materials developed on the fault surfaces (Obata and Karato, 1995; Andersen and Austrheim, 2006) as well as numerical simulation of fault structures (Connor et al., 2000) all demonstrate that frictional heating of fault activity can trigger mantle melting. As revealed by seismic data, the Erqis fault dips northward, passes beneath the Ashele basalts, and penetrates into the lithospheric mantle at deeper levels (Zhang et al., 1992; Xu et al., 2001; Windley et al., 2002). Therefore, we propose that the late Triassic activation of the Eriqs fault (Briggs et al., 2009) caused the eruption of the Ashele basalts, which were generated by partial melting of the juvenile depleted lithospheric mantle possibly accreted underneath the Chinese Altai during the Paleozoic orogeny. Geological evidence shows that the Triassic activation of the Erqis fault was possibly due to the continuous convergence of the Altai orogen with the Junggar terrane (Briggs et al., 2009).

Hot debate exists on the nature of the Chinese Altai, e.g. it was considered as a micro-continent (Hu et al., 2000, 2002; Li et al., 2006) or an active margin (Long et al., 2007; Yuan et al., 2007; Sun et al., 2008). Based on whole-rock Sm–Nd isochron dating (Hu et al., 2002; Li et al., 2006) and Nd isotopic model ages (Hu et al., 2000), gneissic rocks in the Chinese Altai were interpreted to be formed during Paleo- to Meso-Proterozoic and therefore represent Precambrian basements. However, more recent work on single-spot zircon U–Pb dating and Hf isotopic geochemistry for gneissic rocks

(Sun et al., 2008) and meta-sediments (Long et al., 2007; Jiang et al., 2011; Dong et al., 2018) as well as granitoids (Cai et al., 2011) reveal predominantly Paleozoic ages (543–368 Ma) and highly positive εHf(t) values. Compared to the old isotopic dating methods (i.e. Sm–Nd isochrons and Nd model ages), modern single-spot ablation zircon U–Pb dating offers significantly higher accuracy and reliability (Jackson et al., 2004). Consequently, the above zircon U–Pb dating results and Hf isotopic compositions suggest a juvenile continental mass and absence of Precambrian basement. The previous studies regarding the nature of the Chinese Altai are largely focused on meta-sediments and granitoids of a crustal origin, and less attention has been paid to mafic rocks originated from the mantle. Our new data for the Triassic Ashele basalts demonstrate that a depleted mantle lithosphere was underneath the Chinese Altai in the Triassic. Backed up by the fact that Precambrian outcrops are absent in the Chinese Altai, results of this study further enhance the conclusions from our previous studies of Paleozoic rocks that the Chinese Altai cannot be a micro-continent. This study shows that the depleted mantle was a dominant, if not the only, component in the lithospheric mantle during the Triassic underneath the Chinese Altai, partial melting of which in the Triassic and later time may have changed its composition toward that of continental lithosphere.

The formation of continental lithospheric mantle has been explained by either mantle melting via plume activities (Herzberg, 1993; Griffin and O'Reilly, 2007) or transformation from depleted lithospheric mantle (Helmstaedt and Schulze, 1989; Canil, 2004; Parman et al., 2004). As discussed above, a mantle plume activity is not favored

beneath the Chinese Altai during the Triassic. Instead, the elemental and isotopic compositions of the Ashele basalt suggest a depleted mantle origin. This supports the transformation of depleted lithospheric mantle to continental lithospheric mantle. After the Paleozoic orogeny, lithospheric mantle underneath the Chinese Altai may include the fossil mantle wedge and the under-thrust or imbricated lithosphere of the subducted plate (Helmstaedt and Schulze, 1989; Parman et al., 2004). The depletion of HFSE for the Ashele basalts may imply derivation from the fossil mantle wedge, and the depletion in LREE may indicate that the incompatible elements were extracted by multiple melting of the mantle source during prolonged Paleozoic orogenesis. The further partial melting of the lithospheric mantle after the orogeny, such as the eruption of the Ashele basalts, may progressively change its composition toward the continental lithosphere.

6. Conclusions

The tholeiitic Ashele basalts in the Chinese Altai were erupted in the Triassic, and are featured by extremely low K_2O (≤ 0.06 wt%) contents and LREE-depleted patterns (both $[La/Sm]_N$ and $[La/Yb]_N$ ratios < 1). Their Sr–Nd–Pb–Hf isotopic compositions ($^{87}Sr/^{86}Sr = 0.7048$–0.7052; $\varepsilon Nd(t) = 4.9$–5.6, $\varepsilon Hf(t) = 14.3$–16.3, $^{206}Pb/^{204}Pb = 17.919$–18.125, $^{207}Pb/^{204}Pb = 15.507$–15.522) indicate depleted mantle as their magma source. This study shows the existence of a depleted lithospheric mantle, not ancient continental lithosphere, in the Triassic underneath the Paleozoic orogen. Taking into account of regional geology, we consider that the Triassic reactivation of the Erqis fault resulted in mantle melting and subsequent genesis of the Ashele basalts. Consecutive partial

melting of the depleted mantle may have changed its composition toward continental lithosphere.

Acknowledgements

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Figure captions:

Fig. 1. (a) Simplified tectonic map of the CAOB (modified after Jahn et al. 2000). (b) Sketch map of the Chinese Altai orogen (modified after Windley et al. 2002). The number 1–5 represent separated terranes, namely the Altaishan, the NW Altaishan, the Central Altaishan, the Qiongkuer–Abagong and the Erqis, respectively.

Fig. 2. Photographs of outcrop (a–b) and mineralogical (c) features of the Ashele basalts. Cpx and Pl represent clinopyroxene and plagioclase, respectively.

Fig. 3. (a) Total alkali–silica (TAS) diagram (after Le Bas et al. 1986), and (b) Zr/Ti–Nb/Y diagram (after Winchester and Floyd, 1977) for the Triassic basalts in the Chinese Altai.

Fig. 4. Binary diagrams showing variations in major oxides of the Ashele basalts.

Fig. 5. Chondrite-normalized REE patterns (a) and primitive mantle-normalized multielement diagrams (b) for the Ashele basalts from the Chinese Altai. Normalizing values are from Sun and McDonough (1989).

Fig. 6. (a) Ti–V and (b) Nb/Ta–Zr/Hf (afer König et al. 2010) diagrams for the Ashele basalts from the Chinese Altai.

Fig. 7. (a) εNd(t) versus $(^{87}Sr/^{86}Sr)_i$ (after Zindler and Hart, 1986), (b) εNd(t) versus εHf(t) (after Kempton et al. 2002), (c) $^{208}b/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ (after Reagan et al., 2010) and (d) $^{208}b/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ (after Pearce et al., 2007) diagrams for the Ashele basalts from the Chinese Altai. AAD represents Australian–Antarctic Discordance.

Fig. 8. (a) δEu versus $(La/Yb)_N$ (after Mazzucchelli et al. 1992) and (b) δEu versus MgO diagrams for the Ashele basalts from the Chinese Altai.

Fig. 9. Binary diagram of $(Nb/La)_{PM}$ versus Age (Ma) for the basaltic rocks in the Altai orogen. Normalizing values are from Sun and McDonough (1989). Geochemical data are from Cai et al. (2007), Zhang et al. (2010), Wan et al. (2013), Yu et al. (2017) and Sheldrick et al. (2018).

Table 1 Bulk-rock major oxides (wt.%) and trace element (ppm) abundances of the Ashele basalt

Sample	13ASL02	13ASL04	13ASL05	13ASL07	13ASL08	13ASL09	13ASL10	13ASL11	13ASL12	13ASL13
SiO_2	52.0	50.7	52.0	52.3	51.3	50.1	50.7	50.9	52.5	52.3
TiO_2	0.98	0.99	0.95	1.03	1.00	0.97	1.05	1.02	1.04	1.01
Al_2O_3	15.1	15.2	15.4	14.9	15.4	14.6	15.1	15.2	15.5	15.3
$Fe_2O_3^T$	13.6	12.9	13.2	13.6	13.2	13.5	13.6	13.9	14.0	13.6
MnO	0.20	0.18	0.18	0.20	0.20	0.19	0.19	0.19	0.20	0.19
MgO	4.76	4.48	4.84	4.53	4.73	4.70	4.64	4.73	4.87	4.62
CaO	6.63	5.84	7.33	6.78	7.73	5.62	5.90	5.68	6.15	6.48
Na_2O	3.74	4.19	3.46	3.86	2.48	3.86	4.52	4.09	4.04	3.84
K_2O	0.04	0.05	0.05	0.05	0.05	0.05	0.05	0.05	0.04	0.04
P_2O_5	0.06	0.05	0.05	0.06	0.05	0.05	0.05	0.06	0.06	0.06
L.O.I	2.51	5.02	2.15	2.31	3.54	5.95	3.84	3.93	1.35	2.25
Total	99.61	99.63	99.64	99.66	99.67	99.68	99.69	99.70	99.71	99.72
Sc	42.6	41.1	42.7	42.5	42.5	43.4	41.0	43.7	43.9	42.9
V	390	370	383	385	375	386	365	390	391	390
Cr	13.3	15.0	13.1	15.6	12.0	12.5	12.1	14.4	11.5	14.2
Co	38.3	36.4	34.1	29.6	29.7	30.0	34.2	34.1	34.7	33.6
Ni	6.02	6.66	6.38	6.20	6.40	6.05	5.87	6.46	6.30	5.99
Cu	44.5	44.5	46.4	46.3	45.9	44.5	46.5	47.1	45.9	44.8
Zn	101	89.0	96.4	95.5	89.7	101	88.6	96.1	101	92.6
Ga	17.0	15.2	16.8	16.2	17.0	16.6	15.1	16.5	16.9	16.3
Ge	2.86	2.61	2.93	2.84	2.76	2.68	2.57	2.75	2.78	2.86
Rb	0.17	0.25	0.30	0.35	0.33	0.41	0.16	0.33	0.32	0.28
Sr	191	127	189	168	234	146	76.0	140	153	165
Y	14.7	14.1	14.6	14.1	14.4	14.3	13.8	14.5	14.5	14.6
Zr	17.9	17.6	18.2	18.4	17.7	18.3	16.6	17.9	18.8	18.2
Nb	0.38	0.37	0.38	0.37	0.37	0.39	0.35	0.36	0.38	0.39
Cs	0.08	0.25	0.13	0.14	0.10	0.11	0.15	0.13	0.10	0.12
Ba	21.4	33.9	34.8	39.0	25.6	25.8	22.0	31.2	27.5	27.3
La	2.13	1.89	1.90	1.93	1.94	1.94	1.79	2.01	2.06	1.95
Ce	5.52	4.95	4.99	4.98	5.00	5.04	4.73	5.08	5.23	5.07
Pr	0.84	0.79	0.80	0.81	0.79	0.81	0.76	0.82	0.83	0.81
Nd	4.39	4.25	4.27	4.28	4.25	4.33	4.07	4.39	4.46	4.35
Sm	1.57	1.49	1.55	1.54	1.52	1.52	1.49	1.57	1.57	1.55
Eu	0.70	0.68	0.68	0.67	0.65	0.66	0.66	0.70	0.69	0.67
Gd	2.02	1.98	2.03	2.00	1.98	2.04	1.93	2.02	2.05	2.07
Tb	0.40	0.39	0.40	0.39	0.39	0.41	0.38	0.40	0.40	0.40
Dy	2.67	2.62	2.70	2.64	2.61	2.71	2.55	2.71	2.71	2.73
Ho	0.62	0.60	0.63	0.60	0.60	0.61	0.58	0.61	0.61	0.62
Er	1.80	1.72	1.83	1.76	1.75	1.81	1.73	1.75	1.76	1.80
Tm	0.28	0.27	0.28	0.27	0.28	0.28	0.27	0.28	0.28	0.28
Yb	1.83	1.71	1.83	1.78	1.82	1.80	1.72	1.83	1.85	1.80
Lu	0.29	0.27	0.29	0.29	0.29	0.29	0.28	0.29	0.30	0.29
Hf	0.70	0.70	0.73	0.73	0.70	0.74	0.67	0.72	0.75	0.73
Ta	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.04
Pb	2.68	1.88	2.45	2.76	2.56	1.98	2.16	3.17	2.68	3.38
Th	0.43	0.43	0.45	0.44	0.43	0.45	0.41	0.45	0.46	0.44

U	0.18	0.20	0.20	0.21	0.29	0.22	0.18	0.21	0.22	0.22
Mg#	40.9	40.8	42.1	39.7	41.5	40.8	40.3	40.3	40.8	40.2

Mg#=100*Mg/(Mg+Fe^{2+}), assuming Fe^{2+}/Fe^{total} =0.90; LOI=loss of ignition; Fe$_2$O$_3^T$ =total Fe oxides as Fe$_2$O$_3$.

Table 1 continued

Sample	13ASL14	13ASL15	13ASL16	13ASL17	13ASL18	13ASL19	13ASL20	13ASL21	13ASL22
SiO$_2$	53.2	51.7	48.2	50.5	51.3	50.1	52.2	52.3	53.1
TiO$_2$	0.97	1.00	1.10	1.06	1.00	1.04	1.00	0.97	1.01
Al$_2$O$_3$	14.4	15.2	16.3	15.1	15.6	15.6	15.2	15.3	15.1
Fe$_2$O$_3^T$	12.9	13.8	14.8	13.8	13.4	14.3	13.6	13.4	13.3
MnO	0.17	0.20	0.19	0.20	0.17	0.20	0.19	0.18	0.19
MgO	4.27	4.82	4.93	5.01	4.27	4.72	4.48	4.68	4.53
CaO	8.39	6.12	6.69	8.65	8.79	7.81	7.71	7.38	7.26
Na$_2$O	3.54	3.94	3.02	2.75	2.46	3.08	2.78	3.12	3.21
K$_2$O	0.04	0.04	0.06	0.04	0.03	0.04	0.03	0.04	0.04
P$_2$O$_5$	0.06	0.06	0.06	0.06	0.06	0.06	0.06	0.05	0.06
L.O.I	1.72	2.78	4.39	2.62	2.66	2.81	2.62	2.34	2.05
Total	99.73	99.74	99.75	99.76	99.77	99.78	99.79	99.80	99.81
Sc	41.2	41.2	47.1	43.6	42.4	46.2	41.4	42.5	42.6
V	374	383	427	393	383	425	378	387	387
Cr	15.2	12.9	9.47	11.6	14.0	15.0	13.9	9.7	20.7
Co	30.9	33.1	33.8	36.2	22.6	37.1	30.6	36.0	36.4
Ni	10.9	5.41	5.63	5.57	10.4	10.7	8.43	5.82	14.1
Cu	42.1	45.2	43.2	41.4	45.0	45.4	43.1	46.8	45.6
Zn	92.1	95.1	109	101	87.9	95.6	93.8	100	95.2
Ga	16.1	16.4	18.0	17.2	17.1	17.5	16.6	16.8	16.9
Ge	2.86	2.69	2.99	2.86	2.68	2.89	2.69	2.72	2.83
Rb	0.27	0.08	0.11	0.10	0.10	0.08	0.10	0.10	0.10
Sr	219	222	168	180	187	165	164	175	181
Y	14.3	14.3	16.3	15.7	14.4	16.8	14.3	14.5	14.7
Zr	17.9	17.4	20.1	18.9	18.8	19.7	18.2	18.2	18.6
Nb	0.38	0.37	0.42	0.39	0.37	0.40	0.38	0.37	0.38
Cs	0.10	0.12	0.21	0.34	0.19	0.13	0.17	0.14	0.20
Ba	24.2	10.3	18.3	21.4	17.8	21.2	18.8	22.7	25.9
La	1.94	1.88	2.12	1.97	1.94	2.14	1.91	1.86	1.99
Ce	4.98	4.95	5.59	5.20	5.01	5.59	5.03	4.98	5.20
Pr	0.79	0.79	0.89	0.84	0.80	0.90	0.80	0.80	0.83
Nd	4.26	4.16	4.84	4.50	4.22	4.82	4.34	4.32	4.47
Sm	1.50	1.52	1.71	1.58	1.53	1.71	1.54	1.52	1.58
Eu	0.67	0.65	0.72	0.68	0.69	0.74	0.68	0.68	0.75
Gd	1.98	1.99	2.25	2.11	1.97	2.27	2.02	2.04	2.08
Tb	0.39	0.39	0.45	0.42	0.39	0.45	0.40	0.40	0.40
Dy	2.62	2.66	2.98	2.83	2.63	3.03	2.68	2.74	2.76
Ho	0.59	0.60	0.68	0.64	0.60	0.71	0.62	0.63	0.63
Er	1.72	1.74	1.98	1.88	1.74	2.04	1.78	1.81	1.81
Tm	0.27	0.27	0.31	0.29	0.27	0.32	0.28	0.28	0.28
Yb	1.75	1.74	2.02	1.89	1.78	2.02	1.81	1.85	1.84
Lu	0.28	0.29	0.32	0.30	0.29	0.33	0.29	0.29	0.30
Hf	0.71	0.67	0.81	0.75	0.72	0.76	0.71	0.74	0.73
Ta	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03
Pb	2.78	9.6	2.04	2.18	2.63	2.12	2.66	8.11	2.34
Th	0.43	0.43	0.50	0.46	0.44	0.48	0.45	0.45	0.44

U	0.20	0.23	0.27	0.22	0.21	0.20	0.19	0.19	0.20
Mg#	39.6	40.8	39.8	41.9	38.7	39.6	39.6	40.8	40.3

Table 2 Rb-Sr and Sm-Nd isotopic analytical results for the Ashele basalt

Sample	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	2σ	$(^{87}Sr/^{86}Sr)i$	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	2σ	$^{143}Nd/^{144}Ndi$	εNd(t)
13ASL02	0.170	191	0.002580	0.70521	0.000008	0.705200	1.57	4.39	0.215571	0.512928	0.000006	0.512611	5.12
13ASL04	0.253	127	0.005764	0.70509	0.000008	0.705072	1.49	4.25	0.212219	0.512948	0.000005	0.512635	5.60
13ASL05	0.301	189	0.004616	0.70508	0.000008	0.705069	1.55	4.27	0.219132	0.512948	0.000005	0.512625	5.40
13ASL06	0.340	165	0.005948	0.70519	0.000009	0.705170	1.54	4.28	0.217322	0.512949	0.000007	0.512629	5.48
13ASL08	0.331	234	0.004095	0.70495	0.000009	0.704941	1.52	4.25	0.216082	0.512936	0.000006	0.512618	5.27
13ASL09	0.409	146	0.008089	0.70514	0.000011	0.705110	1.52	4.33	0.212484	0.512954	0.000008	0.512642	5.72
13ASL10	0.164	76	0.006243	0.70522	0.000007	0.705202	1.49	4.07	0.220790	0.512947	0.000007	0.512622	5.35
13ASL11	0.329	140	0.006800	0.70516	0.000011	0.705135	1.57	4.46	0.212826	0.512944	0.000007	0.512631	5.51
13ASL13	0.275	165	0.004834	0.70511	0.000009	0.705098	1.55	4.35	0.215350	0.512961	0.000006	0.512644	5.77
13ASL14	0.265	219	0.003505	0.70513	0.000007	0.705119	1.50	4.26	0.212606	0.512938	0.000007	0.512625	5.40
13A	0.08	222	0.00	0.70	0.00	0.70	1.52	4.16	0.221	0.512	0.00	0.512	5.4

13ASL15	0		1045	510	0008	5092			043	951	0006	626	2
13ASL16	0.110	168	0.001894	0.70487	0.000010	0.704859	1.58	4.50	0.212586	0.512961	0.000006	0.512648	5.85
13ASL18	0.097	187	0.001501	0.70491	0.000010	0.704910	1.71	4.82	0.213979	0.512947	0.000006	0.512632	5.53
13ASL20	0.103	164	0.001813	0.70486	0.000011	0.704850	1.52	4.32	0.212870	0.512935	0.000005	0.512622	5.33
13ASL22	0.103	181	0.001648	0.70480	0.000008	0.704795	1.58	4.47	0.213424	0.512943	0.000005	0.512629	5.47

Table 3 Th-U-Pb and Lu-Hf isotopic analytical results for the Ashele basalt

Sample	Th (ppm)	U (ppm)	Pb (ppm)	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	$^{206}Pb/^{204}Pbt$	$^{207}Pb/^{204}Pbt$	$^{208}Pb/^{204}Pbt$
13ASL02	0.183	0.433	2.676	18.196	15.523	37.965	18.046	15.515	37.850
13ASL04	0.203	0.434	1.875	18.305	15.529	38.033	18.068	15.517	37.868
13ASL05	0.196	0.447	2.448	18.150	15.519	37.923	17.975	15.511	37.793
13ASL06	0.279	0.440	2.658	18.150	15.521	37.918	17.919	15.509	37.800
13ASL08	0.289	0.433	2.561	18.141	15.519	37.913	17.894	15.507	37.792
13ASL09	0.221	0.445	1.979	18.271	15.528	38.024	18.026	15.515	37.864
13ASL10	0.182	0.407	2.162	18.230	15.532	38.004	18.045	15.522	37.870
13ASL11	0.214	0.445	3.167	18.253	15.527	38.009	18.105	15.519	37.908
13ASL13	0.222	0.442	3.380	18.173	15.523	37.942	18.029	15.516	37.849
13ASL14	0.203	0.433	2.780	18.124	15.520	37.907	17.964	15.512	37.796
13ASL15	0.230	0.428	9.639	18.177	15.520	37.942	18.125	15.518	37.910
13ASL1	0.272	0.504	2.043	18.224	15.524	37.984	17.932	15.509	37.808

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13ASL18	0.212	0.441	2.628	18.150	15.520	37.925	17.973	15.511	37.805
13ASL20	0.191	0.445	2.658	18.194	15.522	37.958	18.037	15.514	37.839
13ASL22	0.199	0.443	2.336	18.201	15.521	37.963	18.014	15.512	37.828

Table 3 continued

Sample	Lu (ppm)	Hf (ppm)	$^{176}Lf/^{177}Hf$	$^{176}Hf/^{177}Hf$	2σ	$^{176}Hf/^{177}Hft$	$\varepsilon Hf(t)$
13ASL02	0.289	0.695	0.057575	0.283305	0.000006	0.283063	14.8
13ASL04	0.273	0.702	0.055242	0.283310	0.000005	0.283078	15.4
13ASL05	0.293	0.728	0.057172	0.283311	0.000004	0.283070	15.1
13ASL06	0.280	0.712	0.055863	0.283314	0.000004	0.283079	15.4
13ASL08	0.286	0.704	0.057708	0.283306	0.000004	0.283064	14.9
13ASL09	0.290	0.736	0.055971	0.283287	0.000004	0.283051	14.4
13ASL10	0.277	0.673	0.058467	0.283344	0.000007	0.283098	16.1
13ASL11	0.288	0.717	0.057058	0.283301	0.000005	0.283061	14.8
13ASL13	0.290	0.732	0.056277	0.282896	0.000010	0.283060	14.7
13ASL14	0.280	0.708	0.056178	0.282891	0.000004	0.283055	14.5
13ASL15	0.286	0.670	0.060637	0.283317	0.000004	0.283062	14.8
13ASL16	0.324	0.808	0.056961	0.283315	0.000005	0.283076	15.3
13ASL	0.287	0.715	0.0570	0.2832	0.0000	0.2830	14.7

18			19	98	04	58	
13ASL 20	0.287	0.714	0.0570 99	0.2833 28	0.0000 07	0.2830 88	15.7
13ASL 22	0.296	0.729	0.0576 78	0.2833 50	0.0000 08	0.2831 08	16.4

Highlights

● Triassic basalts in Ashele have chondrite-normalized REE patterns depleted in LREE and are akin to MORB-type basalts.

● Sr-Nd-Pb-Hf isotopic compositions of the basalts are also indicative of a depleted mantle source.

● A Triassic depleted lithosphere mantle, not ancient continental lithosphere, existed underneath the Paleozoic orogen.

● Consecutive partial melting of the depleted mantle may have changed its composition toward continental lithosphere.

Graphical abstract

Post-collision

Ashele
Basalt (225Ma)

Continental
Crust of Chinese Altai

Erqis
Fault

Continental
Mantle
Transformed
from
Juvenile
Sub-arc
Mantle

Mantle Melting