Postcollisional potassic and ultrapotassic rocks in southern Tibet: Mantle and crustal origins in response to India–Asia collision and convergence

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Available online 8 April 2014

Abstract

Postcollisional potassium-rich magmatism in southern Tibet provides an important insight into the deep processes inducing accelerated plateau uplift and associated geologic episodes during early Miocene. As a major outcropping of postcollisional magmatic rocks in the southwestern Tibet, ultrapotassic and potassic volcanism in Xungba basin occurred coevally at ~23 Ma, consisting of bimodal-like volcanic sequence. The mantle-derived ultrapotassic rocks (Group 1) are latites and marked by both mantle and crustal geochemical signatures, similar to other younger (19–10 Ma) ultrapotassic rocks elsewhere in the southern Tibet. The high Th/Yb ratios, low Ba/La and Hf/Sm ratios observed in ultrapotassic rocks strongly imply derivation from a metasomatized lithospheric mantle regions enriched by inputs of pelagic sediment and carbonate during previous Tethyan oceanic subduction, while their relatively high SiO₂ and Dy/Yb, low Ni/MgO and CaO/Al₂O₃, and convex upward ⁸⁷Sr/⁸⁶Sr–δ¹⁸Oᵥ−SMOW relationship suggest additional contributions of overthickened lower crust and ancient basement except for enriched mantle sources. The potassic rocks (Group 2), which underlie Group 1 rocks, are intermediate to silicic and exhibit adakitic geochemical affinities with high Sr/Y and La/Yb ratios, and low Y and Yb concentrations. The fingerprint of overthickened lower crust identified both from ultrapotassic and potassic rocks may be an important feature for magmatism occurred under the background of continental collision. And the bimodal volcanic sequence of Xungba postcollisional magmatism may further corroborate that the removal of lower part of over-thickened lithospheric mantle at depth have triggered an extension setting associated with Miocene plateau uplift.

1. INTRODUCTION

Following the closure of the Neo-Tethys ocean in the early-Paleogene (Mo et al., 2007, 2008; Royden et al., 2008; Lee et al., 2009; Bouilhol et al., 2013), continued convergence of India–Asia has lead to large-scale mantle–crust
interactions that may have been responsible for a variety of geological phenomena and the uplift of Tibetan Plateau in the Miocene (DePaolo and Ingrahm, 1985; Harrison et al., 1992; Yin and Harrison, 2000; Guo et al., 2008; Kapp et al., 2008; Molnar and Stock, 2009; Enkelmann et al., 2011; Kirstein, 2011). The postcollisional magmatism such as the late Oligocene–Miocene adakitic rocks of deep crustal origin and the ultrapotassic rocks of mantle origin offer unprecedented opportunities to reveal deep processes (e.g., deep subduction of Indian continental materials, convective thinning of the lithosphere, and delamination of overthickened lower crust) associated with the India–Asia continental collision and continued convergence (Miller et al., 1999; Ding et al., 2003; Nomade et al., 2004; Williams et al., 2004; Chung et al., 2005; Zhao et al., 2009; Guo et al., 2013; Liu et al., 2014). An important observation concerns the geochemical signatures of continental crust displayed by mantle-derived ultrapotassic rocks. Such signatures have long been interpreted as resulting from mantle sources enriched by deep processes, including mantle metasomatism by oceanic sediments and/or Indian continental crust (Miller et al., 1999; Ding et al., 2003; Williams et al., 2004; Zhao et al., 2009; Guo et al., 2013), which then led to the tectonic interpretation that the ultrapotassic magmatism marks the removal and melting of southern Tibet continental mantle lithosphere with metasomatic histories.

However, whether the metasomatic agents are subducted seafloor sediments and/or Indian continental crustal remains in dispute (Williams et al., 2004; Gao et al., 2007; Zhao et al., 2009; Tommasini et al., 2010; Guo et al., 2013). In the Lhasa terrane, the Mesozoic–early Cenozoic magmatism resulting from the Tethyan seafloor subduction would have led to the Tibetan mantle lithosphere metasomatized with either or both of the aforementioned agents before India–Asia continental collision (DeBon et al., 1986; Dewey et al., 1988; Chung et al., 2005; Chu et al., 2006; Gao et al., 2007; Ji et al., 2009; Tommasini et al., 2010; Zhu et al., 2011a, 2013; Liu et al., 2014). Geophysical studies show that the north-dipping Indian lithosphere has reached the the Bangong-Nujiang suture zone at depth (Kosarev et al., 1999; Nábelek et al., 2009), which supports the the scenario of Indian crust material as a possible metasomatic agent (Zhao et al., 2009; Liu et al., 2011a; Guo et al., 2013). Importantly, recent studies have found crustal xenoliths and a variety of zircon xenocrysts with pre-eruptive U–Pb ages in the ultrapotassic rocks (Sun et al., 2008; Hébert et al., 2014; Liu et al., 2013, 2014), providing evidence for the Lhasa terrane crust contributions to the Tibetan ultrapotassic magmatism. Previous studies proposed that an over-thickened crust may have formed as a result of continuous continental convergence (Chung et al., 2005, 2009) and the mature Precambrian basement of the central Lhasa subterrane shares similar isotopic composition to that of the Indian crust (Zhu et al., 2011b, 2012a; Zhang et al., 2012). All these observations point to the significance of the Lhasa terrane crustal material (i.e., in addition to Tethyan seafloor sediments and underthrust Indian basement) in the petrogenesis of the postcollisional potassic–ultrapotassic magmatism in southern Tibet.

In this paper, we present the results of our comprehensive study of postcollisional magmatism from the Xungba Basin (Fig. 1), which forms a “bimodal” volcanic sequence composed of early-Miocene ultrapotassic lavas and the underlying potassic lavas with adakitic geochemical signature (Liu et al., 2011b). The results include high quality zircon U–Pb age data and Hf isotopes, whole rock major and trace element compositions, and Sr–Nd–Pb–O isotopes. These data allow us to (i) clarify the relationship between ultrapotassic and adakitic magmatism and their constraints on the onset of the regional extension, (ii) place the timing on the lithospheric thinning/delamination in the Lhasa terrane, and (iii) probe the nature of the overthickened southern Tibetan lithosphere. Built on the previous studies, this study highlights the significance of crustal contributions (thickened lower crust and ancient Lhasa basement) in the petrogenesis of the Tibetan postcollisional magmatism, and provides enhanced understanding of crustal control on postcollisional magmatism in the context of India–Asia convergence.

2. GEOLOGICAL BACKGROUND

The Lhasa terrane, bounded by the Indus-Yarlung Zangbo suture zone (IYSZ) in the south and the Bangong-Nujiang suture zone (BNS) to the north, was subjected to northward Neo-Tethyan seafloor subduction prior to the India–Asia continental collision and postcollisional underthrusting of the Indian continent (Owens and Zandt, 1997; Yin and Harrison, 2000). Recent studies suggest that the Lhasa terrane can be divided into three sub-units (the northern, central, and southern Lhasa subterrane), separated by the Shiquanhe-Nam Tso mélangé zone (SNMZ) and Luobadui-Milashan fault (LMF), respectively (Zhu et al., 2011a, 2013) (Fig. 1a). These three units have significantly different sedimentary, magmatic and tectonic histories and appear to have distinctive crustal compositions.

In the northern Lhasa subterrane, the oldest sequences are middle Triassic sedimentary strata interbedded with volcanic rocks (Zhu et al., 2013). The Cretaceous magmatism, including the Duoni formation volcano-sedimentary strata (ca. 116–91 Ma; Kang et al., 2008) and Baingoin batholith (ca. 132–113 Ma; Zhu et al., 2013), is widely distributed in the northern Lhasa subterrane and is interpreted to have resulted from southward subduction of the Bangong-Nujiang Tethyan seafloor and continental collision thereafter (Zhu et al., 2011a, 2013). In the southern Lhasa subterrane, the Mesozoic sedimentary cover is restricted to the eastern portion and separated from the passive continental margin sedimentary sequence in the Tethyan Himalaya by the IYSZ (Zhu et al., 2013). The widespread Gangdese Batholith and syncollisional Linizzong volcanic succession are regarded as the product of northward subduction and subsequent break-off of the Yarlung Zangbo Tethyan oceanic slab (Coulon et al., 1986; Chung et al., 2005; Mo et al., 2007, 2008; Ji et al., 2009; Lee et al., 2009; Zhao et al., 2011; Niu et al., 2013). The Mesozoic–Cenozoic granitoid magmatism in both the northern and southern Lhasa subterrane is characterized by high zircon ε44(t) values, low radiogenic Sr isotope and unradiogenic
Potassic–ultrapotassic rocks in the central Lhasa subterrane include Shiquanhe (24–21 Ma; Williams et al., 2004), Xungba-Bangba-Yare (25–18 Ma; Miller et al., 1999; Liu et al., 2011b), Sailipu (17–16 Ma; Sun et al., 2008), and Yangying (11–10 Ma; Nomade et al., 2004). Data of adakitic rocks and leucogranite in the southern Lhasa subterrane and Tethyan Himalaya zones related to continental extrusion (Harrison et al., 1992). (c) Simplified geological map showing outcrops of postcollisional volcanic rocks and localities where volcanic rocks are dated in the Xungba basin (Miller et al., 1999; Liu et al., 2011b). Major outcrops of postcollisional potassic–ultrapotassic rocks in the central Lhasa subterrane include Shiquanhe (24–21 Ma; Williams et al., 2004), Xungba-Bangba-Yare (25–18 Ma; Miller et al., 1999; Chen et al., 2011; Liu et al., 2011b), Sailipu (17–16 Ma; Sun et al., 2008), and Yangying (11–10 Ma; Nomade et al., 2004). Data of adakitic rocks and leucogranite in the southern Lhasa subterrane and Tethyan Himalaya are from Guo et al. (2007), Chung et al. (2009), Zeng et al. (2011), Chen et al. (2011), Guan et al. (2012), Guo and Wilson (2012), and Hou et al. (2012). BNS = Bangong-Nujiang suture zone; SNMZ = Shiquan River-Nam Tso mélangé zone; LMF = Luobadui-Milashan fault; KF = Karakorum fault; IYXS = Indus-Yarlung Zangbo suture zone.

Nd isotope ratios largely due to the mantle input (Chung et al., 2009; Ji et al., 2009; Wu et al., 2010; Chu et al., 2011; Zhu et al., 2011a).

In contrast, the central Lhasa subterrane underlain by Archean to Proterozoic basement may represent a fragment of a micro-continent separated from east Gondwana that drifted across the Tethyan Ocean (Zhu et al., 2011b; Zhang et al., 2012). Although the outcrops of the Precambrian basement have only been identified in the eastern part of the central Lhasa subterrane to date (Nyainqentanglha Group, 780–690 Ma; Kapp et al., 2005; Dong et al., 2010), inherited Precambrian zircons have been identified in Mesozoic magmatic rocks from the western part of the central Lhasa subterrane (Zhu et al., 2011a, 2013). The sedimentary sequences in the central Lhasa subterrane, ranging from Ordovician to the early Mesozoic, are mainly composed of fine-grained marine clastic and carbonate rocks (Yin and Harrison, 2000). Late Jurassic to early Cretaceous volcano-sedimentary rocks of the Zenong formation, as well as age-equivalent granitoids, are abundant throughout the central Lhasa subterrane and have also been attributed to the subduction and slab break-off of the Bangong-Nujiang
oceanic lithosphere (Kang et al., 2008; Zhu et al., 2011a, 2013). Besides, the contribution from ancient basement to the Mesozoic–Cenozoic magmatism in the central Lhasa subterrane is evident because of the enriched Sr–Nd–Hf isotopic compositions (Zhao et al., 2011; Zhu et al., 2011a; Lee et al., 2012).

The postcollisional magmatism in the Lhasa terrane, including potassic-ultrapotassic volcanic rocks, adakitic intrusives, and minor leucogranite, is distributed roughly parallel to the sutures along the Lhasa terrane and Himalayan orogenic belt (Chung et al., 2005; Zhao et al., 2009) (Fig. 1a) and is spatially related to the N–S trending extensional faults (cf. Coulon et al., 1986; Zhao et al., 2006). The postcollisional magmatism in the central Lhasa subterrane extends to Shiquanhe to the west and Tangra Yum Co to the east, which consists of potassic volcanic rocks (PVR) and ultrapotassic volcanic rocks (UPVR). These potassium-rich rocks, with eruptive ages ranging from 27 to 12 Ma, show extremely high Sr but low Nd isotopic ratios (Miller et al., 1999; Ding et al., 2003; Williams et al., 2004; Zhao et al., 2006, 2009; Gao et al., 2007, 2009; Chen et al., 2010, 2012; Guo et al., 2013).

In the southern Lhasa subterrane, the postcollisional magmatism is dominated by adakitic intrusive rocks and minor PVR–UPVR that occurred as veins, volcanic and sub-volcanic rocks of 20–10 Ma age with a juvenile crustal signature (Coulon et al., 1986; Williams et al., 2001; Zhao et al., 2001; Hou et al., 2004; Chung et al., 2005, 2009). In the Himalayan orogenic belt, Miocene leucogranites formed by crustal anatexis display isotopic compositions that may represent subducted Indian basement (Harrison et al., 1997; Zeng et al., 2011; Guo and Wilson, 2012; Hou et al., 2012). Near the eastern syntaxis, Eocene–Oligocene high Sr/Y rocks have been identified in the southern Lhasa subterrane (26–38 Ma; Harrison et al., 2000; Hou et al., 2004; Guo et al., 2007; Chung et al., 2009; Guan et al., 2012) and the Tethyan Himalaya (43–44 Ma; Aikman et al., 2008, 2012; Zeng et al., 2011; Hou et al., 2012).

3. LOCATION AND PETROGRAPHY OF SAMPLES

The Xungba basin, situated in the northern part of the N–S trending Xungba-Yare rift, contains the largest known outcrops of postcollisional magmatic rocks in the central Lhasa subterrane (Fig. 1a). Folded Mesozoic strata, which mainly consist of marine facies sedimentary rocks interbedded within volcanic flows, cover the northern part of the basin. The subduction-related Zenong group volcano-sedimentary strata are restricted to the southern part of the basin (Fig. 1c). According to the composition and abundances of phenocrysts (brief description has been summarized in Supplementary Online Material (SOM) Table 1) and stratigraphic contact relationships (Fig. 2), the Xungba postcollisional volcanic rocks can be subdivided into ultrapotassic latite (Group 1; UPVR) and potassic intermediate-silicic rocks (Group 2; PVR).

The latitic upper portion (Group 1) of the Xungba lava flows is mainly found in southern Xungba basin, in clear unconformable contact with underlying carbonate strata (Figs. 2 and 3a) and contains crustal xenoliths (Fig. 3b). They have porphyritic and synneusis textures (Fig. 3c) with 10–15 vol.% in phenocrysts (clinopyroxene, orthopyroxene, olivine, plagioclase, and phlogopite). Zoning and pervasive resorption textures are very common for pyroxenes from the UPVR (Fig. 3d and e), which exhibits irregular dark finely spongy cellular structure in the center (Fig. 3d) or coarsely cellular structure to the rim (Fig. 3e). In a latite sample, subhedral rutile xenocrysts were also found (Fig. 3g). Compared with the phenocrysts in lattites (Group 1), the potassic intermediate-silicic rocks (Group 2), which occurred as the lower part of the Xungba lava flows (Fig. 2), contain more sanidine and plagioclase (Fig. 3h and i) with minor biotite, hornblende, and orthopyroxene.

In addition to the field superposition relationship between UPVR and PVR (lithostratigraphic columns C in Fig 2), their similar eruptive ages, which are consistent with previously published data (Table 1), further suggest that these postcollisional rocks formed as the result of the same tectono-magmatic event. The early-Miocene ages indicate that they are the earliest postcollisional magmatic rocks in southern Tibet (Fig. 1a) (cf. Chung et al., 2005; Zhao et al., 2006, 2009).

4. ANALYTICAL METHODS

4.1. Zircon U–Pb dating, trace element and Hf isotopes

Zircons were extracted from coarsely crushed samples (80 mesh). They were separated by using combined methods of heavy liquid and magnetic separation before hand-picking under a binocular. The selected zircons were mounted in epoxy resins and were polished to expose grain interiors. Cathodoluminescence (CL) images were obtained by using scanning electron microscope (SEM, Leo 1450VP, Germany) at the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGG–CAS) before in situ U–Pb and Hf isotopes analyses.

Zircon U–Pb dating and trace element analysis were acquired simultaneously using LA-ICP-MS at the State Key Laboratory of Geological Processes and Mineral Resources (GPMR), China University of Geosciences, Wuhan. Laser sampling was performed using 193 nm excimer laser ablation system (GeoLas 2005), and ion intensities were acquired using ICP-MS (Agilent 7500a). The diameter of spot was ~32 μm. Zircon standard 91500 was used as external standard for U–Pb dating (Wiedenbeck et al., 1995), and every five sample analyses was followed by analysis of two 91500 zircon standards. Each analysis includes a background acquisition of 20–30 s (gas blank) and 50 s data acquisition on the sample. Trace element abundances were calibrated against USGS multiple-reference materials (BCR-2G and BIR-1G) using 29Si as an internal standard (Liu et al., 2010). Off-line selection and integration of background and sample signals, time-dependent drifts for U–Th–Pb isotopic ratios correction, U–Pb dating and quantitative calibration for zircon trace element analyses were all performed by ICPMSDataCal_ver8.0 (Liu et al., 2010). Detailed operating conditions for the
LA-ICP-MS and data reduction are the same as those described in Liu et al. (2008, 2010). Common lead correction procedure was based on Andersen (2002). Tera–Wasserburg diagrams and mean square of weighted deviates (MSWD) were calculated using Isoplot/Ex_ver3 (Ludwig, 2003).

Zircon Hf isotope were analyzed using a Nu Plasma HR MC-ICPMS (Nu Instruments Ltd., UK) coupled to a GeoLas 2005 excimer ArF laser-ablation system at the State Key Laboratory of Continental Dynamics, Northwest University (CDNU), Xi’an. The energy, pulse rate, and

Fig. 2. Lithostratigraphic columns of Xungba postcollisional volcanic rocks. The locations of these columns are indicated in the Fig. 1c. Bi: biotite, Cpx: clinopyroxene, Hbl: hornblende, Ol: olivine, Opx: orthopyroxene, Phl: phlogopite, Pl: plagioclase, Qtz: quartz, and Sa: sanidine.

Fig. 3. (a) The unconformable contact between early-Miocene ultrapotassic rocks and underlying early Cretaceous limestone and clastic rocks. (b) Crustal xenolith entrained by ultrapotassic magma. (c) Phenocrystal clots composed of olivine with rounded or triangular shapes and subhedral clinopyroxene in the UPVR (Group 1). (d) Subhedral orthopyroxene and clinopyroxene phenocrysts of ultrapotassic latite (Group 1). (e and f) Poikilitic olivine textures in the UPVR (Group 1): (e) orthopyroxene with coarsely cellular texture as a rim of irregular-shaped olivine grain, and (f) elliptical olivine grains included in clinopyroxene. (g) Rutile xenocryst in the UPVR (Group 1). (h) Sanidine phenocrystal clots in the PVR (Group 2). (i) Plagioclase phenocrystal clots with polysynthetic twinning in the PVR (Group 2).
207Pb/208Pb/86Sr/146Nd/206Pb/87Sr/208Pb/143Nd/

Table 1

<table>
<thead>
<tr>
<th>Affinity</th>
<th>Sample No.</th>
<th>Location</th>
<th>Dating results</th>
</tr>
</thead>
<tbody>
<tr>
<td>UPVR</td>
<td>TE011/93</td>
<td>~10 km South of Xungba</td>
<td>Minerals (Ar–Ar MS)</td>
</tr>
<tr>
<td>UPVR</td>
<td>TE138/93</td>
<td>~15 km South of Xungba</td>
<td>Minerals (Ar–Ar MS)</td>
</tr>
<tr>
<td>UPVR</td>
<td>10YR06</td>
<td>~20 km Southeast of Xungba</td>
<td>Zircon U–Pb</td>
</tr>
<tr>
<td>UPVR</td>
<td>08YR05</td>
<td>~25 km south of Xungba</td>
<td>Zircon U–Pb</td>
</tr>
<tr>
<td>PVR</td>
<td>TE025/93</td>
<td>~25 km South of Xungba</td>
<td>Minerals (Ar–Ar MS)</td>
</tr>
<tr>
<td>PVR</td>
<td>TE148/93</td>
<td>~25 km South of Banga</td>
<td>Minerals (Ar–Ar MS)</td>
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<tr>
<td>PVR</td>
<td>TE150/93</td>
<td>~25 km South of Banga</td>
<td>Minerals (Ar–Ar MS)</td>
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<tr>
<td>PVR</td>
<td>07BB03</td>
<td>~50 km South of Banga</td>
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</tr>
<tr>
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<td>10XB06</td>
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<td>Zircon U–Pb</td>
</tr>
<tr>
<td>PVR</td>
<td>GJ0602</td>
<td>~28 km East of Xungba</td>
<td>Zircon U–Pb</td>
</tr>
<tr>
<td>PVR</td>
<td>GJ0619</td>
<td>~3 km Northwest of Xungba</td>
<td>Zircon U–Pb</td>
</tr>
<tr>
<td>PVR</td>
<td>GJ0628</td>
<td>~15 km Southeast of Banga</td>
<td>Zircon U–Pb</td>
</tr>
<tr>
<td>PVR</td>
<td>10XB01</td>
<td>~15 km West of Xungba</td>
<td>Zircon U–Pb</td>
</tr>
</tbody>
</table>

UPVR, ultrapotassic volcanic rock; PVR, potassic volcanic rock.

spot size in this work were 100 mJ, 8 Hz, and 44 μm, respectively. High-purity helium was used as the carrier gas within the ablation cell and was mixed with argon (makeup gas) in the homogenizers. Each analysis includes 30 s data acquisition on the sample, and every ten-sample analysis was followed by one analysis of zircon standard 91500, GJ-1 and Monastery. Detailed operating conditions for MC-ICP-MS and the correction for interference by Lu and Yb have been described by Yuan et al. (2008). In this work, the average 176Hf/177Hf ratio for zircon standard 91500 was 0.282299 ± 15 (2σ, n = 38), which agrees well with the recommended value within error (Woodhead et al., 2004; Yuan et al., 2008).

4.2. Major and trace elements

For whole-rock geochemical analysis, well-cleaned fresh rock sample chips (dried at 50 °C) were crushed to 200 mesh after rejecting amygdale and xenoliths (crust and mantle) by hand-picking. Rock powders were mixed with 5.2 g Li2B4O7 + 0.4 g LiF + 0.3 g NH4NO3, and fused in a Pt crucible for major elements analysis using X-ray fluorescence spectrometry (XRF) at CDNU Xi’an. Trace element abundances were determined using ICP-MS (Agilent 7500a) at GPMR, Wuhan. Whole-rock powders (50 mg) for trace elements analysis were dissolved in Teflon bombs by purified HF + HNO3, and the sealed Teflon bombs have been heated at 190 °C in oven for 48 h. After evaporating the solution in Teflon to incipient dryness at 115 °C, the resultant salt was re-dissolved by mixture of 30% HNO3. The final solution was diluted to 100 g with 2% HNO3, and fused in a Pt crucible for major elements analysis using X-ray fluorescence spectrometry (XRF) at CDNU Xi’an. Trace element abundances were determined using ICP-MS (Agilent 7500a) at GPMR, Wuhan. Whole-rock powders (50 mg) for trace elements analysis were dissolved in Teflon bombs by purified HF + HNO3, and the sealed Teflon bombs have been heated at 190 °C in oven for 48 h. After evaporating the solution in Teflon to incipient dryness at 115 °C, the resultant salt was re-dissolved by mixture of 30% HNO3. The final solution was diluted to ~100 g with 2% HNO3 in a polyethylene bottle. Detailed sample-preparing procedure, operating conditions for the laser ablation system and ICP-MS instrument, USGS rock reference materials (BCR-2, BHVO-2, AGV-2, RGM-1 and GSR-1) and data reduction have been described by Liu et al. (2008).

4.3. Whole-rock Sr–Nd–Pb–O isotopes

For whole-rock Sr–Nd isotopic analysis, rock powders were dissolved in Teflon bombs using HF + HNO3 + HClO4. Rb, Sr, Sm and Nd were separated by conventional ion exchange procedures and measured using a Finnigan MAT261 thermal ionization mass spectrometer (TIMS) in static mode at GPMR, Wuhan. The isotopic ratios were normalized to 86Sr/88Sr = 0.1194 and 144Nd/142Nd = 0.7219, respectively, for mass fractionation correction. Total procedural blanks are <1 ng for Sr and <50 pg for Nd. The Sr standard (NBS 607) and Nd standard (BCR-2) yielded 87Sr/86Sr = 0.710257 ± 12 (2σ) and 144Nd/142Nd = 0.512118 ± 12 (2σ), respectively. Detailed descriptions of the analytical procedures and operating conditions for TIMS are given in Liu et al. (2004). The εNd(t) was calculated with reference to CHUR 143Nd/144Nd ratios of 0.512638, and the neodymium isotopic depleted mantle model age (TDM) was calculated by assuming its parental magma had been derived from depleted mantle source with (143Sm/144Nd)DMM = 0.21375 and (143Nd/144Nd)DMM = 0.51315. The decay constant of Nd is 6.54 × 10−12 yr−1.

For Pb isotopic analysis, rock powders (100–150 mg) were dissolved in Teflon bombs by purified HF + HNO3 and were heated at 195 °C for 48 h. Separation of Pb was achieved by conventional ion exchange procedures and HCl was used as leaching agent. Pb isotopic ratios were determined using Nu Plasma HR MC-ICPMS at the CDNU, Xi’an and Thermo-Finnigan TRITON mass spectrometry equipped with an oxygen gas leak valve, nine Faraday cups and an ion counting multiplier at the Institute of Geology and Mineral Resources, Tianjin. In Xi’an, total procedural blanks were <20 pg. For mass fractionation correction, NBS 981 standards was analyzed in Xi’an with average values and associated errors (2σ) being 206Pb/204Pb = 16.942 ± 0.002, 207Pb/204Pb = 15.499 ± 0.002, and 208Pb/204Pb = 36.72 5 ± 0.004. In Tianjin, total procedural blanks were <1 ng, and NBS 982 standard was used to for mass fractionation. The average values of NBS 982 in this study were: 206Pb/204Pb = 36.624 ± 0.002, 207Pb/204Pb = 17.078 ± 0.001, and 208Pb/204Pb = 36.508 ± 0.002.

Oxygen isotopes were measured on a Finnigan MAT-253EM mass spectrometry at the Institute of Mineral
Resources, Chinese Academy of Geological Sciences. The mixture of purified BrF$_5$ and 9.5 mg sample powder was heated at 700 °C for 9 h under high vacuum condition. For oxygen isotopic analysis, the resultant oxygen was converted to CO$_2$ by reacting with graphite resistance rod for 12 min. Every five sample analysis was followed by one analysis of quartz standard NBS 28 ($^{18}O_{SMOW}$ = 10.92‰, and the analytical uncertainty is better than 0.2‰). All δ$_{18}O$ values were reported with reference to the V-SMOW standard.

Zircon U–Pb age data, trace element (sample 08YR05) and Hf isotopic data are given in SOM Tables 2–4, respectively. Major and trace element data are given in Table 2, and whole-rock Sr–Nd–Pb–O isotopic compositions are given in Table 3.

5. RESULTS

5.1. Zircon morphology, trace elements and U–Pb geochronology

In CL images (Fig. 4), zircons from Group 1 UPVR samples (08YR05) are subrounded, and exhibit weak zoning, unzoned or uniform internal textures (Fig. 4a), which may imply a high-temperature crystallization environment (Corfu et al., 2003). In contrast to zircons from the PVR, zircons from the UPVR have relatively lower Th (94–468 ppm) and U (296–2420 ppm), with Th/U ratios >0.1 (Fig. 5a). Zircons from the UPVR can be subdivided into two types according to their different heavy rare earth element (HREE) patterns (Fig. 5b). The first type shows a rising slope from Ho to Lu ([Dy/Yb]$^N$ = 0.20–0.43) (Fig. 5b), together with homogeneous internal structure, consistent with magmatic zircons with mantle affinity (Belousova et al., 2002; Hoskin and Schaltegger, 2003), albeit with relatively higher U and Th concentrations (SOM Table S2). The second type is marked by generally flat HREE patterns ([Dy/Yb]$^N$ = 0.81–1.34) (Fig. 5b), which are similar to zircons crystallized under a very limited HREE supply (Rubatto, 2002; Liu et al., 2013, 2014). In spite of the divergence between two kinds of zircon, the U–Pb dating of UPVR zircons yield a weighted mean $^{206}$Pb/$^{238}$U ages of 23.2 ± 0.4 Ma, after rejecting discordant ages (Fig. 4e).

Zircons from three PVR samples (GJ0602, GJ0619 and GJ0628) of Group 2 rocks commonly occur as doubly-terminated prismatic crystals, ranging from about 100 to 250 μm in size (Fig. 4b–d). Most zircons have narrow and frequent oscillatory zoning, and only a few of them show broad zoning. Except for a zircon grains (GJ0628-03) having extremely high Th concentration (19278 ppm) and relatively low Th/U ratios (0.23), most zircons in the PVR display higher Th (397–6115 ppm) and U (475–3737 ppm) concentrations than zircons in the UPVR (Fig. 5a). Their Th/U ratios (0.24–2.66) indicate a magmatic zircon origin (Hoskin and Schaltegger, 2003). After rejecting inherited zircon analytical spots, zircons from these PVR samples yield $^{206}$Pb/$^{238}$U mean ages of 23.5 ± 0.2, 23.3 ± 0.2 and 23.4 ± 0.1 Ma, respectively (Fig. 4f–h).

5.2. Zircon Hf isotopes

A total number of 58 dated zircons were analyzed by LA-MC-ICPMS for Hf isotope (SOM Table 4). Zircons from Group 1 UPVR sample (08YR05) exhibit highly variable $^{176}$Hf/$^{177}$Hf ratios (0.282173–0.282868), $^{176}$Hf(t) values (−21.2 to −3.0) and depleted mantle model ages (TDM = 0.8–1.5 Ga). By contrast, zircons in two PVR samples (GJ0602 and GJ0619) of Group 2 show relatively uniform Hf isotopic composition ($^{176}$Hf/$^{177}$Hf = 0.282330–0.282427), with $^{176}$Hf(t) and TDM ranging from −15.6 to −12.2 and 1.9 to 2.1 Ga, respectively. Except for one analytical spot with most radiogenic Hf isotopic ratio ($^{176}$Hf/$^{177}$Hf = 0.282531, $^{176}$Hf(t) = −8.4, TDM = 1.6 Ga), most zircons in this PVR sample (GJ0628) have a narrow Hf isotopic range ($^{176}$Hf/$^{177}$Hf = 0.282342–0.282462; $^{176}$Hf(t) = −15.6 to −12.8; TDM = 1.8 to 2.0 Ga).

5.3. Major and trace element geochemistry

In this study, postcollisional volcanic rocks from the Xungba basin show high K$_2$O contents (4.36–9.00 wt.%) and K$_2$O/Na$_2$O (1.64–6.16) (Table 2) with a wide compositional range from latite to rhyolite (SiO$_2$ = 57.2–70.1 wt.%) (Fig. 6a). Group 1 rocks are relatively more Mg-rich (MgO = 4.57–6.75 wt.%, Mg$_{eq}$ = 66.0–69.5) and metaluminous (A/CNK = 0.63–0.68) (Fig. 5c), with high K$_2$O (5.69–8.62 wt.%) and K$_2$O/Na$_2$O (2.62–3.34) (Fig. 6b). By contrast, Group 2 rocks are intermediate-silicic, showing varying K$_2$O/Na$_2$O (1.64–6.16) but lower Mg$_{eq}$ (28.9–55.6), TFe$_2$O$_3$ (2.52–4.09 wt.%), TiO$_2$ (0.55–1.02 wt.%), and P$_2$O$_5$ (0.14–0.38 wt.%), and wider range of A/CNK (0.90–1.16) (Table 2).

Overall, the Xungba K-rich rocks display a significant light REE (LREE) enrichment (Fig. 7a). The UPVR (Group 1) show a moderate LREE fractionation ([La/Sm]$_N$ = 1.64–1.89), a negative Eu anomaly (Eu/Eu$^*$ = 0.64–0.72), a large (La/Yb)$_N$ variation (22–27), and lower total REE contents ($^{\sum}$REE = 360–459 ppm) (Table 2) (Fig. 7a), overlapping the field defined by postcollisional ultrapotassic rocks from the Lhasa terrane (cf. Zhao et al., 2009). The REE patterns of the PVR (Group 2) are comparable in light and middle REE patterns to Cambrian silicic lavas (Zhu et al., 2012a) and in HREEs ([La/Yb]$_N$ = 30–74) to S-type granites (Liu et al., 2006), both of which might originate from the Lhasa basement (Fig. 7a).

The most distinctive features of postcollisional Tibetan K-rich rocks are the pronounced enrichment in Th, U, K, and Pb, and relative depletion in Ba, Nb, Ta, Sr, and Ti (Fig 7b). Aside from these similarities, the UPVR are characterized by high (Th/U)$_N$ ratios (1.62–1.96), lower Hf/Sm ratios (0.55–0.62) (Table 2), more remarkably enrichment of K and weak depletion of Sr (Fig. 7b), falling into the field of postcollisional ultrapotassic magmatism in the Lhasa terrane (Fig 7b) (cf. Zhao et al., 2009). As for the Xungba PVR, they have relatively lower abundances of incompatible and compatible elements than the UPVR, with strong Nb and Ta depletions, higher Hf/Sm ratios.
### Whole-rock major and trace element compositions for postcollisional volcanic rocks from the Xungba Basin.

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**Major element (wt.%)**

- **SiO₂**: 63.1
- **TiO₂**: 1.01
- **Al₂O₃**: 15.2
- **Fe₂O₃**: 3.90
- **MnO**: 0.03
- **MgO**: 2.08
- **CaO**: 2.76
- **Na₂O**: 1.48
- **K₂O**: 8.68
- **P₂O₅**: 0.34
- **LOI**: 1.16
- **Total**: 99.8
- **K₂O/Na₂O**: 5.86
- **Mgₑ**: 51.4
- **A/CK**: 0.90
- **A/NK**: 1.29

**Trace element (ppm)**

- **Be**: 9.63
- **Sc**: 12.2
- **V**: 83.6
- **Cr**: 105
- **Co**: 9.28
- **Ni**: 18.2
- **Cu**: 18.5
- **Zn**: 83.8
- **Ga**: 26.6
- **Rb**: 457
- **Sr**: 491
- **Zr**: 563
- **Nb**: 31.9
- **Cs**: 121
- **Ba**: 1908
- **Hf**: 17.4
- **Ta**: 3.55
- **Nb**: 58.9
- **U**: 12.0
- **La**: 135
- **Ce**: 274
- **Pr**: 30.4
- **Nd**: 108
- **Sm**: 15.1
- **Eu**: 2.41
- **GdCORR**: 7.62
- **Tb**: 0.86
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(continued on next page)
### Table 2 (continued)

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**Trace element (ppm)**

- Be: 7.32, 7.06, 7.38
- Sc: 7.27, 7.37, 7.56
- V: 51.5, 54.1, 60.2
- Cr: 33.2, 35.6, 41.3
- Co: 5.33, 4.66, 6.62
- Ni: 13.6, 9.69, 10.2
- Cu: 9.82, 12.7, 12.2
- Zn: 84.8, 80.5, 84.0
- Ga: 29.2, 30.8, 29.5
- Rb: 337, 354, 341
- Sr: 167, 205, 192
- Zr: 266, 269, 266
- Nb: 11.5, 10.2, 9.85
- Cs: 19.0, 19.1, 19.6
- Ba: 638, 650, 566
- Hf: 7.41, 7.40, 7.59
- Ta: 0.86, 0.75, 0.72
- Pb: 38.3, 39.1, 38.5
- Th: 40.4, 42.9, 41.9
- U: 8.84, 7.40, 8.61
- La: 76.4, 87.1, 77.9
- Ce: 161, 170, 166
- Pr: 20.5, 23.3, 20.9
- Nd: 69.3, 78.6, 71.9
- Sm: 11.9, 13.3, 12.0
- Eu: 1.24, 1.28, 1.16
- GdCORR: 7.20, 7.41, 6.87
- Tb: 0.89, 0.89, 0.83
- Dy: 0.35, 0.35, 0.37
- Ho: 0.64, 0.55, 0.52
- Er: 1.81, 1.53, 1.47
- Tm: 0.24, 0.19, 0.19
- Yb: 1.40, 1.19, 1.14
- Tm: 0.24, 0.19, 0.19

**Notes:**

- SiO₂ (wt.%): 67.3, 68.1, 66.6
- Major element (wt.%): Mg#, Al₂O₃, MgO, SiO₂, CaO, Na₂O, K₂O, TiO₂, Fe₂O₃, FeO, MnO, MgO, FeO, CaO, MgO, Al₂O₃, SiO₂, K₂O, Na₂O
- A/CNK = Al₂O₃/(CaO + Na₂O + K₂O) (molar ratio)
- A/NK = Al₂O₃/(Na₂O + K₂O) (molar ratio)
- Mg²⁺ = Mg²⁺/(Mg²⁺ + Fe²⁺) (molar ratio)

**References:**

- Liu et al., 2014.
- Loss on ignition (LOI)
- Duplicate for showing analytical precision.

**TAS, total alkalics vs. silica; Tr., trachyte; TrD, trachydacite; Lat, latite; Dac, dacite; Rhy, rhyolite. LOI, loss on ignition.**
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<td>0.732739</td>
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<td>0.734000</td>
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<td>11.9</td>
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<td>12.0</td>
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<td>526.9</td>
<td>658.7</td>
<td>0.718822</td>
<td>4</td>
<td>0.7181</td>
<td>20.0</td>
<td>0.718822</td>
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<td>4</td>
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<td>23.8</td>
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<td>0.720312</td>
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<td>0.7188</td>
<td>21.3</td>
<td>0.719501</td>
<td>4</td>
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<td>21.3</td>
<td>0.719501</td>
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<td>450.3</td>
<td>636.6</td>
<td>0.718966</td>
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<td>18.6</td>
<td>0.718966</td>
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<td>0.7188</td>
<td>19.9</td>
<td>0.719615</td>
<td>5</td>
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<td>19.9</td>
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<td>0.7188</td>
<td>19.8</td>
<td>0.719596</td>
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<td>0.7188</td>
<td>19.8</td>
<td>0.719596</td>
<td>5</td>
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Table 4
Isotopic compositions of melts derived from three geochemistry end-members.

<table>
<thead>
<tr>
<th>Geochemistry end-members</th>
<th>SCLM</th>
<th>Hypothetical ancient lower crust</th>
<th>Ancient Lhasa basement</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rb (ppm)</td>
<td>26.3</td>
<td>-</td>
<td>136.6</td>
</tr>
<tr>
<td>Sr (ppm)</td>
<td>1027</td>
<td>300</td>
<td>77.2</td>
</tr>
<tr>
<td>(^{87}\text{Sr}/^{86}\text{Sr})</td>
<td>0.705</td>
<td>0.710</td>
<td>0.756</td>
</tr>
<tr>
<td>Sm (ppm)</td>
<td>7.06</td>
<td>-</td>
<td>5.76</td>
</tr>
<tr>
<td>Nd (ppm)</td>
<td>40.3</td>
<td>26.0</td>
<td>28.8</td>
</tr>
<tr>
<td>(^{143}\text{Nd}/^{144}\text{Nd})</td>
<td>0.5126</td>
<td>0.5115</td>
<td>0.5119</td>
</tr>
<tr>
<td>(^{6}^{18}\text{O}/^{16}\text{O}) (ppm)</td>
<td>5.3 ± 0.4</td>
<td>≤5.3</td>
<td>9.5–14</td>
</tr>
</tbody>
</table>

The concentrations of Rb, Sr, Sm, and Nd, together with Sr–Nd isotopic compositions for these three end-members, are averages based on Miller et al. (1999), Ding et al. (2003), and Zhu et al. (2012a). Pb concentrations and \(^{206}\text{Pb}/^{204}\text{Pb}\) ratios are from Zhao et al. (2009), which is consistent with derivation from a phlogopite-bearing mantle source. This interpretation is supported by the phlogopite-bearing spinel facies mantle xenoliths.

Table 5
Mineral/melt partition coefficients used in the trace element modeling.

<table>
<thead>
<tr>
<th>Phlogopite-spinel Iherzolite</th>
<th>Ol</th>
<th>Opx</th>
<th>Cpx</th>
<th>Phl</th>
<th>Spl</th>
</tr>
</thead>
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<tr>
<td>DLa</td>
<td>0.0000046</td>
<td>0.00057</td>
<td>0.0586</td>
<td>0.0004</td>
<td>-</td>
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<tr>
<td>DHy</td>
<td>0.0053</td>
<td>0.0365</td>
<td>0.5382</td>
<td>0.0182</td>
<td>-</td>
</tr>
<tr>
<td>DV</td>
<td>0.0214</td>
<td>0.2075</td>
<td>0.5819</td>
<td>0.0330</td>
<td>-</td>
</tr>
<tr>
<td>Rutile-bearing eclogite</td>
<td>Grt</td>
<td>Cpx</td>
<td>Rt</td>
<td></td>
<td></td>
</tr>
<tr>
<td>DLa</td>
<td>0.0027</td>
<td>0.0281</td>
<td>0.3000</td>
<td></td>
<td></td>
</tr>
<tr>
<td>DHy</td>
<td>2.6594</td>
<td>0.4588</td>
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<td></td>
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</tr>
<tr>
<td>DV</td>
<td>10.0544</td>
<td>0.5983</td>
<td>0.0120</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

These mineral-melt D values (partition coefficients) are averages of available data for different phases. (1) In spinel facies Iherzolite, the partition coefficients for Ol are from McDade et al. (2003) and Lee et al. (2007). D values of Opx and Phl are from Grégoire et al. (2000) and Sun and Liang (2012), respectively. (2) In rutile-bearing eclogite facies, partition coefficients for Grt are from Green et al. (2000) and Pertermann et al. (2004), and, for Cpx, D values are from Pertermann and Hirschmann (2002) and Pertermann et al. (2004). D values of Rt are based on Pilet et al. (2011).

Values in brackets denote the mineral percentages in the residue.
found in the UPVR in the Sailipu basin (Zhao et al., 2008; Liu et al., 2011a). The consistence between REE patterns of the UPVR and the model melts in equilibrium with clinopyroxenes in the mantle xenolith (Liu et al., 2011a) further corroborates that the ultrapotassic melt is originated from a phlogopite-clinopyroxene-olivine assemblage mantle source represented by the lherzolite/harzburgite mantle xenoliths (i.e., metasomatized lithospheric mantle origin). As the Xungba early-Miocene UPVR share the same geochemical characteristics mentioned above, it is reasonable to assume the same lithospheric mantle origin.

However, compared with basalts, the relatively high silica contents of the UPVR cannot be formed alone by partial melting of hydrous spinel-facies peridotites even if the mantle beneath southern Tibet had experienced protracted hydration during Tethyan oceanic subduction. Because partial melts produced in the hydrous spinel stability field (1100–1400 °C; 1.2–2.0 GPa) are still basaltic with only 1–3 wt.% SiO₂ increase even though the H₂O content can be up to 6.3 wt.% (cf. Gaetani and Grove, 1998).

Although ultrapotassic rocks can form in various tectonic settings (Müller et al., 1992), crustal material always plays an important role in ultrapotassic magma evolution (Avanzinelli et al., 2009; Conticelli et al., 2009a,b). In the Lhasa terrane, the UPVR commonly display continental-crust-like geochemical features (e.g., enriched Sr–Nd–Pb–O isotopes) (Miller et al., 1999; Williams et al., 2004; Zhao et al., 2009). The extensive fractional crystallization

Fig. 4. (a–d) Cathodoluminescence images, and (e–h) Tera–Wasserburg diagrams for zircons from Xungba potassic and ultrapotassic rocks.
can be ruled out as the major cause of SiO₂ elevation because the steep positive La/Yb–La correlation is consistent with varying extent of partial melting rather than fractional crystallization (Fig. 10a). Therefore, partial melting of metasomatized mantle lithosphere and the melt modified by Lhasa terrane crust assimilation during ascent altogether readily explain the petrogenesis of the UPVR. The metasomatism may have genetically associated with Neo-Tethyan subduction involving seafloor sediments and/or terrigenous sediments (e.g., Himalayan basement, i.e., the northern margin of the Indian continent) (Miller et al., 1999; Ding et al., 2003; Williams et al., 2004; Guo et al., 2006, 2013; Gao et al., 2007; Zhao et al., 2009).

6.1.1. Enriched isotopic compositions and crustal contamination by ancient Lhasa basement

Because of the differences in crustal composition between the central and southern Lhasa subterraines (Zhu et al., 2011a, 2013), and because of the spatial distribution of ultrapotassic magmatism mentioned above (Fig. 1a), we infer that the crust must have contributed to the postcollisional magmatism in southern Tibet. Many lines of evidence support this inference: (i) the majority of the UPVR are localized in the central Lhasa subterrane with only a few ultrapotassic veins found in the southern Lhasa subterrane; (ii) compared with the juvenile crusts in the northern and southern Lhasa subterraines (Zhu et al., 2011a), the central Lhasa subterrane has an old basement of Proterozoic and Archean age (zircon Hf model ages up to 2870 Ma; Dong et al., 2010; Zhu et al., 2011a,b); (iii) Paleozoic–Proterozoic zircon xenocrysts have been found in the UPVR (Fig. 5c; Sun et al., 2008; Liu et al., 2011b, 2014); (iv) the least radiogenic Hf isotopic composition of co-magmatic zircons from the UPVR overlaps that of Lhasa basement (Fig. 5c). Together with the crustal xenoliths widely identified from the ultrapotassic rocks (Fig. 3a; Miller et al., 1999; Hébert et al., 2014; Liu et al., 2014), all the above data and observations signify the contribution of the ancient Lhasa basement (upper-middle crust) to the petrogenesis of the UPVR.

In the plot of δ¹⁸O vs. δ⁴⁰Ar/³⁶Ar (Fig. 9a), which is often used to assess the extent of crustal contamination in source regions (cf. Macpherson et al., 1998), the UPVR plot on strongly convex-upward hyperbolas consistent with mixing between lithospheric mantle and ancient Lhasa basement. This process also accounts for the negative correlation between δ¹⁸O and MgO in UPVR samples with MgO > 6 wt.% (Fig. 9b). The increasingly δ⁴⁰Ar/³⁶Ar with decreasing MgO (Fig. 9c) lends further support for the crustal contamination interpretation. In addition, the petrography also point to the importance of the crustal contamination. For example, crystallization of pyroxenes (both Opx and Cpx) at the expanse of olivine is a straightforward consequence of the reaction of the form olivine + SiO₂ = pyroxenes as a result of crustal contamination (addition of crustal SiO₂) or magma mixing (addition of SiO₂ from the induced crustal melt) (Fig. 3d–f). Therefore, the enriched isotopic signatures of the Tibetan
ultrapotassic rocks may be partially derived from the ancient crust of the Lhasa terrane.

6.1.2. "Garnet-signature" and material input from thickened lower crust

Crustal contamination by ancient Lhasa basement can produce extremely enriched isotopic composition of ultrapotassic magmas (Fig 9). We show here the deep crust can also contribute to the petrogenesis of the UPVR. The positive La/Yb-Dy/Yb correlation defined by the UPVR (Fig. 10b) indicates the presence of garnet as a residual phase in the source region (cf. Miller et al., 1999; Zhao et al., 2009). Previously, the “garnet-signature” in the UPVR was widely regarded as melt input from melting of garnet facies mantle lithosphere because melts derived from spinel facies mantle cannot explain the large HREE fractionation (i.e., high Dy/Yb ratios, Miller et al., 1999; Zhao et al., 2009; Prelević et al., 2012). However, it seems unlikely that the “garnet-signature” in the UPVR to have derived from partial melts from the garnet-facies mantle. This is because partial melting of garnet peridotite, even in hydrous condition ($H_2O = 1.5–5$ wt.$\%$; 1200–1450°C; 3.5 GPa), generates SiO$_2$-poor but MgO-rich melts (cf. Tenner et al., 2012) and these melts commonly display higher CaO/Al$_2$O$_3$, yet it is unobserved (Fig. 10c). In addition, mantle xenoliths (e.g., Sailipu xenoliths), which were thought to be fragments of mantle sources of the ultrapotassic magmas (Zhao et al., 2008; Liu et al., 2011a), are derived from depths of 50–65 km that is shallower than the spinel-garnet transition depth (Liu et al., 2011a). Overall, additional components with low CaO/Al$_2$O$_3$ and high SiO$_2$ are required for the petrogenesis of the UPVR (Fig. 10c).

Recent studies suggest that the pyroxene-rich veins or layers (i.e., olivine-poor pyroxenite, garnet pyroxenite or eclogite) may be the possible magma sources of alkali magmas (Hirschmann and Stolper, 1996; Hirschmann et al., 2003; Keshav et al., 2004; Sobolev et al., 2005). Moreover, partial melting of pyroxene-rich lithologies can impart the “garnet signature” independent of the great depth required by garnet-facies peridotite melting (Hirschmann and Stolper, 1996) and produce melts with moderate MgO and relatively high SiO$_2$ (Kogiso et al., 2003; Keshav et al., 2004). However, partial melts derived from mantle pyroxenite (without olivine) would generate magmas with high SiO$_2$ (up to 55$\%$) but high Ni/MgO (Sobolev et al., 2005, Herzberg, 2006), which cannot explain the low Ni/MgO and high SiO$_2$ ultrapotassic rocks in the Lhasa terrane (Fig. 10d). Also, some partial melts of garnet pyroxenite, in experiments, are too aluminous to be appropriate for the UPVR (Hirschmann et al., 2003; Kogiso et al., 2003).

Thickened lower crust (eclogite-facies or garnet-bearing granulite-facies) could be another candidate to explain the “garnet-signature” (Spandler et al., 2008; Wang et al., 2010) (Fig. 10b), especially when the continued thickening associated with the Indian–Asia convergence since the Eocene is taken into consideration (Chung et al., 2005, 2009). Compared with the contribution of garnet-facies mantle-derived melts, melt derived from the eclogitic lower crust would have high SiO$_2$ and low MgO, CaO/Al$_2$O$_3$, and Ni/MgO (Sobolev et al., 2005; Spandler et al., 2008; Wang et al., 2010). What’s more, the positive trends between Sr–$O$ isotopic ratios and MgO also support a potential contribution from the lower crust that is characterized by low $^{87}Sr/^{86}Sr$ and $\delta^{18}O_{V-SMOW}$ values (Fig. 9b and c). Other lines of evidence supporting lower crust contribution include: (i) the close spatial association between

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Fig. 6. Classification diagrams of the Xungba postcollisional volcanic rocks: (a) Total alkalis vs. SiO$_2$ (Le Maitre et al., 2005), (b) K$_2$O vs. SiO$_2$, and (c) A/NK vs. A/CNK. The smaller red circles and green squares are data from the literature (Miller et al., 1999; Chen et al., 2011; Liu et al., 2011b). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
ultrapotassic and adakitic rocks (Chen et al., 2011, 2012),
(ii) gabbro and granulite xenoliths identified in the UPVR,
which may represent the contribution from juvenile crust
(Miller et al., 1999; Chan et al., 2009; Liu et al., 2014).

Fig. 7. (a) Chondrite normalized REE patterns, and (b) primitive mantle normalized trace element patterns for Xungba volcanic rocks. Chondrite and primitive mantle values are from Boynton (1984) and Sun and McDonough (1989), respectively. Shaded fields of Miocene UPVR are from Zhao et al. (2009), and data of the Cambrian silicic metavolcanic rocks and Jurassic S-type granitoids are from Zhu et al. (2012a) and Liu et al. (2006).

Fig. 8. (a) Plot of initial $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ for postcollisional magmatic rocks in the Lhasa terrane. (b and c) Plots of lead isotopic ratios: (b) $^{207}\text{Pb}^{204}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$ and (c) $^{208}\text{Pb}^{204}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$. Qiangtang Na-rich basalts (cf. Ding et al., 2003), the PVR–UPVR in the Lhasa terrane (see references in Fig. 1), ancient Lhasa basement (Zhu et al., 2012a), GLOSS (Plank and Langmuir, 1998), gabbroic xenolith entrained by ultrapotassic magma (Miller et al., 1999), and Himalayan basement (Zhao et al., 2009) are illustrated shown for comparison. Isotopic compositions of three geochemistry end-members are given in Table 4. Three calculated isotopic mixing hyperbolas with 10% intervals are also shown for comparison. Bulk Silicate Earth (BSE), enriched mantle components (EM I and EM II) and prevalent mantle (PREMA) are from Zindler and Hart (1986). Northern Hemisphere Reference Line (NHRL): $^{207}\text{Pb}^{204}\text{Pb} = 0.1084\times^{206}\text{Pb}^{204}\text{Pb} + 13.491$; $^{208}\text{Pb}^{204}\text{Pb} = 1.209\times^{206}\text{Pb}^{204}\text{Pb} + 15.627$. Other data sources are illustrated for showing the crustal affinity of the PVR–UPVR in the Lhasa terrane: Yarlung Zangbo ophiolite (Göpel et al., 1984; Mahoney et al., 1998; Xu and Castillo, 2004; Zhang et al., 2005; Niu et al., 2006), granitoids derived from ancient Lhasa basement (Gariépy et al., 1985; Liao, 2003), and Himalayan basement (Vidal et al., 1982; Göpel et al., 1984). $^{87}\text{Sr}/^{86}\text{Sr}_{0}$, $^{143}\text{Nd}/^{144}\text{Nd}_{0}$, and $\varepsilon_{\text{Nd}}(t)$ are corrected to 23 Ma.
tions of geochemical end-members are listed in Table 4. Modeling crust (K) are shown. Therefore, considering zircons’ magmatic origin from such incompletely extracted interstitial residual melt, which is highly depleted in HREEs, would have flat HREE pattern (Liu et al., 2014). However, this explanation requires that the adakitic magmatism occurred prior to the ultrapotassic magmatism. It can be supported by Xungba postcollisional magmatism because, in the Xungba basin, ultrapotassic lavas and underlying potassic lavas with adakitic signatures form a “bimodal” volcanic sequence (Fig. 2).

The Mesozoic–Cenozoic magmatic zircons, with varying but overall positive εHf(t) values, have also been found in the Tibetan ultrapotassic rocks, indicating contributions from juvenile crust produced during episodic oceanic subduction events and India–Asia convergence (Fig. 3c) (Sun et al., 2008; Chan et al., 2009; Liu et al., 2011b, 2014). Therefore, input from thickened lower crust that had been reworked by juvenile mantle input after Mesozoic–Cenozoic magmatism is required to explain the petrogenesis of the Tibetan ultrapotassic rocks.

6.1.3. Mantle metasomatism during Neo-Tethyan oceanic subduction

In southern Tibet, the most primitive UPVR, with MgO content up to 12 wt.% (Ding et al., 2006; Zhao et al., 2009), have higher Sr–O isotopic ratios than those most involved ultrapotassic magma whose MgO content is close to 3 wt.% (Fig. 9b and c). This paradox can be explained by the metasomatism of the Tibetan mantle lithosphere prior to the assimilating crustal components mentioned above. The Fe-rich olivine and hydrous mineral phases (i.e., phlogopite) in the mantle xenoliths entrained by ultrapotassic magma also point to a metasomatized mantle source (Liu et al., 2011a). The mantle metasomatic event most likely occurred recently in response to recent Tethyan seafloor subduction and Indian lithosphere underthrusting, as is evidenced by the Mesozoic–Cenozoic magmatic zircon xenocrysts in the UPVR, which are consistent with outcropped magmatism genetically associated with the Neo-Tethyan seafloor subduction, slab rollback and break-off (Sun et al., 2008; Liu et al., 2014).

In Fig. 10, average upper, middle, and lower continental crust and partial melts derived from the ancient Lhasa basement have been shown for comparison. Crustal contamination cannot explain some trace element features of the UPVR because the continental crust contaminants would only dilute the enriched primitive melts (cf. Conticelli, 1998). Therefore, some incompatible element ratios (e.g., Th/Yb, Ba/La, Hf/Sm and Zr/Hf), which are largely immune to the crustal contamination, can be used as effective tracers for estimating mantle metasomatic agents (Fig. 10e and f) (Ben Othman et al., 1989; Dupuy et al., 1992; Ionov et al., 1993; Weyer et al., 2003; Prelević et al., 2012).

In the Th/Yb vs. Ba/La plot (Fig. 10e), ultrapotassic rocks in the Lhasa terrane display elevated Th/Yb with low and limited Ba/La. Because of the difference of element mobility in hydrous fluids (i.e., LILEs are more soluble

![Fig. 9. Plots of (a) δ18O vs. 87Sr/86Sr, (b) δ18O vs. MgO, and (c) 87Sr/86Sr vs. MgO. Strontium and oxygen isotopic data of the postcollisional UPVR and the Himalayan basement are from Zhao et al. (2009) and references therein. Isotopic compositions of geochemical end-members are listed in Table 4. Modeling crustal contamination hyperbolas with different ratios of mantle to crust (K) are shown.](image)
than HFSEs and REEs), the large Th/Yb (vertical) variation reflects source input from subducted sediments rather than slab-derived fluids (Woodhead et al., 2001). Additionally, considering the similar geochemical behavior among Zr, Hf, and Sm during partial melting (McDonough, 1990), Tibetan ultrapotassic rocks have fairly constant chondritic Zr/Hf ratios (30–40) with Hf/Sm ratios varying over a large range and deviating from the chondritic value for Hf/Sm (0.69) (Fig. 10f). The Hf/Sm ratios of continental crust components, terrigenous sediments, turbidites and mature sands, with high abundances of zircons, are commonly super-chondritic (Fig. 10f) because zircon is Hf-enriched and LREE-depleted (Hoskin and Schaltegger, 2003; Prelević et al., 2012). The super-chondritic Hf/Sm ratios of some UPVR samples therefore may imply an additional input of zircon-rich sediments (e.g., terrigeneous sediments, turbidites, mature sands) to the mantle source region or assimilating upper-middle continental crust components during magmatic ascent. However, many ultrapotassic samples, including the Xungba UPVR, display sub-chondritic Hf/Sm ratios and negative variation between Hf/Sm and Zr/Hf (Fig. 10f), suggesting other
metasomatic agents in the mantle sources. The pelagic sediments and marine pelite, which contain relatively few zircons, are characterized by much lower Hf/Sm ratios (Fig. 10f). Furthermore, the appearance of low Hf/Sm but high Zr/Hf ratios in some UPVR samples may reflect a mantle source region modified by carbonate-related metasomatism (Fig. 10f) (Dupuy et al., 1992; Ionov et al., 1993). Thus, the mantle sources of the UPVR in the Lhasa terrane may have also been enriched by pelagic sediments and marine carbonates derived from the Neo-Tethyan seafloor subduction. This interpretation is further supported by the high (La/Yb)N but low Ti/Eu ratios displayed by clinopyroxenes in mantle xenoliths (Liu et al., 2011a), which points to carbonate-related mantle metasomatism (Coltorti et al., 1999).

Considering the underthrusting of Indian continental lithosphere following the Indian–Asia collision (cf. Freymueller, 2011), it is apparent that the Himalayan basement is a possible agent in metasomatizing the source region of ultrapotassic rocks (Ding et al., 2003; Zhao et al., 2009). However, both the Indian continent and central Lhasa subterrane are derived from the breakup and dispersal of Gondwanaland (Zhu et al., 2011b; Zhang et al., 2012) and have similar Sr–Nd–O isotopic compositions (Fig. 8a) (Vidal et al., 1982; Göpel et al., 1984; Zhu et al., 2013). Accordingly, it is difficult to distinguish Indian material inputs into mantle sources of UPVR from the influence of crustal contamination of ancient Lhasa basement. If the spatial distribution of the ultrapotassic magmatism and the compositional differences between the central (ancient) and northern/southern (juvenile) subterranes are considered, we can find that the Tibetan mantle lithosphere metasomatism may have more complex histories and the recent metasomatized lithosphere is only one end-member controlling the geochemical signatures of the UPVR (Williams et al., 2004; Zhao et al., 2009; Guo et al., 2013).

In summary, postcollisional ultrapotassic rocks in the Xungba basin as well as those from the entire Lhasa terrane were derived from a phlogopite-bearing spinel peridotite source, which had been metasomatized during previous Tethyan oceanic subduction, and then mixed with melts derived from thickened lower crust (eclogite-facies or garnet-bearing granulate facies) and ancient Lhasa basement (middle or upper crust) during magmatic ascent. More work is needed to distinguish the Indian input into the mantle sources of ultrapotassic rocks, if this process did exist during continental convergence, from crustal contamination of the central Lhasa subterrane crust.

6.2. Petrogenesis of adakitic potassic volcanic rocks (Group 2)

In the Sr–Nd isotopic array (Fig. 8a), Xungba potassic volcanic rocks (Group 2) show higher Sr initial isotopic ratios relative to the overlying UPVR and coeval potassic magmatism in the central Lhasa subterrane (Chen et al., 2010, 2011). Compared with the Xungba UPVR, the PVR have lower incompatible elements (e.g., Th, U, Sr, Ba; Fig. 7b) and distinct Sr–Nd isotopes, ruling out an origin by fractional crystallization from the mantle-derived ultrapotassic magmas. Furthermore, the high SiO2, low MgO and compatible elements (e.g., Cr, Ni) concentrations also indicate that the Xungba PVR were derived from crustal sources rather than a mantle source (Table 2).

Despite the similarities in Sr–Nd–O isotopic compositions, the Xungba PVR could not have formed by partial melting of subducted Indian continent for two reasons: (i) their Pb isotopic compositions resemble the Lhasa basement rather than the Himalayan basement (Gariepy et al., 1985; Liao, 2003), and (ii) the PVR have low MgO and compatible elements abundances (Cr = 32–129 ppm, Ni = 4–74 ppm), which is inconsistent with melts derived from subducted continental crust since this kind of melt will inevitably react with the mantle wedge and acquire excess compatible elements. The crustal contamination trend between δ18Ov,SMOW and 87Sr/86Sr and negative correlations between Sr-O isotopic ratios and MgO confirm that the evolved isotopic compositions of the PVR are due to assimilation of ancient Lhasa basement (Fig. 9b and c).

In discrimination diagrams (Fig. 11a and b) (Defant and Drummond, 1990; Richards and Kerrich, 2007), except for a few samples with lower Sr/Y ratios, most Xungba PVR have adakitic geochemical signatures with high Sr/Y and La/Yb ratios but low Y and Yb abundances. However, the negative Sr and Eu anomalies observed in the PVR suggest that the Group II rocks are not typical “adakite”. Aside from melting a subducted young oceanic crust, an adakitic signature can be achieved through: (i) fractional crystallization of clinopyroxene that is selectively enriched in HREEs (Castillo et al., 1999; Richards and Kerrich, 2007); (ii) partial melting of high Sr/Y sources under low pressure conditions (cf. Moyen, 2009); and (iii) melting of thickened lower crust under eclogite-facies conditions (cf. Chung et al., 2009). In the case of the Xungba PVR, no fractionation trends of pyroxene and amphibole can be found in the Rb/Sr–Sr–Cr diagrams (Fig. 11c), precluding the adakitic signature from fractional crystallization of HREE-rich minerals. The adakite-like signature of the PVR could not have originated from high-Sr/Y sources, as the Sr/Y ratio of gabbro xenolith entrained by ultrapotassic magma is as low as 32.8 (Miller et al., 1999), slightly higher than that of average lower crust (Sr/Y = 21.8, Rudnick and Gao, 2003). If we consider the compressional regime during the Eocene (Chung et al., 2005), the Xungba PVR is best interpreted as resulting from partial melting of thickened lower crust. However, in comparison with typical adakitic melts (Richards and Kerrich, 2007; Castillo, 2012), the PVR are marked by a weak negative Eu anomaly in general and a few PVR samples have relatively low Sr/Y ratios and strong negative Eu anomalies (Fig. 11a and d). These differences can be reconciled with the existence of plagioclase fractional crystallization. As illustrated in Fig. 11a, a rapid decreasing trend in Sr/Y ratios with plagioclase fractionation can be modeled by assuming the parent adakitic magma containing 1000 ppm of Sr and 12 ppm of Y. The plagioclase fractionation trend shown in mineral vector diagram (Fig. 11e) and the positive correlation trend between Eu/Eu* and Sr (Fig. 11d) are consistent with fractional crystallization of plagioclase in the petrogenesis of the PVR. Therefore, the Xungba PVR with adakitic
geochemical signatures were derived from thickened lower crust and modified by crustal contamination with ancient Lhasa basement and by fractional crystallization of plagioclase.

6.3. Geodynamic implications

Although the exact timing and processes of plateau uplift remain in dispute, postcollisional magmatism and geologic episodes, including accelerated uplift of southern Tibet (Harrison et al., 1992; Yin and Harrison, 2000), changes in crustal stress state and deformation style (Blisniuk et al., 2001; Kapp et al., 2008; Mitsuishi et al., 2012), high rates of denudation and sedimentation (Clift and Gaedicke, 2002; Uddin and Lundberg, 2004; Carter et al., 2010; Enkelmann et al., 2011; Kirstein, 2011), rapid increase of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in seawater (DePaolo and Ingram, 1985; Krishnaswami et al., 1992) and the transformation of global climate system from zonal to a monsoonal pattern (Guo et al., 2008), all appear to have taken place simultaneously in the early Miocene (Fig. 12). Therefore, as the earliest magmatic record, the “bimodal” volcanic sequences formed by early-Miocene ultrapotassic lavas and underlying potassic lavas in Xungba basin may symbolize the onset of geodynamic processes in southern Tibet during postcollisional stage and point to an extensional tectonic regime.

However, the synchronous adakitic magmatism and the petrogenetic relationship between Tibetan ultrapotassic rocks and eclogite-facies lower crust require a thickened lower crust for postcollisional magmatism. An effective geodynamic model should be able to reconcile the “paradox” between extensional tectonic regime and the overthickened crust, and to accommodate northward underthrusting of the Indian lithosphere with the existence of thickened Tibetan lower crust. A possible scenario is that the lower part of overthickened lithospheric mantle was delaminated, which may have triggered postcollisional magmatism and surface uplift and extension in the southern Tibet as suggested by previously (Chung et al., 2005; Zhao et al., 2009). In this scenario, with continued northward underthrusting of Indian continental lithosphere, the rheologically heterogeneous lithospheric mantle would become overthickened. Because of the gravitational instability and mantle convection, the delamination started from the northern front of the overthickened lithospheric mantle and was gradually replaced by hotter asthenosphere. Consequently, the upwelling of the asthenosphere resulted in partial melting of the residual part of lithospheric mantle metasomatized during Mesozoic–Cenozoic Neo-Tethyan seafloor...
subduction and thickened lower crust, generating the ultrapotassic and potassic volcanism in the central Lhasa subterrane, respectively. The poloidal mantle flow around the Lhasa terrane (Fig. 1 a).

7. CONCLUSIONS

(1) The postcollisional volcanic sequences in the Xungba basin consist of ultrapotassic (UPVR) and potassic volcanic rocks (PVR) that erupted coevally at ~23 Ma.

(2) The UPVR resulted from partial melting of hydrous mantle lithosphere under spinel facies conditions. The mantle lithosphere has experienced multiple metasomatic events during the Neo-Tethyan seafloor subduction. The thickened lower crust and ancient Lhasa basement rocks have both contributed to the Tibetan ultrapotassic magmatism.

(3) The PVR with adakitic signature resulted from melting of the thickened lower crust with the melt contaminated by ancient Lhasa basement rocks during ascent. Plagioclase crystallization is responsible for the low Sr/Y and high Y for some of the PVR samples, but has little effect on the La/Yb and Yb.

(4) The extensional tectonic regime and the thickened crust together make the UPVR and PVR magmatism possible, and the UPVR and PVR rocks offer prime opportunities to discuss deep processes in the early Miocene. Delamination of lower part of the thickened mantle lithosphere is a possible scenario to explain the plateau uplift, surface extension and the close relationship between postcollisional magmatism and thickened lower crust.

ACKNOWLEDGEMENTS

The work was supported by the National Key Project for Basic Research (2011CB403102), NSF of China (Grants 41273044, 40873023, 41225006), NSF of U.S. (Grants EAR1111959, 1111586), CGS-121201121260, Sinoprobe-04-02, IRT1108, 111 Program (B07011), and IGCP/SIDA-600. We appreciate three anonymous reviewers for constructive comments and Editors Marc Norman and Weidong Sun for handling the manuscript. We thank Yongsheng Liu, Zhaocuo Hu and Honglin Yuan for assistance in the zircon analysis, and Shihong Tian in whole rock oxygen isotope analysis.

APPENDIX A. SUPPLEMENTARY DATA

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.gca.2014.03.031.

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Associate editor: Weidong Sun