**Crustal Melting above a Mantle Plume: Insights from the Permian Tarim Large Igneous Province, NW China**

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**1. Introduction**

Large igneous provinces (LIPs) are the most voluminous emplacement of mantle material at the Earth’s surface after basaltic magmatism at mid-ocean ridges and at subduction zones (Coffin and Eldholm, 1994). The ascent of mantle material is not only responsible for the outpouring of huge volume of basaltic magma over a short period, but can also induce partial melting of overlying crust and result in the formation of crustally-derived silicic magmas. The study of these crustally-derived silicic rocks from LIPs has the potential to yield significant information on crustal melting regimes (Xu et al., 2008). Previous studies on rhyolites from LIPs focused on identifying their volume, eruption rate and style, and source region (Drew et al., 2013; Nash et al., 2006; Watts et al., 2011), with relatively less constraints on their temporal variation of magmatic temperature, which has hampered our understanding of crustal melting behavior associated with a thermal anomaly. The recently recognized Permian Tarim LIP (Xu et al., 2014; Yang et al., 2013; Zhang et al., 2010; Zhou et al., 2009) in northwestern China is well suited to address this issue because it contains abundant low Nb-Ta rhyolites which have been interpreted to have a crustal origin (Liu et al., 2014; Tian et al., 2010). While mafic-ultramafic rocks of the Tarim LIP have been the subject of a number of petrological and geochemical studies (He et al., 2016; Tian et al., 2010; Wei et al., 2014; Zhang et al., 2010), the study of low Nb-Ta rhyolites remains relatively scarce, and little is known about the crustal melting behavior above a mantle plume.

In this contribution, newly acquired geochemical data, combined with available data in the literature, are used to constrain the partial melting temperature for the low Nb-Ta rhyolites of the Permian Tarim LIP. It is demonstrated that there is a temporal variation for the Permian Tarim rhyolites with an increase of zircon saturation temperature (TZr) during the early stage and decreasing of TZr during the latter stage of the Permian Tarim magmatism. The temporal variation in TZr is interpreted as being indicative of the variation in magma temperature in the source region. It is argued that, given a thermal anomaly, fusible components in the crust melt at lower temperature initially and refractory materials melt at higher temperature at a later stage. The decreasing TZr towards the final stage may correspond to the waning of the thermal anomaly when the influence of the mantle plume diminishes in the region.

**2. Geological background and samples**

Bounded by the Tianshan orogen to the north and west, and the Kunlun-Karakorum orogen to the south, the Tarim craton in the Xinjiang Uygur Autonomous Region of northwestern China is one of the three major crustal blocks in China. It is covered by a vast Cenozoic desert and its basement rocks mostly outcrop along the northern (Kuerle area) and southwestern margins (Tiekelike area, Fig. 1). The craton experienced a northward drift and was amalgamated with the southern Central Asian Orogenic Belt (CAOB) during the late Paleozoic. During the late Carboniferous-early Permian, there was no significant plate motion for the Tarim craton (Fang et al., 2001). The oldest crystalline basement of the Tarim craton is interpreted to be composed of Neoarchean to early Paleoproterozoic tonalite-trondjemite-granodiorite (TTG) gneisses and supracrustal rocks (Hu et al., 2000). The supracrustal sequences are mainly composed of meta-sedimentary rocks with amphibolite layers (Dong et al., 1998; Long et al., 2011). Precambrian sequences of metasediments including marine sediments, glacial deposits and sedimentary-volcanogenic series are distributed along the northern and southwestern margins of the Tarim craton (Wang et al., 2009). Neoproterozoic granitoids mainly consisting of gneissic biotite granites, granodiorites and tonalities, intruded the basement rocks and Paleoproterozoic to Mesoproterozoic sedimentary sequences (Wu et al., 2009). Little is known about the Mesoproterozoic-Neoproterozoic history of the Tarim craton. Recently discovered Proterozoic-age zircon grains both in metasedimentary and igneous rocks (Wang et al., 2009; Wu et al., 2009), however, may indicate the existence of extensive Mesoproterozoic-Neoproterozoic crystalline basement in the Tarim craton.

Several phases of igneous activity have been reported in the north Tarim craton, i.e., the Archean, early Paleoproterozoic, Neoproterzoic and early Permian events (Gao et al., 2011; Hu et al., 2000; Tian et al., 2010; Zhang et al., 2010). Amongst these magmatic events, the early Permian phase, named Permian Tarim LIP, is notable for its synchronous emplacement of ultramafic-mafic-silicic rocks. It covers an area of 250000-300000 km2 (Fig. 1B; Tian et al., 2010; Yang et al., 2013), consisting of ~300 Ma small-volume diamond-bearing kimberlites and V-Ti magnetite bearing gabbroic intrusion, ~290 Ma continental flood basalts (CFB), ~280 Ma layered mafic-ultramafic intrusions and bimodal dykes, and associated rhyolites, and minor andesites and silicic ignimbrites (He et al., 2016; Liu et al., 2014; Tian et al., 2010; Xu et al., 2014; Zhang et al., 2013). The Tarim flood basalts show fractionated light rare earth elements (REE) and flat heavy REE patterns with small negative Nb-Ta anomalies on primitive mantle normalized diagrams, with high (87Sr/86Sr)i from 0.70623 to 0.70803, relatively uniform εNd(t) values (mainly from -3.8 to -2.3) and low (206Pb/204Pb)i (17.43–17.57; Li et al., 2012; Wei et al., 2014; Xu et al., 2014). The ~280 Ma basaltic magmas display oceanic island basalt (OIB)-like trace element characteristics with lower (87Sr/86Sr)i (from 0.70472 to 0.70684), and higher εNd(t) from -0.3 to 4.8 and (206Pb/204Pb)i from 17.50 to 18.11 than the ~290 Ma basalts (Wei et al., 2014; Xu et al., 2014; Zhou et al., 2009). To account for such a temporal variation in magma types and chemical composition in the Tarim LIP, an incubating mantle plume model has been proposed (Xu et al., 2014). Within this tectonic scenario, the first two episodes of magmatism are mainly derived from the melting of enriched components of the subcontinental lithospheric mantle (SCLM) as a result of heating from the incubating mantle plume. Only thin-spots within the Tarim craton and at the craton-orogen boundary permitted decompression melting of the mantle plume, which generated the ~300 Ma and ~280 Ma layered mafic-ultramafic intrusions and dykes (He et al., 2016; Xu et al., 2014).

In the northern part of the Tarim basin, the Permian Tarim silicic volcanics are covered by the Cenozoic desert. Combined with drilling and log records, seismic profiles reveal that the Permian Tarim silicic volcanics marginally overlie the Permian CFB (Fig. 1B; Liu et al., 2014; ShangGuan et al., 2011; Tian et al., 2010; Xu et al., 2014; Yu et al., 2011). The Permian Tarim silicic volcanics in the inner part of the Tarim basin have an area of ~48000 km2 and thickness of 100-300 m, consisting of rhyolites, minor andesites and silicic ignimbrites (Fig. 1; Liu et al., 2014; Tian et al., 2010; Yang et al., 2013). The Permian rhyolites also crop out alongside the northern margin of the Tarim craton in the Wenquan, Kuchehe and Dawanqi regions (Fig. 1C), defining a nearly E-W striking volcanic belt more than 180 km long, 2-12 km wide, and 1-4 km thick with a total exposure area of ~570 km2 (Fig. 1C). Stratigraphically, these Permian Tarim rhyolites belong to the upper part of the Xiaotikanlike Formation.

The Permian rhyolites from the northern Tarim basin and the Wenquan section were subdivided into two groups based on their Nb-Ta contents: the low Nb-Ta rhyolites are characterized by significant negative Nb-Ta anomalies (drilling holes: Ha1, Ha2, S79, S99, S102 and S114; outcrops: Wenquan section) and the high Nb-Ta rhyolites exhibit small negative to positive Nb-Ta anomalies (drilling holes: SX1, SL1-6, MN1-1, MN1-2, YM5-8, YM16-1, YM16-2, Liu et al., 2014; Tian et al., 2010). Eighteen zircon SIMS U-Pb analyses yield a concordia age of 287.3 ± 2.0 Ma for the rhyolite Ha2 from the Halahatang region of the northern Tarim basin (ShangGuan et al., 2011), which is comparable with CA-ID-TIMS analysis on zircons recovered from one sample from the top of the Wenquan section (WQ09-2: 286.8 ± 0.5 Ma; Liu et al., 2014). Younger zircon U-Pb ages were also reported for the low Nb-Ta rhyolites in the neighboring regions (S79-114; 282 ± 3-274 ± 3 Ma; Yu et al., 2011).

The Kuchehe silicic volcanics, located ~80 km north of Kuche county, has an outcrop area of ~130 km2 (Fig. 1C). They unconformably overlie the Devonian volcanic strata to the north and are unconformably overlain by the Triassic terrigenous clastic rocks to the south. The Kuchehe silicic volcanics are emplaced as successive lava flows; they are massive, with pink coloration in the lower part of the succession and the purple-pink color in the upper part. Cogenetic agglomerates and ignimbrites can be found among the top of the upper part of purple rhyolite. The agglomerates have a texture and mineralogy similar to the host rock. Purple-red, euhedral garnet (1-10 mm) and black biotite (5-10 mm) crystals have been identified in hand specimen. Thin sections show that phenocrysts have proportions of ~30-50 % in volume, consisting of euhedral quartz (10-30%), euhedral to anhedral plagioclase and alkali feldspar (20-30%), euhedral biotite (2-3%) and euhedral garnet (1-2%, Fig. 2A and B). In the Dawanqi region, the Permian silicic volcanics have an exposure area of ~370 km2. The Dawanqi silicic volcanics conformably overlie the sedimentary rocks of the Xiaotikanlike Formation to the north and unconformably underlie the Triassic sedimentary strata to the south (Fig. 1). They are emplaced as flow deposits and massive with black grey to grey color, and show pyroclastic texture in thin sections, containing euhedral to anhedral quartz (15-20%) and euhedral to anhedral plagioclase and alkali feldspar (5-20%) as phenocryst phases, and the remainder is quartz-feldspathic groundmass (75-80%; Fig. 2C and D). Seventeen samples from the Kuchehe (10) and Dawanqi (7) sections were collected for chemical analysis (K07-2 is an agglomerate sampled at the top of the Kuchehe rhyolites). Two samples from the top of each section were selected for zircon U-Pb analysis (Kuchehe: K07-1; Dawanqi: DWQ09-6) and three samples were prepared for zircon in-situ Hf analysis (Kuchehe: K03-1, K07-1; Dawanqi: DWQ09-6). For comparison, a summary of mineral assemblages of silicic rocks from the Tarim LIP, Chon Aike SLIP of Patagonia and SRP-YP volcanic Provinces is presented in Table 1.

**3. Analytical methods**

Cathodoluminescence (CL) images of the zircon grains were obtained using an electron microprobe (JXA-8100, JEOL) equipped with a Mono CL3 detector (Gatan) at the Guangzhou Institute of Geochemistry, Chinese Academy of Sciences (GIGCAS). Zircon grains recovered from the Kuchehe sample (K07-1) suffered post-alteration because cracks are observed in plane-polarized light (not shown). The Dawanqi zircons (DWQ09-6), however, are relatively fresh as no cracks are observed in plane-polarized light. Thus, CA-ID-TIMS method was chosen for the Kuchehe zircons and the LA-ICPMS method was chosen for the Dawanqi zircons. The U-Pb CA-ID-TIMS analyses were carried out at the Berkeley Geochronology Center, following the procedures described in Mundil et al. (2004). Comparison of age results is facilitated by inter-laboratory cross calibration (Black et al., 2003b, 2004) and analyses of reference materials as well as analyses of calibration solution distributed by the Earthtime initiative (Irmis et al., 2011). In-situ U-Pb analyses for the Dawanqi zircons (DWQ09-6) were performed using an Agilent 7500a ICP-MS coupled with a Resonetic RESOLution M-50 ArF-Excimer laser source (λ=193nm) at GIGCAS. During analysis, spot sizes of ~31 μm, with a laser frequency of 10 Hz at 80 mJ, were used. Helium was used as the carrier gas to enhance the transport efficiency. NIST610 and TEMORA (417 Ma) were used as external calibration standards (Black et al., 2003a), and 29Si as internal standard. Each block of 5 samples was bracketed by analyses of standards. Off-line inspection and integration of background and analyte signals, and time-drift correction and quantitative calibration for U-Pb dating were performed using ICPMSDataCal (Liu et al., 2010). Detailed analytical procedures can be found in Li et al. (2012). Concordia diagrams are plotted and weighted mean 206Pb/238U ages are calculated by using Isoplot/Ex\_ver3 (Ludwig, 2003). Analytical errors of 206Pb/238U are better than 0.4 % (2σ) for CA-ID-TIMS and 2 % (1σ) for LA-ICPMS (Table A.1).

Major oxides of whole-rocks were carried out by X-ray fluorescence spectrometry (XRF) on fused glass disks by utilizing Rigaku ZSX-100e XRF instrument at GIGCAS. Detailed analytical procedure is given by Goto and Tatsumi (1996). Trace elements were determined with Perkin-Elmer Sciex ElAN 6000 ICP-MS, following the techniques described by Liu et al. (1996). Analytical uncertainties for major oxides are mostly between 1 % and 5%, and for REE and other incompatible elements are generally better than 5% (Table A.2).

For Nd isotope analyses, rare earth elements (REEs) were separated by cation exchange columns, followed by REE separation using HDEHP-coated Kef columns to obtain Nd. Nd isotopic ratios were measured with a Micromass Isoprobe Multi-Collector ICPMS at GIGCAS. The JNDi-1 standards were used as monitors of detector efficiency drift of the instrument for Nd isotopes. It was repeatedly measured during the analysis and gave an average of 0.512117 ± 5 (2σ, n=4). BHVO-2 was used as the standard to monitor the instrumental performance for Nd isotopes. Multiple analyses of BHVO-2 yielded an average 143Nd/144Nd ratio of 0.512945 ± 9 (2σ). Detailed analytical procedures can be found in Wei et al. (2002) and Liang et al. (2003).

In-situ zircon Hf isotopic analyses were measured by using the Neptune multi-collector ICPMS equipped with a 193 nm laser, at the Institute of Geology and Geophysics, Chinese Academy of Sciences in Beijing, China. During the course of this study, a laser repetition rate of 6 Hz at 100 mJ was used and beam diameters were constant at ~44 μm. During analyses, the 176Hf/177Hf and 176Lu/177Hf ratios of the standard zircon (GJ-1) were 0.282009 ± 10 (2σ, n = 20) and 0.00025, similar to a weighted mean 176Hf/177Hf ratios of 0.282000 ± 5 (2σ) measured using the solution analysis (Morel et al., 2008). The detailed analytical technique and data correction procedure are described in Wu et al. (2006).

**4. Results**

4.1 Zircon U-Pb dating

Zircon U-Pb isotopic results and cathodoluminescence (CL) images are presented in Table A.1, Fig. A.1 and Fig. 3. Zircons from the Kuchehe and Dawanqi silicic volcanic rocks exhibit fine-scale oscillatory zoning and have high Th/U ratios of 0.3-0.6, consistent with crystallization from a silicic magma. Twelve CA-ID-TIMS U-Pb analyses for the Kuchehe sample yielded a weighted mean 206Pb/238U age of 292.6 ± 0.5 Ma (2σ, MSWD = 5.3, Fig. 3A). Except spot 9 with an exceptionally high U-Pb age and plotting off the concordia curve, nineteen LA-ICPMS U-Pb analyses for the Dawanqi sample form a relatively coherent cluster and yield a weighted mean 206Pb/238U age of 274 ± 2 Ma (2σ, MSWD = 1.2, Fig. 3B). These results suggest that the Kuchehe and Dawanqi volcanic rocks were erupted at ~293 Ma and ~274 Ma respectively.

4.2 Major and trace elements

Whole-rock major and trace elements for the Kuchehe and Dawanqi silicic volcanics are reported in Table A.2 and Figs. 4-6. For the purpose of this study, published data for silicic rocks from the Permian Tarim LIP, Chon Aike silicic LIP and Snake River Plateau-Yellowstone Plateau volcanic provinces (SRP-YP; SiO2>60 wt.%) are also plotted for comparison. Although other compositions of silicic rocks are present in these LIPs or volcanic provinces (Fig. 4), the collective term “rhyolite” is used for simplicity. Both the Kuchehe and Dawanqi rhyolites are peraluminous with high A/CNK (molar Al2O3/[CaO+Na2O+K2O]>1.0) and A/NK values (molar Al2O3/[Na2O+K2O]>1.1). However, they are different in other geochemical characteristics with the Kuchehe rhyolites having low (Zr+Nb+Ce+Y) contents (256-344 ppm) and 10000Ga/Al ratios (molar ratio Ga/Al, 2.3-3.0), whereas the Dawanqi rhyolites having high (Zr+Nb+Ce+Y) contents (338-2390 ppm) and 10000Ga/Al ratios (>2.6, Table A.2, Fig. 5A, B and C). The agglomerate collected from the Kuchehe section (K07-2) has a chemical composition within the range defined by other Kuchehe rhyolites (Figs. 4-6).

In the Chondrite-normalized rare earth element (REE) diagrams, the Kuchehe and Dawanqi rhyolites exhibit LREE-enriched patterns (LaN/YbN = 1.2-62.9; mainly > 12) characterized by negative Eu anomalies with the exception of one Dawanqi sample (DWQ09-5). In the primitive mantle (PM)-normalized multi-trace element diagrams, the Kuchehe and Dawanqi rhyolites exhibit negative Ba, Sr, Eu, Nb and Ta anomalies (Fig. 6). The Dawanqi rhyolites exhibit a wide variation in both REE and multi-trace element diagrams (Fig. 6B and E). The Kuchehe and Dawanqi rhyolites exhibit negative Nb-Ta anomalies in multi-trace element diagrams, belonging to the low Nb-Ta rhyolite type defined by Liu et al. (2014) (Fig. 6D and E). For convenience, the low Nb-Ta rhyolites in the following discussion denote both the Kuchehe and Dawanqi rhyolites and the other low Nb-Ta rhyolites from northern Tarim unless it is specified.

4.3 Whole-rock Nd isotopes and in-situ zircon Hf isotopes

Whole-rock Nd isotopic data and zircon Hf isotopic data are reported in Table A.3, A.4 and Fig. 7. Nd-Hf isotopes for mafic-ultramafic rocks and other rhyolites from the Tarim LIP are also presented for comparison. The Kuchehe rhyolites have unradiogenic 143Nd/144Nd ratios (εNd[t] = -8.5 to -7.9, corrected to eruption ages), corresponding to two-stage Nd model ages (T2DM, using formulas and parameters provided in Li et al. (2003)) of 1710-1755 Ma. The Dawanqi rhyolites have unradiogenic 143Nd/144Nd (εNd[t] = -5.4 to -5.2) ratios and T2DM of 1472-1485 Ma, which marginally overlaps with other low Nb-Ta rhyolites (Fig. 7A). As a whole, the Tarim low Nb-Ta rhyolites have marginally lower εNd(t) values than those of the Tarim LIP mafic-ultramafic rocks (εNd[t] = -5.2 - +5.2, Fig. 7A; Wei et al., 2014; Xu et al., 2014). The high Nb-Ta rhyolites, however, have εNd values within the range of the Tarim LIP mafic-ultramafic rocks. Zircon 176Hf/177Hf ratios of the Kuchehe rhyolites display a wide range of values, with εHf(t=292.6 Ma) values ranging from -9.0 to -2.4 and T2DM varying from 1459 Ma to 1876 Ma, and partially overlap with whole rock Hf isotope composition of the Tarim LIP mafic-ultramafic rocks(εHf[t]=-6.2 - +2.8, Fig. 7B; Li et al., 2012). Variation of Hf isotopic compositions for the Dawanqi and other low Nb-Ta rhyolites are more limited with εHf(t) values of -4.8 to -1.6, corresponding to T2DM of 1395 Ma to 1593 Ma and lying within the range of Tarim LIP mafic-ultramafic rocks (Table A.4 and Fig. 7B).

4.4 Zircon saturation temperature

For the purpose of this study, geochemical data for the low Nb-Ta rhyolites from the Tarim LIP, the Chon Aike silicic LIP of Patagonia and the SRP-YP volcanic provinces are used to calculate zircon saturation temperature, using the thermometer of Watson and Harrison (1983):

TZr=12900/(2.95 + 0.85 M + ln[496,000/Zr]); M=(Na + K + 2·Ca)/(Al·Si).

The results are presented in Table 1 and Fig. 8. The Permian Tarim rhyolites were erupted in three distinct stages: the early stage rhyolites (~293 Ma) from the Kuchehe section, the middle stage rhyolites from boreholes Ha1, Ha2, and the Wenquan section (~287 Ma, Liu et al., 2014; Yu et al., 2011), and the late stage rhyolites from borehole S79, S99, S102, S114, and the Dawanqi section (282-274 Ma, Yu et al., 2011; and this study), respectively. The early stage rhyolites have lower TZr (793 oC to 827 oC) relative to the middle stage rhyolites (886-936 oC), whilst the late stage rhyolites have lower TZr (803-908 oC, Table 1 and Fig. 8A) than the middle stage volcanic event. Overall the Tarim rhyolites exhibit a decline in TZr between the middle and late stages, preceded by an upward trend between the early and middle stages. The Chon Aike rhyolites yield TZr range of 696-938 oC with the majority between 750 oC and 850 oC (Fig. 8B). Rhyolites from the Marifil Formation, Mapple Formation, and Chon Aike Formation exhibit increasing trends from 719-805 oC to 821-908 oC, 744-861 oC to 748-885 oC and 737-800 oC to 744-938 oC at earlier stages, but decline to 779-891 oC, 696-867 oC and 739-766 oC respectively. The El Quemado rhyolites exhibit an upward trend of TZr from 746-805 oC at ~182 Ma to 775-835 oC at ~163 Ma, and then most of the samples decrease to 721-801 oC, with the exception of two samples that have relatively high TZr values of 834-931 oC (Table 1 and Fig. 8B). However, no regular pattern is identified for rhyolites from the Ibañez Formation. TZr of the SRP-YP rhyolites range from 746 to 969 oC with the majority above 800 oC (Table 1 and Fig. 8C). Three cycles are recognized for the SRP-YP rhyolites based on their eruption ages and TZr results: The first cycle rhyolites (16.0-15.2 Ma) are from the High Rock Caldera complex (HRCC) and Virgin Valley, Paradise Valley, Huntington Creek, Antonne Wash, Overton Wash and Peacock (OV) volcanic provinces; The second cycle rhyolites (15.2-7.5 Ma) are from the Bruneau-Jarbidge (B-J), Mount Bennett Hills (MBH) and Twin Falls (TF) volcanic provinces; and the third cycle rhyolites (7.5-0 Ma) are from the Heise and Yellowstone volcanic provinces. In each cycle, TZr exhibits an increasing trend at an earlier stage but a declining trend in a later stage (Fig. 8C). This classification is approximately synchronous to Nash et al. (2006)’s classification, which is based on major changes in eruption frequency, composition and magma temperature. TZr results for the Brennecke and Mount Poster Formation (early, ~183 Ma) samples of the Chon Aike silicic LIP and the peralkaline samples of the SRP-YP are excluded in the following discussion because the formers have undergone extensive upper crustal contamination (Riley et al., 2001) and the latter are peralkaline and are not suitable for TZr calculation (Watson and Harrison, 1983). Boehnke et al. (2013) optimized the calculation of the coefficients of Watson and Harrison (1983) by using a Bayesian approach. However, it yielded similar temporal patterns and in some cases, unrealistic low TZr values for rhyolites from the Tarim LIP, the Chon Aike silicic LIP of Patagonia and the SRP-YP volcanic provinces (not shown). The calculation of Watson and Harrison (1983) is therefore favored in this study.

**5. Discussion**

5.1 Post-eruption alteration and crystal fractionation

Before the chemical compositions of the Kuchehe and Dawanqi rhyolites are used to constrain their petrogenesis and magmatic temperatures, the effects of post-eruption alteration and crystal fractionation must be evaluated. Other low Nb-Ta rhyolites from the Permian Tarim LIP are also presented for discussion (Figs. 9 and 10; Liu et al., 2014; Tian et al., 2010; Yu et al., 2011). K, Na and Ca are the most mobile elements in post-eruption alteration process. The Dawanqi rhyolites exhibit low CaO (< 0.36 wt.%) and Na2O contents (< 2.10 wt.%), and high K2O contents (> 6.79 wt.%) (Fig. 9), indicative of significant post-eruption alteration. Plots of major elements and key parameters versus LOI are used to double check whether the rest samples suffered significant post-eruption alteration of feldspars because: 1) feldspars are the most abundant minerals in silicic rocks; and 2) their alteration products such as kaolinites contain considerable structure water. Two Kuchehe rhyolites (K05-2 and K06-1) may have also been subject to post-eruption alteration as they are anomalous in comparison to other Kuchehe rhyolites in a series of plots (Figs. 9 and 10). Sample K05-2 is interpreted to have undergone Na loss as it deviates from other Kuchehe rhyolites in the plots of A/CNK and Na2O vs. LOI (Fig. 9B and E). Sample K06-1 might have gained Na and undergone CaO loss as it exhibits the highest Na2O and lowest CaO contents (Fig. 9). However, the effects of post-eruption alteration on key parameters such as A/CNK and M values for the majority of the Kuchehe, Halahatang, S and Wenquan rhyolites are relatively minor, given their low loss on ignition (LOI, mainly < 2.1 wt.%; Fig. 9; Table A.2) and relatively consistent A/CNK and M values over variable range of LOI (Fig. 9E and F). Given the uncertainty regarding the alteration history of samples (K05-2, K06-1, and all the Dawanqi samples), they are excluded in the following discussion on major element compositions. Their Zr contents, however, could be used in the estimation of zircon saturation temperature because: 1) Zr is relatively immobile during post-eruption alteration; and 2) a large uncertainty in major element composition (M: 0.9-1.9) would only introduce a small error in estimated temperature (within 50 oC, Miller et al., 2003). Trace element ratios such as Zr/Nb, Nb/La and Th/La are relatively immobile during post-eruption alteration and are also used to constrain the petrogenesis. This is supported by the broadly positive correlations between Zr and Nb for the low Nb-Ta rhyolites (Fig. 11A). Two Dawanqi rhyolites (DWQ09-4 and 09-6) from the Dawanqi section have extremely high Zr and Nb contents (Figs. 10-11) in comparison to other Dawanqi samples, which might have undergone upper crustal contamination and/or significant post-eruption alteration and are excluded in the following discussion.

Feldspar fractionation is interpreted to have taken place during the petrogenesis of the Permian Tarim rhyolites as suggested by negative anomalies of Eu in REE diagrams and negative anomalies of Ba, Sr and Eu in multi-trace element diagrams (Fig. 6). Fractionation of Nb-bearing phases in silicic rocks leads to the depletion of Nb-Ta relative to U and La in multi-trace element diagram. The negative correlation between Nb and silica would be expected if the negative Nb-Ta anomalies are induced by crystal fractionation. However, each suite of the Permian Tarim rhyolites except two samples (DWQ09-4 and 09-6) have consistent Nb contents over variable silica contents (Fig. 10). This feature suggests that the negative Nb-Ta anomalies for the majority of the Permian Tarim low Nb-Ta rhyolites are likely inherited from the source.

5.2 Genetic type and constraints on the petrogenesis of the Kuchehe and Dawanqi rhyolites

Although both the Kuchehe and other low Nb-Ta rhyolites form the Tarim LIP belong to the low Nb-Ta rhyolites type, they are different in genetic types based on their whole rock geochemistry. The Kuchehe rhyolites are characterized by high A/CNK values and low Na2O contents, 10000Ga/Al ratios and (Zr+Nb+Ce+Y) contents (Fig. 5 and Table A.2). These chemical signatures, coupled with the presence of garnet in the samples, indicate that the Kuchehe rhyolites are geochemically similar to S-type granitoids (Chappell and White, 1992). Other low Nb-Ta rhyolites from the Wenquan section and the drill holes in the northern Tarim basin are metaluminous to peraluminous with high A/CNK values, 10000Ga/Al ratios and (Zr+Nb+Ce+Y) contents (Liu et al., 2014; Tian et al., 2010; Yu et al., 2011; and this study), consistent with the characteristics of aluminous A-type granitoids (Fig. 5; King et al., 1997). Identifying the genetic type for the Dawanqi rhyolites is difficult because their major elements were subject to significant post-eruption alteration. However, they might be equivalent to aluminous A-type granitoids based on their high 10000Ga/Al ratios and (Zr+Nb+Ce+Y) contents (Fig. 5C).

Silicic magmatism in LIPs is typically the result of variable thermal and mass exchange between mantle-derived basaltic magma and crustal material. In order to explore the petrogenetic relationships between the rhyolites and mafic-ultramafic rocks in the Tarim LIP, geochronological data and Nd-Hf isotope compositions of ultramafic-mafic-silicic rocks are compiled and shown in Fig. 7 and the Y-axis of Fig. 8A for comparison respectively (Li et al., 2012; Tian et al., 2010; Wei et al. 2014; Xu et al., 2014; Yang et al., 2006; Zhang et al., 2010; Zhou et al., 2009). The Kuchehe (292.6 ± 0.5 Ma) and Dawanqi rhyolites (274 ± 2 Ma) lie within the outcrop extent of the Tarim mafic-ultramafic magmatism (Fig. 8A) and exhibit whole rock Nd isotope compositions marginally overlap with those of the Tarim mafic-ultramafic rocks. Also, the low Nb-Ta rhyolites exhibit in-situ zircon Hf isotope compositions partially overlap within the range of whole rock Hf isotope composition of the Tarim mafic-ultramafic rocks (Fig. 7). This allows the possibility that the silicic rocks may have been generated via fractionation of associated basaltic rocks, coupled with crustal assimilation (AFC) or melting-assimilation-storage-hybridization model (MASH, Hildreth and Moorbath, 1988). Although the primitive geochemical signature of the basaltic rocks might be overprinted by the crustal materials in the magma chamber, these two models emphasize the compositional contribution of injected basaltic/parental magmas from the mantle (Hildreth and Moorbath, 1988).

Intermediate geochemical characteristics between the asthenosphere mantle and the continental crust would be expected if a suite of silicic magmas were generated by AFC or MASH processes of mantle-derived basaltic magmas. This is because the mantle and the continental crust have different geochemical characteristics such as trace element ratios and Nd isotope composition (Figs. 7 and 11). The crust-mantle interaction in the Tarim LIP is recorded by the geochemical characteristics of the high Nb-Ta rhyolites. The high Nb-Ta rhyolites exhibit Nd isotope compositions within the range of the Tarim LIP mafic-ultramafic rocks, with trace element ratios such as Nb/La and Th/La, intermediate between the Tarim LIP mafic-ultramafic rocks and the Tarim basement rocks (Fig. 7 and 11). The high Nb-Ta rhyolites and the Tarim LIP mafic-ultramafic rocks form a coherent trend in the plot of Nb versus Zr. Moreover, the high Nb-Ta rhyolites exhibit a decrease of εNd(t) with the increase of silica (Fig. 11; Liu et al., 2014). These features had been interpreted as hybrid products of crystal fractionation of mafic rocks, coupled with crustal assimilation (AFC, Liu et al., 2014).

However, the Kuchehe and Dawanqi rhyolites display continental crust-like trace element signatures in multi-trace element diagrams (negative Nb-Ta anomalies, Fig. 6) and deviate from the Nb-Zr trend defined by the high Nb-Ta rhyolites and the Tarim LIP mafic-ultramafic rocks (Fig. 11A). They also exhibit trace element ratios within the Tarim basement or near the lower crust end-member (Fig. 11). Moreover, the Kuchehe and Dawanqi rhyolites have relatively constant εNd(t) values over a variable range of silica (Fig. 11C). Besides, rhyolite is the only rock type in the Kuchehe and Dawanqi sections. These features indicate that the Permian Tarim mantle plume may have only provided the heat source for the generation of the Kuchehe and Dawanqi rhyolites. These results, together with high δ18Ozircon values (7.7 - 9.2 ‰) for the Wenquan and Ha rhyolites (Liu et al., 2014), collectively suggest that the low Nb-Ta rhyolites are derived from partial melting of the Tarim basement rocks. Newly underplated juvenile crust might have contributed to the low Nb-Ta rhyolites as they exhibit Nd-Hf isotope composition between the depleted mantle and Early-Middle Proterozoic crust (Fig. 7). The juvenile basaltic materials could be incorporated into the Tarim lower crust at previous magmatic events, during which the Nd isotope composition of the crustal source was homogenized by AFC or MASH processes as the low Nb-Ta rhyolite exhibit relative constant Nd isotope composition over a variable range of silica (Fig. 11C).

5.3 Temporal variation of magma temperature for crustally-derived rhyolites from the Tarim LIP

5.3.1 Zircon saturation thermometry as a viable method to assess silicic magma temperature

Estimating temperature of silicic magma generation is problematic because: 1) the absence of suitable mineral pairs with temperature-sensitive exchange reactions; and 2) fractional crystallization and melt segregation during magmatic processes (Miller et al., 2003). An alternative method to assess the temperature of magma generation is to constrain their minimum magmatic temperature at an early stage of crystallization because silicic magmas usually ascend adiabatically (Miles et al., 2013). Zircon saturation thermometry provides a simple but robust means of estimating magmatic temperature for crustally-derived silicic rocks (Miller et al., 2003; Watson and Harrison, 1983). This is because: 1) The solubility of zircon is strongly dependent on temperature, and to a lesser extent on other factors (Watson and Harrison, 1983); and 2) Zircon is ubiquitous in intermediate to silicic magmas (Miller et al., 2003). For silicic rocks, the presence or absence of inherited zircons crystals is a useful indicator as to whether a suite of samples gives maximum or minimum estimates of initial magma temperature (Miller et al., 2003). The successful application of TZr to natural silicic samples is robust if the magmatic liquid can be accurately assessed and the whole-rock composition is as close as possible to the primitive melt composition when zircons crystallize (Miller et al., 2003). Miller et al. (2003) successfully recognized the low- and high- temperature (or cold- and hot-) granitoids by the utilization of TZr.

5.3.2 Geochemical and experimental constraints on temporal variation of initial magma temperature

Water is essential to generate silicic magmas at crustal depths, because anhydrous melting requires unrealistically high temperature (Miller et al., 2003). Melting experiments emphasize the dehydration of muscovite, biotite or amphibole as the source of water at crustal levels (Gardien et al., 1995; Patiño Douce and Johnston, 1991; Vielzeuf and Montel, 1994), with the breakdown reactions of muscovite, biotite and amphibole approximately in the order of increasing temperature at crustal pressure (Fig. 12; Johannes and Holtz, 1996; Wyllie and Wolf, 1993). All of the TZr results for the Tarim LIP give minimum estimates of initial magma temperature prior to extensive crystallization because of the absences of any inherited zircon in rhyolitic rocks (Miller et al., 2003; and this study). The TZr of the Tarim rhyolites show an increasing trend from 293 Ma to 287 Ma and then a decreasing trend from 287 Ma to 274 Ma (Fig. 8A). The early stage rhyolites of the Tarim LIP have TZr values of 794-827 oC and overlap with the biotite dehydration field, consistent with biotite dehydration-melting reaction in their source rocks. The middle to late stage rhyolites have higher TZr of 803-936 oC, overlapping the biotite and amphibolite dehydration fields, suggesting biotite±amphibole dehydration reactions in the source. The comparison of zircon saturation temperature of the Tarim rhyolites and dehydration reactions therefore suggests that the early stage rhyolites and the two later stage rhyolites were derived from fusible and refractory materials respectively.

5.3.3 Comparison with the Chon Aike silicic LIP and the SRP-YP volcanic provinces

Almost all the TZr results for the Chon Aike silicic LIP and SRP-YP silicic rocks give minimum estimates of initial magma temperature because there are essentially no inherited zircon grains in host rocks (Coble and Mahood, 2016; Drew et al., 2013; Pankhurst et al., 2000; Perkins et al., 1995; Riley et al., 2001). Several samples from the Mapple Formation (~170 Ma) of the Chon Aike silicic LIP and the Picabo, Heise and Yellowstone volcanic provinces of the SRP-YP may yield maximum estimates given the presence of inherited zircon grains (Drew et al., 2013; Pankhurst et al., 2000; Watts et al., 2011). The TZr of silicic rocks from the Marifil (~185 Ma), Mapple and Chon Aike (~170 Ma) Formations of the Chon Aike silicic LIP and each cycle of the SRP-YP volcanic provinces exhibits an increasing trend at early-middle stages but decreasing trend at middle-late stages, similar to that of the Tarim rhyolites. The El Quemado Formation (~154 Ma) rhyolites of the Chon Aike Province generally exhibit a similar pattern with the exception of two samples emplaced during the latter stage that deviate from the decreasing trend with high TZr temperatures, which might result from significant upper crustal contamination (Riley et al., 2001). No regular pattern is found for rhyolites from the Ibañez Formation (~154 Ma) (Fig. 8B), probably due to lack of sufficient geochronologic and geochemical data. Independent studies using other mineral pair geothermometers such as two feldspars, clinopyroxene-fayalite and Fe-Ti oxides thermometers yield estimates of magma temperature for the SRP-YP silicic rocks that range from 755 oC to 1050 oC (Bonnichsen et al., 2008; Cathey and Nash, 2004; Ellis et al., 2010; Hildreth et al., 1984; Honjo et al., 1992; Vazquez et al., 2009). These results are essentially similar to TZr estimates obtained in this study. Specifically, the temporal variation of Fe-Ti oxide temperatures for the SRP-YP silicic rocks of 12.7-7.5 Ma and 2.0-0.0 Ma is similar to that of zircon saturation temperature (Fig. 8C; Bonnichsen et al., 2008; Cathey and Nash, 2004; Hildreth et al., 1984; Vazquez et al., 2009).

5.4 Insights into crustal melting behavior above a mantle plume

Silicic rocks from the Tarim LIP, the Marifil, Mapple and Chon Aike Formations of the Chon Aike silicic LIP of Patagonia and each cycle of the SRP-YP volcanic provinces show similar temporal variations in zircon saturation temperature estimates, highlighting the general behavior of crustal melting above a mantle plume (Fig. 8). Specifically, TZr results for the Tarim LIP rhyolites increase and then decrease from the early stage to the later stages, which reflects the melting of fusible to refractory crustal components during the early stage and then from refractory to fusible components during the later stage. First, the early stage rhyolites were derived from the melting of biotite-bearing rocks, i.e., fusible components (Fig. 12A and B); Second, the middle to late stage rhyolites were derived from an amphibole-bearing source which has a refractory nature and requires higher temperature to melt (Fig. 12A and B; Johannes and Holtz, 1996); Finally, although erupted at the final stage of the Tarim LIP, the late stage rhyolites have higher TZr than the early stage rhyolites, and show a gradual decrease in TZr. This decrease in TZr might result from the thermal decay of the mantle plume in the region (Campbell and Griffiths, 1990). First, although the thermal impact is waning, the overlying lithospheric mantle might have been further eroded and thinned by persistent thermomechanical erosion of the mantle plume, which would permit conductive heat transfer to a shallower depth in the crust and result in partial melting of fusible materials, producing low temperature rhyolites. Second, hot mantle plume materials might flow to thinner spots or weak zones such as the craton-orogen boundary due to persistent mass supply from the feeder conduit, and cause partial melting of fusible components and produce low temperature magmas at a new site in the lithosphere. Third, when the mantle plume moves to a new site because of plate drifting, fusible components would start to melt and yield low temperature magmas. It may initiate a new cycle but it constitutes the late stage of the previous cycle.

The impact of the thermal effects of a mantle plume on the overlying lithosphere is of fundamental importance to understanding the temporo-spatial evolution of magmatism in mantle plume-related LIPs or volcanic provinces (Davies, 1994; Gibson et al., 2006; Xu et al., 2008, 2014). The mantle plume heat transfer to the crust could be accomplished by conduction and magmatic underplating (Xu et al., 2008). Mantle geochemical signatures would be usually anticipated in silicic rocks if they are generated by magmatic underplating because of mass exchange between the basaltic magmas and ambient rocks (Xu et al., 2008). In this case, the mantle plume heat is transferred by direct contact between the basaltic magmatism and crustal materials. The Tarim low Nb-Ta rhyolites, especially those from the Kuchehe, Wenquan and Dawanqi sections, however, are emplaced near the craton-orogen boundary and exhibit continental crust geochemical characteristics, precluding that the heat source for the Tarim low Nb-Ta rhyolites was provided by magmatic underplating. Conductive heating from the Permian Tarim mantle plume through the lithospheric mantle, without direct contact between the mantle plume materials and the overlying crust, is therefore preferred to explain the generation of the Tarim LIP rhyolites. A conductive heating model (Fig. 12A and B), augmented by a coupled thermomechanical erosion process, to investigate the behavior of crustal melting above a mantle plume following the method described by Davies (1994) is described here. In this model, a potential temperature of 1600 oC is used for a ‘generic’ mantle plume and an initial lithosphere thickness of 120 km for the northern margin of the Permian Tarim lithosphere is estimated because a craton margin is relatively thin. The initial geotherm is calculated for a mantle potential temperature of 1300 oC using the method of Davies (1994). A temporal dynamical model with emphasis on the generation of the Permian Tarim low Nb-Ta rhyolites is also presented to illustrate how the temporo-spatial evolution of the Permian Tarim magmatism respond to the plume-lithosphere interaction (Fig. 12 C-F).

At ~300 Ma, a mantle plume impinged at the base of the thick lithosphere beneath the Tarim craton and may have resulted in either of two differing scenarios: 1) The plume did not melt directly because of a “lid effect” (Xu et al., 2014), however an increase of temperature due to conductive heating from the mantle plume triggered partial melting of metasomatized materials at the base of thick lithosphere, producing the diamondiferous kimberlites (Fig. 12C; Xu et al., 2014); 2) The mantle plume materials might have ponded beneath thinner lithosphere at the Tarim Craton-Tianshan Orogen boundary due to deflection by the thick keel of the Tarim craton. At these thin-spots/weak zones (< 120 km), plume materials underwent partial melting and produced the V-Ti magnetite bearing gabbros (Fig. 12C, He et al., 2016).

At ~293 and ~287 Ma, the Tarim lithosphere, especially thin-spots/weak zones near the craton-orogen boundary, were significantly thinned (~ 60 km) by thermomechanical erosion of the mantle plume (Fig. 12D and E). Given the compositional contrasts between the low Nb-Ta rhyolites and the basalts, the mantle plume might have only provided the heat source for the rhyolites. At the crust-mantle boundary of ~30 km, the steady-state geotherm is ~300 oC below the crust solidus (Fig. 12A). Conductive heating resulting from an increasing temperature of ~300 oC at the crust-mantle boundary will raise the temperature of the crust to the biotite breakdown condition in ~8 Myrs time interval (Fig. 12B). However, only after ~10 Myrs could conductive heating lead to amphibole-bearing crustal materials melting (Fig. 12B). This explains why fusible components will melt at an early stage and produce low temperature rhyolites, while refractory components melt at a later stage and produce high temperature rhyolites.

After ~287 Ma, it becomes increasingly difficult for the crust to melt and generate low temperature rhyolites because: 1) Fusible components might have been eliminated by previous thermal events, and 2) The thermal impact of mantle plume on the lithosphere might have decayed (Campbell and Griffiths, 1990). However, the lithosphere might have been further thinned by persistent thermomechanical erosion of the mantle plume, which would permit conductive heat transfer to a shallower depth in the crust and result in partial melting of fusible materials, producing low temperature rhyolites such as those low Nb-Ta rhyolites in the northern Tarim basin (Fig. 12F).

**6. Conclusions**

1) Zircon CA-ID-TIMS and LA-ICPMS U-Pb dating indicate that the Kuchehe and Dawanqi rhyolites were erupted at 292.6 ± 0.5 Ma and 274 ± 2 Ma respectively. The Kuchehe and Dawanqi rhyolites are characterized by negative Nb-Ta anomalies, crust-like trace element ratios and negative εNd-Hf(t) values, consistent with a crustal origin.

2) There is a temporal changes in zircon saturation temperature for the Permian Tarim low Nb-Ta rhyolites with an increasing trend of TZr (794-827 oC to 886-936 oC) in the early-to-middle stage (293-287 Ma) but a decreasing trend of TZr (886-936 oC to 803-908 oC) in the middle-to-late stage (287-274 Ma) of the Tarim magmatism.

3) The temporal changes in TZr for the Tarim rhyolites reflect that, above a mantle plume, fusible materials in the crust melt at an early stage and refractory materials melt at a later stage. The gradual decrease in TZr, however, is interpreted to reflect the thermal ‘decay’ of the mantle plume’s impact on the lithosphere. This interpretation on crustal melting behavior is applicable to data from the Chon Aike silicic LIP of Patagonia and Snake River Plateau-Yellowstone Plateau volcanic provinces, which show similar temporal variations in zircon saturation temperature.

**Appendix A. Supplementary data**

Supplementary data to this article can be found online at: xxx

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**Figure Captions:**

Figure 1. Simplified geological map showing distribution of Permian magmas in the Tarim Large Igneous Province (B, after Liu et al., 2014). Regional map showing distribution of Permian rhyolites in the northern part of the Tarim craton (C, modified after Gao et al., 2011). Zircon/Baddeleyite U-Pb ages showing the eruption sequence of Tarim LIP magmas in the northern Tarim and in the neighboring Tianshan Orogen (He et al., 2016; Zhang et al., 2010; Zhang et al., 2013; and this study).

Figure 2. A and B: Microphotography of biotite and garnet as phenocrysts of the Kuchehe rhyolites; C and D: Microphotography of the Dawanqi rhyolites.

Figure 3. Concordia plots of zircon CA-ID-TIMS U-Pb results for the Kuchehe rhyolites (A) and zircon LA-ICPMS U-Pb results for the Dawanqi ignimbrites (B). Insets are cathodoluminescence (CL) images showing their magmatic texture and spots of in-situ Hf (yellow cycles) and U-Pb (black cycle) isotopic analyses. Shaded point is excluded from age calculation. More CL images could be found in Fig. A.1.

Figure 4. Alkalis vs. silica contents for rhyolites from the Tarim LIP, Chon Aike silicic LIP and SRP-YP volcanic provinces. Data source: The low Nb-Ta rhyolites are from drilling holes (Ha: Halahatang, Liu et al., 2014, Tian et al., 2010; S: other low Nb-Ta rhyolites, Yu et al., 2011) in Tarim basin and Wenquan section (Liu et al., 2014); Chon Aike silicic LIP (Chon Aike SLIP), Pankhurst and Rapela (1995), Riley et al. (2001) and their references, Wever and Storey (1992); SRP-YP, Bonnichsen et al. (2008), Coble and Mahood (2016), Drew et al. (2013), Perkins et al. (1995; 1998), Watts et al. (2011), and Georoc (based on search in April 2016 at <http://georoc.mpch-mainz.gwdg.de/georoc/>).

Figure 5. Chemical discrimination diagrams for rhyolites from the Tarim LIP. A: A/NK vs. A/CNK diagram showing that the Tarim rhyolites are metaluminous to peraluminous. B: (Na2O+K2O)/CaO vs. (Zr+Nb+Ce+Y) plots showing the characteristics of I- and S-type granitoids for the Kuchehe rhyolites and aluminous A-type granitoids for the other low Nb-Ta rhyolites. C: (Zr+Nb+Ce+Y) vs. 10000Ga/Al plots showing the characteristics of I- & S-type granitoids for the Kuchehe rhyolites and aluminous A-type granitoids for the other low Nb-Ta rhyolites. Abbreviations: FG, fractionated granites; OGT, unfractionated granites. Data source are the same as in Fig. 4.

Figure 6. Chondrite-normalized rare earth element patterns (left panel) and primitive mantle-normalized multi-trace element diagram (right panel) for the Kuchehe and Dawanqi rhyolites compared to the Tarim mafic-ultramafic rocks (grey fields: Tian et al., 2010; Wei et al., 2014; Yang et al., 2007; Zhang et al., 2010; Zhou et al., 2009) and high Nb-Ta rhyolites (blue lines: Liu et al., 2014; Tian et al., 2010). Data source for the low Nb-Ta rhyolites are the same as in Fig. 4. Normalization values and OIB are from Sun and McDonough (1989). Bulk continental crust is from Rudnick and Gao (2003).

Figure 7. Plots of initial εNd-Hf(t) value vs. emplacement age for the Permian Tarim rhyolites. Data source for Nd isotope composition of the Tarim mafic-ultramafic rocks and the low Nb-Ta rhyolites are the same as in Fig. 4 and 6. Hf isotope composition for the Tarim mafic-ultramafic rocks are from Li et al. (2012).

Figure 8. Plots of zircon saturation temperature (TZr) showing temporal variation for rhyolites from the Tarim LIP (A), Chon Aike silicic LIP (B, Chon Aike SLIP) and SRP-YP volcanic provinces (C) vs. emplacement age. Lines showing temporal variation in TZr. Trend lines are drawn based on the highest TZr results at given ages because they represent the closest estimates of melting temperature at which the crustal source could achieve. A: Grey histogram is count of reliable ages for mafic-ultramafic rocks in the Tarim LIP (Xu et al., 2014). B: Ibanez silicic rocks fail to exhibit temporal variation due to insufficient geochronologic constraints and geochemical data. C: Three cycles for the SRP-YP are defined based on temporal variation in TZr. Data source for the low Nb-Ta rhyolites are the same as in captions of Figs. 4. TZr results for the Brennecke and Mount Poster samples, and peralkaline samples from the SRP-YP are abandoned (see text for details). TFe-Ti denote magma temperature constrained by Fe-Ti oxides thermometry (Bonnichsen et al., 2008; Cathey and Nash, 2004; Hildreth et al., 1984; Vazquez et al., 2009). Abbreviations: OV represents rhyolites from Virgin Valley, Paradise Valley, Huntington Creek, Antonne Wash, Overton Wash and Peacock in Perkins et al. (1998), B-J (Bruneau-Jarbidge), MBH (Mount Bennett Hills), TF (Twin Falls), HRCC (High Rock Caldera complex).

Figure 9. Variation of K2O, Na2O, CaO, Zr, A/CNK (molar Al2O3/[CaO+Na2O+K2O]) and M (molar [Na+K+2·Ca]/[Al·Si]) versus LOI. Data source for the low Nb-Ta rhyolites are the same as in Fig. 4.

Figure 10. Nb versus SiO2 diagram. Data source for the low Nb-Ta rhyolites are the same as in Fig. 4.

Figure 11. Nb versus Zr (A), Nb/La versus Th/La diagrams (B), and εNd(t) versus SiO2 (wt.%) (C). Data of OIB and PM (primitive mantle) are from Sun and McDonough (1989); Data of lower crust (LC), middle crust (MC) and upper crust (UC) are from Rudnick and Gao (2003). Horizontal and oblique black arrows (B) denote middle-upper crustal contamination respectively. The field of the Tarim basement rocks are from Liu et al. (2014).

Figure 12. Left: Modelling of conductive heating showing dehydration reactions of micas start earlier than that of amphibole-bearing rocks (amphibolite) at the same crustal pressure (A and B). The modeling approach follows method of Davies (1994). A: The initial geotherm is calculated for a mantle potential temperature of 1300 oC using the method of Davies (1994). The change in temperature 2, 4, 6, 8, 10 and 12 Ma after the temperature at the base of the lithosphere was increased by 300 oC is also illustrated. Thermal and rheological parameters: upper mantle density (ρ), 3300kg/m3; surface density contrast (Δρ), 3300kg/m3; specific heat, 1000 J/kgK; radiogenic heat productivity (μ), 3.0×10-6 W/m3; thermal diffusivity (κ), 1.0×10-6 m2/s; effective viscosity (ŋ), 1.0×1023 pa·s. Water-saturated solidi of granite, tonalite and basalt are after Stern et al. (1975) and Yoder (1976). B: Approximate ranges of dehydration of micas and amphibolite at crustal depths are after Miller et al. (2003) and Wyllie and Wolf (1993). Right: Schematic illustration of the role of mantle plume-lithosphere interaction in the generation of the Permian Tarim magmas (C-F). C: At ~300 Ma, as a result of conductive heating from the mantle plume, metasomatized enriched material at the base of the thick lithosphere beneath Tarim started to melt, producing the diamondiferous kimberlites. At the same time, mantle plume material started to melt at thin-spots of the Tarim craton-Tianshan Orogen boundary and resulted in the emplacement of V-Ti magnetite bearing gabbros. D: At ~293 Ma, the Tarim lithosphere was significantly thinned by thermomechanical erosion and conductive heating triggered partial melting of fusible materials in the crust. E: At ~287 Ma, persistent conductive heating triggered partial melting of refractory materials in the crust. F: After ~287 Ma, the lithosphere was thinned through persistent thermal erosion by the hot mantle plume. However, only fusible materials in the crust could melt as the thermal decay of the mantle plume. The Tarim basalts are also presented in (D) and (E) for illustration. Note that the Tarim basalts were actually erupted at ~291-287 Ma (Tian et al., 2010; Wei et al., 2014; Yang et al., 2006).