**Carbonatites of Tarim (NW China): First evidence of crustal contribution in
carbonatites from a large igneous province**

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ABSTRACT

Many carbonatites are associated both spatially and temporally with large igneous provinces (LIPs), and considered to originate from a mantle plume source lacking any contribution from recycled crustal materials. Here, we report an occurrence of carbonatite enriched in rare-earth elements (REE) and associated with the Tarim LIP in northwestern China. The Tarim LIP comprises intrusive and volcanic products of mantle plume activity spanning from ~300 to 280 Ma. The carbonatites at Wajilitage in the northwestern part of Tarim are dominated by calcite and dolomite varieties, and contain abundant REE minerals (principally, monazite and REE-fluorcarbonates). Th-Pb age determination of monazite yielded an emplacement age of 266 ± 5.3 Ma, i.e. appreciably younger than the eruption age of flood basalts at ~290 Ma. The carbonatites show low initial $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7037-0.7041) and high $\epsilon_{\text{Nd}(t)}$ (1.2-4) values, which depart from the isotopic characteristics of plume-derived basalts and high-Mg picrites from the same area. This indicates that the Wajilitage carbonatites derived from a mantle source isotopically distinct from the one responsible for the voluminous (ultra)mafic volcanism at Tarim. The carbonatites show $\delta^{26}\text{Mg}_{\text{DSM3}}$ values (−0.99 to −0.65 ‰) that are significantly lower than those in typical mantle-derived rocks and rift carbonatites, but close to marine sediments and orogenic carbonatites. We propose that the carbonatites in the Tarim LIP formed by decompressional melting of recycled sediments mixed with the ambient mantle peridotite. The enriched components in the Tarim plume could be accounted for by the presence of recycled sedimentary components in the subcontinental mantle.

Keywords: Carbonatite; large igneous province; plume; subducted sediments; carbonate recycling

1. Introduction

Carbonatites are mantle-derived carbonate-rich igneous rocks commonly occurring in intracontinental rift settings and associated with alkaline silicate rocks. Carbonatites provide valuable information on the composition of their mantle source because: (1) their Sr-Nd-Hf isotopic characteristics are generally inherited from, and representative of, the source owing to the initially high concentrations of Sr, rare-earth and high-field-strength elements (REE and HFSE, respectively) in their parental magmas (e.g., Eby, 1975; Nelson et al., 1988; Chakhmouradian et al., 2016), and (2) these magmas have extremely low viscosities (Treiman, 1989), which facilitate their rapid ascent to the surface. Carbonate melts also play a significant role in mantle metasomatism and as a result, are intimately linked to the secular evolution of the subcontinental mantle (Rudnick et al., 1993). There is increasing evidence that many carbonatites around the world are linked to mantle plume activity (Bell and Tilton, 2001; Ernst and Bell, 2010). Evidence for the involvement of plumes in carbonatite petrogenesis includes the spatial and temporal association between these rocks and large igneous provinces (LIPs), and similarities in Sr, Nd, Hf and Pb isotopic characteristics between young carbonatites and ocean island basalts (OIBs) (Nelson et al., 1988; Bell and Tilton, 2001; Ernst and Bell, 2010). Examples, summarized by Bell (2001), Bell and Simonetti (2010) and Ernst and Bell (2010), include the Paraná-Etendeka (South America and southwestern Africa), Keweenawan (Canada),

Maimecha-Kotuy portion of the Siberian Trap (Russia) and Deccan (India) LIPs. It was proposed that carbonatitic magmas in these regions were formed by low-degree partial melting of volatile-rich and colder outer edges of the plume, and without any contribution from recycled crustal materials (Bell and Simonetti, 2010). The LIPs and carbonatites have been considered to evolve along different pathways, representing a single magma-generation process (Ernst and Bell, 2010). However, Fischer et al. (2009) found that $^3\text{He}/^4\text{He}$ ratios of carbonatite lavas in Oldoinyo Lengai, Tanzania, show considerable overlap with those for samples of mid-ocean ridge basalts (MORB) and lithospheric mantle. Here, we report newly discovered carbonatites associated with the Tarim LIP of northwestern China. The major objective of the present work is to use geochronological and isotopic data for constraining the relationship (if any) between the carbonatites and other igneous rocks in the Province.

2. Geological setting

The large volume of Early Permian volcanic rocks in the Tarim Block of northwestern China has been recognized as a LIP (Tian et al., 2010; Zhang et al., 2013; Xu et al., 2014). The Tarim Block is surrounded by the Tianshan Mountains to the north and west, and western Kunlun and Altyn orogenic belts to the south (Fig. 1a). It was amalgamated with the southern part of the Central Asian Orogenic Belt during the Late Paleozoic and is composed of a Precambrian crystalline basement, including volcano-sedimentary and high-grade metamorphic rocks, and a thick Phanerozoic sedimentary cover which comprises Ordovician, Permian and Cretaceous strata (Long et al., 2010). The Block is

also the largest basin in China, occupying an area of 600,000 km². The LIP consists of voluminous volcanic successions and associated intrusions, including flood basalts, dike swarms, Fe-Ti oxide-bearing mafic-ultramafic intrusions, picrites, porphyritic pyroxenite-dunite, lamprophyres, nephelinites, syenites, A-type granites and rhyolites (Xu et al., 2014). It covers an area of more than 300,000 km² with a thickness of igneous package ranging from several hundred meters to 3 km, according to recent geophysical surveys and oil exploration deep-drill logs (Tian et al., 2010; Xu et al., 2014). Drill-hole data indicate that the thickness of volcanic rocks ranges from 200 to 800 m, and the entire volume of the Tarim basalts is estimated to be 150,000 km³ (Xu et al., 2014). Based on the currently available geochronological data, the Tarim magmatism can be divided into several major phases (Xu et al., 2014): 1) ~300 Ma event, which produced small-volume porphyritic pyroxenite-dunite and is interpreted to mark the onset of plume-induced magmatism (Zhang et al., 2013); 2) ~290 Ma outpourings of flood-basalts, which produced the bulk of igneous rocks at Tarim; (3) rhyolitic activity at 293–287 Ma, whose extent is presently unknown; and 4) emplacement of small-volume mafic-ultramafic dikes, quartz syenites and A-type granites at ~280 Ma contemporaneously with another episode of rhyolitic volcanism (282-277 Ma).

The Wajilitage region is located in the northwestern part of the Tarim LIP. A variety of igneous rocks, including porphyritic pyroxenite-dunite, mafic-ultramafic intrusions, diabase dikes, alkaline lamprophyres, syenites, nephelinites and carbonatites, have been documented in this area. The carbonatites occur as dikes (Fig. 1b), and form a NWW-SEE trend within an area of 5 km². They are not typical alkaline-rock carbonatite complexes

(e.g., Mitchell, 2005). Similar examples include Gifford Greek from Australia (Pirajno et al., 2010), Haast River in New Zealand (Cooper et al., 2015). The Wajilitage carbonatites intruded pyroxenite and gabbro, as well as older (Devonian) clastic sequences. Individual dikes range from several tens of meters to > 1 km in length and from 1 to 15 m in thickness. They are composed predominantly of fine- to medium-grained calcite and dolomite (> 85 vol.%) and can be roughly grouped into calcite and dolomite-calcite carbonatites. The calcite carbonatites are mainly composed of calcite. Minor and accessory phases include augite, aegirine, potassium feldspar, albite, biotite, dolomite, barite, celestine, pyrite, magnetite, fluorapatite and REE minerals (some of the listed silicate minerals are probably xenocrysts derived from the wall-rocks). The REE minerals include monazite-(Ce), bastnäsite-(Ce), parisite-(Ce) and synchysite-(Ce). Monazite-(Ce) usually occurs as isolated crystals and their aggregates, or as intergrowths with barite and celestine. The dolomite-calcite carbonatites are mainly composed of dolomite and subordinate calcite, with minor and accessory minerals of augite, aegirine, albite, barite, celestine, pyrite, fluorapatite and REE minerals. Monazite-(Ce) and/or parisite-(Ce) form intergrowths with small aggregates of fine-grained dolomite. Based on drill-hole and surface sampling, Geological Team No.11 of the Xinjiang Geological Bureau reported an average composition of 0.54 to 1.92 wt.% total REE₂O₃ for the carbonatite dykes. The sampling locations in this study are shown in Fig. 1b.

3. Analytical Methods

The 10 carbonatite samples were used for major, trace element and isotopic

measurements, and 6 samples for age determination. The rocks were manually crushed and grinded into powders using an agate mortar which was cleaned by distilled water every time to avoid contamination. The monazite grains were separated from whole rocks using standard crushing and separation techniques and, then handpicked under a binocular microscope. They were mounted into a 25 mm epoxy resin and polished to expose the cores of the grains in readiness for geochronology and Nd isotope analysis.

3.1. Monazite U-Th-Pb dating

U-Th-Pb age determination of monazite was performed by secondary-ion mass-spectrometry (SIMS) using a Cameca IMS-1280 instrument at the Institute of Geology and Geophysics, Chinese Academy of Sciences (CAS). An O^{2-} primary ion beam was accelerated at 13 kV with an intensity of ca. 2-3 nA. The aperture illumination mode (Kohler illumination) was used with a 200 μm primary beam mass filter aperture to produce even sputtering over the entire analyzed area representing an elliptical spot $\sim 20 \times 30 \mu m$ in size. Positive secondary ions were extracted with a 10 kV potential. Sample charging effects were minimized by optimizing the energy offset to maximum transmission in a 60 eV energy window at the start of each analysis, using Th^+ as a reference peak. A mass resolution of ca. 5,500 (defined at 1% peak height) was used, which was enough to separate most of the measured U, Th and Pb isotopes from their isobaric interferences (such as REE oxides). Monazite 44069 was used as a standard. Detailed analytical procedures have been described by Li et al. (2013).

3.2. Major- and trace-element analysis

Major element compositions of carbonate minerals were measured on C-coated polished section using a JXA-8230 electron microprobe equipped with wavelength-dispersive X-ray spectrometers at the Key Laboratory of Metallogeny and Mineral Assessment, Institute of Mineral Resources, Chinese Academy of Geological Sciences. The microprobe was operated at an accelerating voltage of 15 kV and a beam current of 20 nA with an electron beam diameter of 2-5 μm . All raw data were corrected with standard ZAF correction procedures.

Major elements in pulverized whole-rock carbonatite samples were analyzed by X-ray fluorescence spectrometry at the Institute of Geochemistry, CAS, with analytical uncertainties varying from 1% to 3%. Loss on ignition was measured using 1.0 g of powdered sample heated to 1100 $^{\circ}\text{C}$ for 1 hour. Trace-element abundances in the whole-rock samples were determined in solution by quadrupole inductively-coupled-plasma mass-spectrometry (ICP-MS) at the Institute of Geochemistry, CAS. Prior to the analysis, 50 mg of rock powder were dissolved in a Teflon reaction vessel using 1 ml of HF (38 wt. %) and 0.5 ml of HNO_3 (68 wt. %). The sealed reaction vessel was placed in an electric oven and heated to 180 $^{\circ}\text{C}$ for 24 hours. The cooling solution was spiked with 1 ml of 1 μgml^{-1} Rh as an internal standard and evaporated on a hot plate. This was followed by two cycles of dilution with 1 ml of HNO_3 and evaporation to dryness. The final residue was re-dissolved in 8 ml of HNO_3 . The reaction vessel was re-sealed, returned to the oven and heated to 135 $^{\circ}\text{C}$ for 5 h. The final solution was diluted to 100 ml with distilled de-ionized water. The analytical error of the ICP-MS analysis was

177 better than 10% for most elements.

179 **3.3. Trace-element analysis of carbonate minerals**

180 In-situ laser-ablation (LA)-ICP-MS analyses of the carbonate minerals were performed
181 at Peking University using a COMPexPro 102 excimer laser and an Agilent7500ce/cs
182 mass-spectrometer. The diameter of the ablation spot was 32 μm . The NIST 610 glass
183 was used as a calibration standard, and the Ca content measured by electron-microprobe
184 analysis as an internal standard for all samples. In-run signal intensity for indicative trace
185 elements was monitored during the analysis to ensure that the ablation spot was confined
186 to the area of interest and did not sample other mineral phases or inclusions. The
187 analytical error of the LA-ICP-MS measurements was better than 5% at the ppm level.

189 **3.4. Multi-isotope analysis**

190 The carbon and oxygen isotopic compositions of the carbonatites were measured at
191 the Institute of Geochemistry, CAS, using an IsoPrime continuous-flow isotope-ratio
192 mass-spectrometer. The results are expressed using the conventional delta notation ($\delta^{13}\text{C}$
193 and $\delta^{18}\text{O}$), as per mille (‰) variations relative to the Vienna PDB and SMOW standards,
194 respectively. Their estimated error is ± 0.1 ‰ for both.

195 Prior to Mg isotopic measurements, the samples were pretreated following the
196 procedure described by Shen et al. (2013). The total procedural blank for Mg was < 15 ng.
197 The abundances of Mg isotopes were quantified using a Neptune MC-ICP-MS at the
198 China University of Geosciences (Beijing). The analytical results are expressed using the

conventional δ -notation in per mille relative to the reference material (DSM3), i.e.
 $\delta^X\text{Mg} = [({}^X\text{Mg}/{}^{24}\text{Mg})_{\text{sample}} / ({}^X\text{Mg}/{}^{24}\text{Mg})_{\text{DSM3}} - 1] \times 1,000$, where X refers to mass 25 or 26, and
DSM3 is a reference solution prepared from pure Mg metal. The analysis of several
external standards (BCR-2, BHVO-2, Kilbourne Hole olivine and Hawaii seawater) gave
results within 2σ error of the recommended values.

The Sr and Nd isotopic compositions of selected whole-rock samples were
determined using multicollector (MC) ICP-MS with a VG Axiom instrument at Peking
University. Mass fractionation corrections for Sr and Nd isotopic ratios were normalized to
 ${}^{86}\text{Sr}/{}^{88}\text{Sr} = 0.1194$ and ${}^{146}\text{Nd}/{}^{144}\text{Nd} = 0.7219$, respectively. Repeated measurements of the
Sr standard NBS987 and Nd standard JNdi yielded ${}^{87}\text{Sr}/{}^{86}\text{Sr} = 0.710252 \pm 10$ (2σ) and
 ${}^{143}\text{Nd}/{}^{144}\text{Nd} = 0.512122 \pm 10$ (2σ), respectively. The Rb, Sr, Sm and Nd concentrations in
the carbonatite samples were analyzed by solution ICP-MS (Thermo Fisher Scientific
X-Series II). Raw data processing included linear drift correction, internal (matrix)
correction, REE and Ba interference corrections, blank subtraction, calibration with
international standards, and a dilution correction. Replicate analyses (including
well-characterized standards) indicated that the error of trace-element measurements was
better than 10%.

In-situ Nd isotopic analysis of monazite was done by MC-ICP-MS using a
Thermo-Finnigan Neptune instrument coupled to a 193 nm ArF Excimer laser-ablation
system at the Institute of Geology and Geophysics, CAS. The laser was focused at 24 μm
above the sample surface and fired using a 6 Hz repetition rate and an energy density of
10 J/cm^2 . An in-house apatite standard (AP1) and Jefferson monazite 44069 were used

for external calibration. Normalized $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ isotopic ratios were calculated using the exponential law. The $^{147}\text{Sm}/^{144}\text{Nd}$ ratio was then externally calibrated against the $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of standard reference materials during the analytical sessions. Detailed analytical procedures have been described by Yang et al. (2008).

4. Results

4.1. Monazite geochronology

We performed high-precision U-Th-Pb geochronology of primary monazite grains separated from the calcite carbonatites. The U concentration in this mineral is too low (Th/U ratios range from 89 to > 10,000) to derive a reliable U-Pb age, but ^{232}Th is consistently well above the limit of detection. The 25 analyzed points show a fairly uniform distribution of ^{232}Th - ^{208}Pb ages regardless of the measured Th content, which indicates no Pb mobility due to post-emplacement deformation or metamorphic overprint. The analyzed grains gave a weighted mean Th-Pb age of 266.0 ± 5.3 Ma (Table S1 and Fig. 2). Thus, the calcite carbonatite is similar in age to the associated alkaline lamprophyre (271 ± 2 Ma; Wang, 2014), but both are appreciably younger than ultramafic and gabbroic rocks from the same area (284.2 ± 2.8 to 281.3 ± 2.2 Ma; Shangguan et al., 2016; Zhang et al., 2016). These data place the carbonatites in the second (Guadalupian) pulse of magmatic activity, postdating the main extrusive events and the emplacement of ultramafic-mafic intrusions at Wajilitage, Puchang and elsewhere in the Tarim Block (Fig. 3; Shangguan et al., 2016).

4.2. Carbonate geochemistry

The carbonate minerals in the carbonatites are represented by calcite and dolomite (Table S2). Dolomite contains variable levels of FeO (3.8-6.9 wt.%), whereas minor to moderate MgO (0.1-2.3 wt.%) and FeO (0.1-0.8 wt. %) were detected in the calcite. The carbonates were also analyzed by LA-ICPMS and determined to have high levels of Sr and Ba, variable REE, as well as very low Rb and HFSE contents (Fig. 4a). The measured Sr concentrations in the carbonates are invariably high in both minerals (> 3000 ppm and up to 1.2 wt. %, respectively), and consistent with a carbonatitic origin (Eby, 1975; Hornig-Kjarsgaard, 1998; Chakhmouradian et al., 2016). The dolomite shows a higher total REE content and stronger enrichment in light lanthanides (LREE) relative to heavy (La/Yb_{CN} = 36-121) (Fig. 4b), and is similar to typical primary carbonates (Chakhmouradian et al., 2016), whereas the calcite is characterized by lower LREE and higher HREE levels, and nearly flat normalized REE patterns (La/Yb_{CN} = 1-4).

4.3. Whole-rock geochemistry

The whole-rock major and trace element compositions of the Wajilitage carbonatites are given in Table S3. As can be expected from their modal composition, the rocks show generally low SiO₂ (≤ 2.0 wt. %) and alkalis ($K_2O + Na_2O \leq 0.5$ wt.%), and variable FeO_t (2.0-4.8 wt.%) and MgO (0.2-14.5 wt.%) depending on the predominant carbonate phase. Based on the CaO, FeO_t + MnO and MgO proportions and using the classification of Woolley and Kempe (1989), the analyzed rocks range from calciocarbonatite to

magnesiocarbonatite.

Both rock types are enriched in most of the incompatible elements, with the exception of Rb, Th, Zr and Hf, relative to the primitive mantle. Their normalized spider diagrams show negative Ta, Pb, Zr and Hf anomalies relative to the overall high REE levels (Fig. 4a). The calciocarbonatite has relatively higher trace-element contents than the magnesiocarbonatite. Chondrite-normalized REE patterns of the carbonatites exhibit consistent LREE enrichment with La/Yb_{CN} ratios ranging 47-75 and 32-83 for the calciocarbonatite and magnesiocarbonatite, respectively (Fig. 4b).

4.4. Isotope geochemistry

Carbon, O, Sr and Nd isotopic data are summarized in Tables S4, S5. The C and O isotopic compositions of five calcite separates both vary within a fairly limited range, from -3.8 to -5.8 ‰ $\delta^{13}\text{C}_{\text{V-PDB}}$ and from 6.4 to 9.5 ‰ $\delta^{18}\text{O}_{\text{V-SMOW}}$. These values are typical for primary, mantle-derived carbonates (Fig. 5), i.e. appear unaffected by surficial secondary processes.

Magnesium isotopic compositions of dolomite and calcite separated from the carbonatites are reported in Table 1 and shown in Fig. 6. All of the investigated samples have remarkably similar $\delta^{26}\text{Mg}_{\text{DSM3}}$ values (-0.99 to -0.68 ‰ in calcite and -0.99 to -0.82 ‰ in dolomite) that are significantly lower than those in mantle-derived xenoliths and magmas, including MORB and OIB (-0.25 ± 0.07 ‰, Teng et al., 2010). Notably, rift-related Oldoinyo Lengai carbonatites, Tanzania, are also appreciably richer in heavy Mg (0.13 - 0.37 ‰; Li et al., 2016) than the Wajilitage rocks, whereas the recently reported

orogenic carbonatites in the Qinling Mts show similarly low $\delta^{26}\text{Mg}_{\text{DSM3}}$ values (–1 to –2 ‰; Song et al., 2016).

Both calcio- and magnesiocarbonatites have similar Sr and Nd isotopic compositions, with low levels of radiogenic Sr [$(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.7037\text{--}0.7041$] and positive $\epsilon_{\text{Nd}(t)}$ values (1.3–4.0). Their Nd isotopic model ages (T_{DM}) range from 0.83 to 1.32 Ga. In-situ Nd isotopic data for the monazite also show little variation in $\epsilon_{\text{Nd}(t)}$ (2.0–3.6) and overlap with the whole-rock range of values, supporting our interpretation that this mineral crystallized from carbonatitic magma, rather than being xenocrysts derived from a crustal source. The Sr–Nd data overlap the range of values reported for the spatially associated nephelinites (Cheng et al., 2015) and alkaline lamprophyre (Wang, 2014). In comparison with flood basalts from the Tarim LIP, the carbonatites show much lower $(^{87}\text{Sr}/^{86}\text{Sr})_i$ and higher $\epsilon_{\text{Nd}(t)}$ values (Fig. 7).

5. Discussion

5.1. Evidence for recycled sediments in the carbonatite source

Carbon isotopic compositions are commonly used to identify the possible presence of subducted sediments in the mantle (Ray, 1999). However, they are effective in identifying recycled organic carbon, but not sensitive to inorganic carbon which dominates the subducted materials (Wang et al., 2014). Indeed, calculated trends of C isotope fractionation in subducted carbonate rocks intersect “Taylor’s box” of typical igneous carbonate compositions at 60–90% decarbonation (Chakhmouradian et al., 2008). In

recent years, Mg isotopes have been used increasingly to identify a recycled sediment signature because they undergo little fractionation due to metamorphic dehydration and magma extraction during subduction (Teng et al., 2010; Wang et al., 2014). Our results show that the Tarim carbonatites are characterized by significantly lower $\delta^{26}\text{Mg}_{\text{DSM3}}$ ratios than typical primary mantle-derived rocks, including MORB, OIB and peridotite xenoliths. Such light-Mg isotopic composition cannot be derived from an unmodified primary mantle source because there is insignificant Mg isotope fractionation between olivine and pyroxenes, olivine and carbonates, or during partial melting of the mantle, and fractional crystallization of ferromagnesian silicates (Teng et al., 2010; Macris et al., 2013; Hu et al., 2016; Li et al., 2016). Based on crystal-chemical considerations, such as the coordination of Mg in mineral structures (e.g., Huang et al., 2013), primary carbonatitic magmas formed by partial melting of carbonated peridotite or eclogite should have $\delta^{26}\text{Mg}_{\text{DSM3}}$ values similar to, or if garnet (which is relatively enriched in light Mg isotopes more than pyroxene and olivine; Huang et al., 2013) is present higher than, that of the mantle source.

The following processes may affect the Mg isotope composition of carbonatites: silicate-carbonate liquid immiscibility, crystal fractionation, crustal contamination, and fluid-driven alteration. The magnitude of isotope fractionation between carbonate-silicate immiscible process is relatively too small (-0.3 to 0.09 ‰; Li et al., 2016) to produce large differences. The light Mg isotope preferentially partitions into fractionating carbonate minerals relative to the liquid (Li et al., 2016), resulting in evolved carbonatitic magmas enriched in heavy Mg isotopes. Crustal contamination of mantle-derived magmas could cause either relative enrichment or depletion of the latter in heavy isotopes (Li et al., 2010;

Yang et al., 2016; Fig. 6). Carbonate sediments could serve as a source of light-Mg enrichment (Galy et al., 2002; Huang et al., 2015), but Ordovician carbonate rocks of the Tarim area (Jiang et al., 2002) show higher $\delta^{13}\text{C}_{\text{V-PDB}}$ (-1.5 to 2.0 ‰) and $(^{87}\text{Sr}/^{86}\text{Sr})_i$ values (0.706-0.710) than the studied carbonatites. A negative Pb anomaly in normalized trace-element patterns (Fig. 4a) and near-chondritic Y/Ho ratios in the carbonatites (29.7 ± 2.3) also rule out any significant crustal contamination. Secondary fluid (e.g., seawater, meteoric water) interaction can in principle cause isotopic re-equilibration of the primary magmatic minerals (Demény et al., 1998), which is not supported by the available stable-isotopic data (Fig. 5).

We will now explore the possibility that a recycled crustal carbonate component in the mantle source of the Tarim carbonatites is the source of their light Mg isotopic signature. Indeed, there are several lines of evidence supporting this interpretation. Firstly, the observed Mg isotopic budget is close to those of marine sedimentary carbonates over geological time scales (-2.8 to -1.0 ‰, Blätter et al., 2015; Huang et al., 2015), and subducted carbonates are expected to retain their initial Mg isotopic signature (Wang et al., 2014). Deeply recycled carbonates and carbonated eclogites are characterized by light Mg isotopic compositions ranging from -2.51 to -0.53 (Wang et al., 2014). Indisputably orogenic carbonatites in the Qinling orogen (Miaoya, Huanglongpu and Huayangchuan) were generated by melting of subducted oceanic crust during the collision between the North and South China Blocks (Xu et al., 2014) and show a clear “marine carbonate” Mg signature ($\delta^{26}\text{Mg}_{\text{DSM3}} = -1 - -2$ ‰; Song et al., 2016).

5.2. Carbonatite origin and mantle plume magmatism in Tarim

It is well established that the Tarim LIP is the product of early Permian mantle plume activities (Tian et al., 2010; Zhang et al., 2013; Xu et al., 2014). According to published geochronological data, the flood basalts formed within a relatively short time period between 292 and 287 Ma. The basalts are associated with high-Mg picrites (MgO = 15-20 wt%: Tian et al., 2010), that probably represent the least contaminated components of the plume source (Campbell, 2002). Both the picrites and flood basalts are enriched in incompatible elements and have Sr-Nd isotopic characteristics indicative of an enriched mantle reservoir (Fig. 7). Intrusions of ultramafic and mafic rocks in the Wajilitage area and elsewhere span from 284 to 273 Ma in age, indicating a long-lived melt source (Zou et al. 2015). According to our data and published work (Cheng et al., 2015; Shangguan et al., 2016), the Tarim carbonatites and other undersaturated alkaline rocks are perceptibly younger (by ~10-20 Ma) than this main stage of plume activity. Campbell and Griffiths (1990) proposed that plume magmatism lasts a relatively short time, with the bulk of magmas emplaced within 2-3 Ma and then followed by smaller-volume mantle melting products over a further 5-10 Ma. Plume-derived carbonatites elsewhere, reported in the previous literature (Bell and Tilton, 2001; Ernst and Bell, 2010) were generated synchronously with the main stage of flood basalt magmatism. It is also significant that the Tarim carbonatites were derived from a relatively depleted mantle source with lower ($^{87}\text{Sr}/^{86}\text{Sr}$)_i and higher $\epsilon_{\text{Nd}(t)}$ values than the earlier-emplaced basalts and picrites (Fig. 7).

Mantle plumes represent buoyant material that separates from the thermal boundary layer at the core-mantle boundary (Campbell and Griffiths, 1990; Hastie et al., 2016). The

available radiogenic and noble gas isotopic evidence indicates that the plume sources contain recycled crustal material derived from subducted slabs (Hofmann, 1997; White, 2010). An alternative view was expressed by Niu and O'Hara (2003), and Niu et al. (2012), who argued that the subducted slab could not be the major source of plume material (and products of plume activity, such as OIB) on the basis of trace-element constraints. The majority of plume-derived magmas from Tarim show consistently enriched isotope characteristics (Fig. 7), implying either a recycled crustal contribution to their plume source, or contamination of the upwelling material by enriched lithospheric components.

Evidence from fresh ultramafic xenoliths hosted by Cenozoic alkali basalts in the western Tarim Block shows that the underlying lithospheric mantle is essentially depleted, with $\epsilon_{\text{Nd(t)}}$ ranging from 9.5 to 14.3 (Chen et al., 2014). Primitive magmas, exemplified by clinopyroxene macrocrysts from the Xiaohaizi mafic dikes just north of Wajilitage, exhibit similar characteristics: $\epsilon_{\text{Nd(t)}} = 4.5\text{--}4.8$, $(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.7035\text{--}0.7037$ (Wei et al., 2015). The flood basalts, picrites and carbonatites show a negligible negative Pb anomaly in their normalized distribution diagrams and lack any correlation between their SiO_2 content and Nd-Pb isotopic signature (Tian et al., 2010; Xu et al., 2014; this work), implying little to nil crustal contamination. This conclusion is consistent with the C-O isotopic data presented in this work (Fig. 5) and unmodified, primitive Y/Ho ratios of the rock-forming carbonates (Table S2). Hence, we interpret the trace-element and isotopic characteristics of the Wajilitage carbonatites as representative of their mantle source. The Mg isotopic signature of the examined rocks indicates the presence of recycled sedimentary material in their source. Evidence for a Proterozoic subduction event below the Tarim Block has

been presented by Zheng et al. (2010). Blueschist-facies metamorphic rocks, conventionally recognized as an indicator of subduction, have been described from the northwestern part of Tarim and interpreted as part of a Proterozoic ocean basin separating the Tarim plate from Australia (Zheng et al., 2010). The youngest Nd model ages of the carbonatites ($T_{\text{CHUR}} = 0.83\text{-}1.32$) are similar to the Sm-Nd isochron age of the blueschist (890 Ma; Zheng et al., 2010), and arc magmatism in northern Tarim (see below), potentially linking the carbonatites to the Proterozoic subduction.

Experimental work suggests that carbonate sedimentary material can be recycled deep into the mantle during subduction, and undergo melting at various depths, potentially down to the transition zone, to produce carbonate melts (Litasov and Ohtani, 2009; Dasgupta and Hirschmann, 2010; Thomson et al., 2016). However, the Wajilitage carbonatites are characterized by depleted Sr and Nd isotopic compositions, clearly distinct from the low- ϵ_{Nd} , high- $^{87}\text{Sr}/^{86}\text{Sr}$ signature of subducted sediments (Plank and Langmuir, 1997). This apparent discrepancy is readily explained if melting took place in the magnesite stability field because magnesite has lower Sr and Nd contents than calcite. Such carbonate melts would inherit the Sr-Nd isotope feature from the depleted mantle they are interacting with, but offset its $\delta^{26}\text{Mg}_{\text{DSM3}}$ to lower values (e.g., Huang and Xiao, 2016). Upwelling hot material from deep mantle would also significantly increase the ambient mantle temperature, resulting in an isotopically heterogeneous reservoir with depleted Sr-Nd isotopic characteristics and low $\delta^{26}\text{Mg}_{\text{DSM3}}$ values. Homogenization explains the uniformity of isotopic composition of the Wajilitage carbonatites regardless of the large variations in their modal composition. That the carbonatitic magmas retained the

“sedimentary” Mg isotopic signature is consistent with little fractionation between Mg isotopes during melting (Macris et al., 2013).

In recent years, published evidence from across the Tarim LIP has revealed the importance of subduction processes in the sub-Tarim mantle. This includes: high water content in the magmas parental to the flood basalts (Xia et al., 2016); the Nd-Hf-Pb isotope systematics of some basalt flows from NW Tarim (Xu et al., 2014; Wei et al., 2015); a record of Neoproterozoic (760-820 Ma) and Silurian (~420 Ma) arc magmatism along the northern margin of the Block (Tang et al., 2016), different to the 820-850 Ma LIP magmatism in the Qaidam block, southern Tarim (Xu et al., 2016). Given this almost 600-Ma history of intermittent subduction along the western edge of the Tarim Block, it remains to be determined which of these events was the most likely source of C for the Wajilitage carbonatites. However, their Mg isotope signature indicative of deep carbon cycling (see above), and T_{DM} values (> 0.8 Ga) are not consistent with a young (i.e., Phanerozoic) recycled-crust contribution. The temporal and spatial scales of the Neoproterozoic Circum-Rodinia subduction system (see discussion in Tang et al., 2016) suggest that it could supply recycled C on the global scale. Further combined Nd-Mg isotopic studies of anorogenic carbonatites elsewhere in the world are desirable.

6. Conclusions

Although in the Tarim Block, calcite and dolomite carbonatites are intimately associated with flood basalts and other common products of mantle-plume activity, these carbonatite rocks were emplaced at 266 ± 5 Ma, i.e. significantly later than the main phase

of basaltic volcanism at 292-287 Ma. The carbonatites differ from the basalts in having a lower ($^{87}\text{Sr}/^{86}\text{Sr}$)_i, but higher $\epsilon_{\text{Nd}(t)}$ values, indicating that their parental magmas did not derive from the same mantle source. The carbonatites exhibit Mg isotopic compositions indicative of a marine sedimentary contribution. We propose that the studied carbonatites derived from melts produced by decompressional melting of upwelling, deep-seated sediments and their mixing with the depleted mantle. Recycled crustal materials may have a significant effect on the Sr-Nd isotopic signature of plume-derived magmas.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at

References

- Bell, K. 2001. Carbonatites: relationships to mantle-plume activity. Special papers, Geological society of America, 267-290.
- Bell, K., Simonetti, A., 2010. Source of parental melts to carbonatites-critical isotopic

464 constraints. *Mineralogy and Petrology* 98, 77-89.

465 Bell, K., Tilton, G.R., 2001. Nd, Pb and Sr isotopic compositions of East African
 466 carbonatites: evidence for mantle mixing and plume inhomogeneity. *Journal of*
 467 *Petrology* 42, 1927-1945.

468 Blätter, C.L., Miller, N.R., Higgins, J.A., 2015. Mg and Ca isotope signatures of authigenic
 469 dolomite in siliceous deep-sea sediments. *Earth and Planetary Science Letters* 419,
 470 32-42.

471 Campbell, I.H., Griffiths, R.W., 1990. Implications of mantle plume structure for the
 472 evolution of flood basalts. *Earth and Planetary Science Letters* 99, 79-93.

473 Campbell, I.H., 2002. Implications of Nb/U, Th/U and Sm/Nd in plume magmas for the
 474 relationship between continental and oceanic crust formation and the development of
 475 the depleted mantle. *Geochimica et Cosmochimica Acta* 66, 1651-1661.

476 Chakhmouradian, A.R., Mumin, A.H., Demény, A., Elliott, B., 2008. Postorogenic
 477 carbonatites at Eden Lake, Trans-Hudson Orogen (northern Manitoba, Canada).
 478 *Lithos* 103, 503-526.

479 Chakhmouradian, A. R., Reguir, E. P., Couëslan, C., Yang, P., 2016. Calcite and dolomite
 480 in intrusive carbonatites. II. Trace-element variations. *Mineralogy and Petrology* 110,
 481 361-377.

482 Chakhmouradian, A.R., Reguir, E.P., Kressall, R.D., Crozier, J., Pisiak, L., Sidhu, R.,
 483 Yang, P., 2015. Carbonatite-hosted niobium deposit at Aley, northern British
 484 Columbia (Canada): Mineralogy, geochemistry and petrogenesis. *Ore Geology*
 485 *Reviews* 64, 642–666.

486 Chen, M.-M., Tian, W., Suzuki, K., Tejada, M.L.G., Liu, F.-L., Senda, R., 2014. Peridotite
487 and pyroxenite xenoliths from Tarim, NW China: evidences for melt depletion and
488 mantle refertilization in the mantle source region of the Tarim flood basalt. *Lithos* 204,
489 97-111.

490 Cheng, Z.-G., Zhang, Z.-C., Hou, T., Santosh, M., Zhang, D.-Y., Ke, S., 2015.
491 Petrogenesis of nephelinites from the Tarim large igneous province, NW China:
492 implications for mantle source characteristics and plume-lithosphere interaction.
493 *Lithos* 220, 164-178.

494 Cooper, A.F., Collins, A.K., Palin, J.M., Spratt, J., 2015. Mineralogical evolution and REE
495 mobility during crystallisation of ancylite-bearing ferrocarbonatite, Haast River, New
496 Zealand. *Lithos* 216, 324-337.

497 Dasgupta, R., Hirschmann, M.M., 2010. The deep carbon cycle and melting in Earth's
498 interior. *Earth and Planetary Science Letters* 298, 1-13.

499 Demény, A., Ahijado, A., Casillas, R., Vennemann, T.W., 1998. Crustal contamination and
500 fluid-rock interaction in the carbonatites of Fuerteventura Canary Islands, Spain: a C,
501 O, H isotope study. *Lithos* 44, 101-115.

502 Eby, G.N., 1975. Abundance and distribution of the rare-earth elements and yttrium in the
503 rocks and minerals of the Oka carbonatite complex, Quebec. *Geochimica et*
504 *Cosmochimica Acta* 39, 597-620.

505 Ernst, R.E., Bell, K., 2010. Large igneous provinces (LIPs) and carbonatites. *Mineralogy*
506 *and Petrology* 98, 55-76.

507 Fischer, T.P., Burnard, P., Marty, B., Hilton, D.R., Furi, E., Palhol, F., Sharp, Z.D.,

508 Mangasini, F., 2009. Upper mantle volatile chemistry at Oldoinyo Lengai volcano and
 509 the origin of carbonatites. *Nature* 459, 77-80.

510 Galy, A., Bar-Matthews, M., Halicz, L., O'Nions, R.K., 2002. Mg isotopic composition of
 511 carbonate: insight from speleothem formation. *Earth and Planetary Science Letters*
 512 201, 105-115.

513 Hastie, A. R., Fitton, J. G., Kerr, A. C., McDonald, I., Schwindrofska, A., Hoernle, K., 2016.
 514 The composition of mantle plumes and the deep Earth. *Earth and Planetary Science*
 515 *Letters* 444, 13-25.

516 Hofmann, A.W., 1997. Mantle geochemistry: the message from oceanic
 517 volcanism. *Nature* 385, 219-229.

518 Hornig-Kjarsgaard, I., 1988. Rare earth elements in sovitic carbonatites and their mineral
 519 phases. *Journal of Petrology* 39, 2105-2121.

520 Hu, Y., Teng, F.-Z., Zhang, H.-F., Xiao, Y., Su, B.-X., 2016. Metasomatism-induced mantle
 521 magnesium isotopic heterogeneity: Evidence from pyroxenites. *Geochimica et*
 522 *Cosmochimica Acta* 185, 88-111.

523 Huang, F., Chen, L., Wu, Z., Wang, W., 2013. First-principles calculations of equilibrium
 524 Mg isotope fractionations between garnet, clinopyroxene, orthopyroxene, and olivine:
 525 Implications for Mg isotope thermometry. *Earth and Planetary Science Letters* 367,
 526 61-70.

527 Huang, K.-J., Shen, B., Lang, X.-G., Tang, W.-B., Peng, Y., Ke, S., Kaufman, A.J., Ma,
 528 H.-R., Li, F.-B., 2015. Magnesium isotopic compositions of the Mesoproterozoic
 529 dolostones: Implications for Mg isotopic systematics of marine carbonates.

530 *Geochimica et Cosmochimica Acta* 164, 333-351.

531 Huang, J., Xiao, Y.L., 2016. Mg-Sr isotopes of low 26Mg basalts tracing recycled
532 carbonate species: implication for the initial melting depth of the carbonated mantle in
533 Eastern China. *International Geology Review* 58, 1350-1362.

534 Jiang, M.-S., Zhu, J.-Q., Chen, D.-Z., Qiao, G.-S., 2002. Carbon and strontium isotope
535 characteristic and response with sea level variation of Ordovician carbonate rocks in
536 Tarim basin. *Science in China (D)* 32, 36-42.

537 Keller, J., Hoefs, J., 1995. Stable isotope characteristics of recent natrocarbonatites from
538 Oldoinyo Lengai. Springer, Berlin, 113-123.

539 Li, Q.L., Li, X.H., Lan, Z.W., Guo, C.L., Yang, Y.N., Liu, Y., Tang, G.Q. 2013. Monazite and
540 xenotime U-Th-Pb geochronology by ion microprobe: dating highly fractionated
541 granites at Xihuashan tungsten mine, SE China. *Contributions to Mineralogy and*
542 *Petrology* 166, 65-80.

543 Li, W.-Y., Teng, F.-Z., Halama, R., Keller, J., Klaudius, J., 2016. Magnesium isotope
544 fractionation during carbonatite magmatism at Oldoinyo Lengai, Tanzania. *Earth and*
545 *Planetary Science Letters* 444, 26-33.

546 Li, W.-Y., Teng, F.-Z., Ke, S., Rudnick, R.-L., Gao, S., Wu, F.-Y., Chappell, B.-W., 2010.
547 Heterogeneous magnesium isotopic composition of the upper continental crust.
548 *Geochimica et Cosmochimica Acta* 74, 6867-6884.

549 Litasov, K., Ohtani, E., 2010. The solidus of carbonated eclogite in the system
550 CaO–Al₂O₃–MgO–SiO₂–Na₂O–CO₂ to 32 GPa and carbonatite liquid in the deep
551 mantle. *Earth and Planetary Science Letters* 295, 115-126.

552 Long, X.P., Yuan, C., Sun, M., Zhao, G.C., Xiao, W.J., Wang, Y.J., Yang, Y.H., Hu, A.Q.,
 553 2010. Archean crustal evolution of the northern Tarim craton, NW China: zircon U–Pb
 554 and Hf isotopic constraints. *Precambrian Research* 180, 272-284.

555 Macris, C.A., Young, E.D., Manning, C.E., 2013. Experimental determination of
 556 equilibrium magnesium isotope fractionation between spinel, forsterite, and
 557 magnesite from 600 to 800°C. *Geochimica et Cosmochimica Acta* 118, 18-32.

558 Mitchell, R. H., 2005. Carbonatites and carbonatites and carbonatites. *The Canadian*
 559 *Mineralogist* 43, 2049-2068.

560 Nelson, D.R., Chivas, A.R., Chappell, B.W., McCulloch, M.T., 1998. Geochemical and
 561 isotopic systematic in carbonatites and implications for the evolution of ocean-island
 562 sources. *Geochimica et Cosmochimica Acta* 52, 1-17.

563 Niu, Y, O'Hara, M.J., 2003. Origin of ocean island basalts: A new perspective from
 564 petrology, geochemistry, and mineral physics considerations. *Journal of Geophysical*
 565 *Research: Solid Earth* 108, 283-299.

566 Niu, Y., Wilson, M., Humphreys, E.R., Hara, M.J.O., 2012. A trace element perspective on
 567 the source of ocean island basalts (OIB) and fate of subducted ocean crust (SOC)
 568 and mantle lithosphere (SML). *Episodes* 35, 310-327.

569 Pirajno, F., Sheppard, S., Johnson, S., González-Álvarez, I., Thorne, A., Cutten, H., 2010.
 570 The Gifford Creek Carbonatite Complex and associated REE, U and Fe oxides
 571 mineralisation, Gascoyne Province, Western Australia. *Geology* 25, 477-494.

572 Plank, T., Langmuir, C.H., 1998. The chemical composition of subducting sediment and
 573 its consequences for the crust and mantle. *Chemical Geology* 145, 325-394.

574 Ray, J.S., Ramesh, R., Pande, K., 1999. Carbon isotopes in Kerguelen plume-derived
575 carbonatites: evidence for recycled inorganic carbon. *Earth and Planetary Science*
576 *Letters* 170, 205-214.

577 Rudnick, R.L., McDonough, W.F., Chappell, B.W., 1993. Carbonatite metasomatism in
578 the northern Tanzanian mantle: petrographic and geochemical characteristics. *Earth*
579 *and Planetary Science Letters* 114, 463-476.

580 Shangguan, S., Peate, I.U., Tian, W., Xu, Y., 2016. Re-evaluating the geochronology of
581 the Permian Tarim magmatic province: implications for temporal evolution of
582 magmatism. *Journal of the Geological Society* 173, 228-239.

583 Shen, B., Wimpenny, J., Lee, C.T.A., Tollstrup, D., Yin, Q.Z., 2013. Magnesium isotope
584 systematics of endokarns: Implications for wallrock reaction in magma
585 chambers. *Chemical Geology* 356, 209–214.

586 Song, W.-L., Xu, C., Smith, M.-P., Kynichy, J., Huang, K.-J., Wei, C.-W., Zhou, L., Shu,
587 Q.-H., 2016. Origin of unusual HREE-Mo-rich carbonatites in the Qinling orogen,
588 China. *Scientific Reports* 6, 37377.

589 Tang, Q.-Y, Zhang, Z.-W, Li, C.-S, Wang, Y.-A, Ripley, E.M., 2016. Neoproterozoic
590 subduction-related basaltic magmatism in the northern margin of the Tarim craton:
591 Implications for Rodinia reconstruction. *Precambrian Research* 286, 370-378.

592 Teng, F.-Z., Li, W.-Y., Ke, S., Marty, B., Dauphas, N., Huang, S.-C., Wu, F.-Y., Pourmand,
593 A., 2010. Magnesium isotopic composition of the Earth and chondrites. *Geochimica*
594 *et Cosmochimica Acta* 74, 4150-4166.

595 Thomson, A.R., Walter, M.J., Kohn, S.C., Brooker, R.A., 2016. Slab melting as a barrier to

596 deep carbon subduction. *Nature* 529, 76-79.

597 Tian, W., Campbell, I.H., Allen, C.M., Guan, P., Pan, W.-Q., Chen, M.-M., Yu, H.-J., Zhu,
598 W.-P., 2010. The Tarim picrite-basalt-rhyolite suite, a Permian flood basalt from
599 northwest china with contrasting rhyolites produced by fractional crystallization and
600 anatexis. *Contributions to Mineralogy and Petrology* 160, 407-425.

601 Wang, L., 2014. The mineralogy, geochemistry and source region of agpaitic lamprophyre
602 in Wajilitage, XinJiang. A Dissertation of Master Degree for China University of
603 Geosciences (Beijing), 1-60.

604 Wang, S.-J., Teng, F.-Z., Li, S.-G., 2014. Tracing carbonate-silicate interaction during
605 subduction using magnesium and oxygen isotopes. *Nature Communications* 5, 5328.

606 Wei, X., Xu, Y.-G., Luo, Z.-Y., Zhao, J.-X., Feng, Y.-X., 2015. Composition of the Tarim
607 mantle plume: Constraints from clinopyroxene antecrysts in the early Permian
608 Xiaohaizi dykes, NW China. *Lithos* 230, 69-81.

609 White, W.M., 2010. Oceanic island basalts and mantle plumes: the geochemical
610 perspective. *Annual Review of Earth and Planetary Sciences* 38, 133-160.

611 Woolley, A.R., Kempe, D.R.C., 1989. Carbonatites: nomenclature, average chemical
612 composition. In: Bell, K., (ed) *Carbonatites: Genesis and Evolution*. Unwin Hyman,
613 London, 1-14.

614 Xia, Q.-K., Bi, Y., Li, P., Tian, W., Wei, X., Chen, H.-L., 2016. High water content in
615 primitive continental flood basalts. *Scientific Reports* 6, 25416.

616 Xu, C., Chakhmouradian, A.N., Taylor, R.X., Kynicky, J., Li, W.-B., Song, W.-L., 2014.
617 Origin of carbonatites in the south Qinling orogen: implications for crustal recycling

618 and timing of collision between the south and north china blocks. *Geochimica et*
619 *Cosmochimica Acta* 143,189-206.

620 Xu, C., Kynicky, J., Chakhmouradian, A.R., Campbell, I.H., Allen, C.M., 2010.
621 Trace-element modeling of the magmatic evolution of rare-earth-rich carbonatite from
622 the Miaoya deposit, Central China. *Lithos* 118, 145-155.

623 Xu, X., Song, S., Allen, M.B., Ernst, R.E., Niu, Y., Su, L., 2016. An 850-820Ma LIP
624 dismembered during breakup of the Rodinia supercontinent and destroyed by Early
625 Paleozoic continental subduction in the northern Tibetan Plateau, NW
626 China. *Precambrian Research*, 282, 52-73.

627 Xu, Y.-G., Wei, X., Luo, Z.-Y., Liu, H.-Q., Cao, J., 2014. The early Permian Tarim large
628 igneous province: main characteristics and a plume incubation model. *Lithos* 204,
629 20-35.

630 Yang, Y.H., Sun, J.F., Xie, L.W., Fan, H.R., Wu, F.Y., 2008. In situ Nd isotopic
631 measurement of natural geological materials by LA-MC-ICPMS. *Chinese Science*
632 *Bulletin* 53, 1062-1070.

633 Yang, W., Teng, F.-Z., Li, W.-Y., Liu, S.-A., Ke, S., Liu, Y.-S., Zhang, H.-F., Gao, S., 2016.
634 Magnesium isotopic composition of the deep continental crust. *American Mineralogist*
635 101, 243-252.

636 Zhang, C. L., Li, X.H., Ye, H.M., 2008. Permian layered ultramafic-mafic-syenite intrusive
637 complex in western Tarim block, NW China: genetically related to a Permian mantle
638 plume under Tarim. *Journal of Geology* 116, 263-281.

639 Zhang, C.L., Li, H.K., Santosh, M., Li, Z.X., Zou, H.B., Wang, H., Ye, H., 2012.

Precambrian evolution and cratonization of the Tarim Block, NW China: Petrology, geochemistry, Nd-isotopes and U–Pb zircon geochronology from Archaen gabbro-TTG-potassic granite suite and Paleoproterozoic metamorphic belt. *Journal of Asian Earth Sciences* 47, 5-20.

Zhang, D., Zhang, Z., Mao, J., Huang, H., Cheng, Z., 2016. Zircon U–Pb ages and Hf–O isotopic signatures of the Wajilitag and Puchang Fe–Ti oxide-bearing intrusive complexes: Constraints on their source characteristics and temporal–spatial evolution of the Tarim large igneous province. *Gondwana Research* 37, 71-85.

Zhang, D.-Y., Zhang, Z.-C., Santosh, M., Cheng, Z.-G., He, H., Kang, J.-L., 2013. Perovskite and baddeleyite from kimberlitic intrusions in the Tarim large igneous province signal the onset of an end-carboniferous mantle plume. *Earth and Planetary Science Letters* 361, 238-248.

Zheng, B.H., Zhu, W.-B, Jahn, B.M., Shu, L.-S, Zhang, Z.-Y, Su, J.-B., 2010. Subducted Precambrian oceanic crust: geochemical and Sr–Nd isotopic evidence from metabasalts of the Aksu blueschist, NW China. *Journal of the Geological Society* 167, 1161-1170.

Zou, S.-Y., Li, Z.-L., Song, B., Ernst, R.E., Li, Y.-Q., Ren., Z.-Y., 2015. Zircon U–Pb dating, geochemistry and Sr–Nd–Pb–Hf isotopes of the Wajilitage alkali mafic dikes, and associated diorite and syenitic rocks: implications for magmatic evolution of the Tarim large igneous province. *Lithos* s212-215, 428-442.

Figure captions

Fig. 1. Schematic location map of the Tarim LIP (a) and geological map of the Wajilitage area (b); modified after Zhang et al. (2013), Xu et al. (2014) and Wang (2014).

Fig. 2. Weighted average plot of the $^{208}\text{Pb}/^{232}\text{Th}$ ages of monazite grains separated from the calcite carbonatites.

Fig. 3. Age comparison of carbonatites with associated igneous rocks in the Tarim LIP. Data for rhyolite, mafic and felsic intrusions, basalt and porphyritic pyroxenite-dunite are from Xu et al. (2014); for nephelinite, lamprophyre and syenite from Cheng et al. (2015), Wang et al. (2015) and Zou et al. (2015).

Fig. 4. Average geochemical data for the Wajilitage carbonatites and their carbonate minerals: (a) trace-element abundances normalized to the Primitive Mantle; (b) chondrite-normalized REE abundances.

Fig. 5. Carbon and oxygen isotopic compositions of the Wajilitage carbonatites; shown for comparison are the fields of primary, unmodified carbonatites (Keller and Hoefs, 1995), and of ambient mantle suggested by Ray et al. (1999).

Fig. 6. Variations in Mg isotopic composition of dolomite and calcite from the Wajilitage carbonatites compared to mantle peridotite, MORB, OIB (Teng et al., 2010), upper and lower crust (Li et al., 2010; Yang et al., 2016), marine carbonates (Huang et al., 2015), rift carbonatites at Oldoinyo Lengai (Li et al., 2016), and orogenic carbonatites at Qinling (Song et al., 2016).

Fig. 7. Variations in Sr-Nd isotopic composition of the Wajilitage carbonatites compared to the published data for other igneous rocks from the Tarim LIP (nephelinite, Cheng et al., 2015; lamprophyres, Wang, 2014; syenite, Zou et al., 2015; basalts, Tian et al., 2010; Xu

684 et al., 2014; and picrite, Tian et al., 2010), and for the Tarim basements (Zhang et al.,
685 2008, 2012).

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Table 1

Mg isotope compositions (‰) of carbonate minerals from carbonatites.

Sample		$\delta^{25}\text{Mg}_{\text{DSM3}}$	2SD	$\delta^{26}\text{Mg}_{\text{DSM3}}$	2SD
WR-1d	dolomite	-0.45	0.04	-0.87	0.08
WR-2d		-0.39	0.01	-0.71	0.06
WR-2d (repeat)		-0.34	0.04	-0.65	0.02
WR-3d		-0.52	0.05	-0.99	0.07
WR-4d		-0.35	0.03	-0.68	0.01
WR-4d (repeat)		-0.35	0.05	-0.69	0.03
WR-5d		-0.41	0.04	-0.81	0.07
WR-6c	calcite	-0.51	0.02	-0.96	0.04
WR-7c		-0.49	0.03	-0.96	0.05
WR-7c (repeat)		-0.51	0.02	-0.99	0.06
WR-8c		-0.53	0.02	-0.99	0.09
WR-9c		-0.43	0.03	-0.82	0.03
WR-10c		-0.53	0.02	-0.99	0.03