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3 **Post-collisional ultrapotassic mafic magmatism in south Tibet: Products of**
4 **partial melting of pyroxenite in the mantle wedge induced by roll-back and**
5 **delamination of the subducted Indian continental lithosphere slab**
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Abstract

Post-collisional (25 Ma to 8 Ma) ultrapotassic mafic magmatic rocks occur to the north of the India-Asia collision zone within the Lhasa terrane of the southern Tibetan Plateau, forming a near 1000 km long semi-continuous igneous belt. They include both extrusive and intrusive facies, though lava flows dominate. In order to understand their petrogenesis, the mineral chemistry of olivine phenocrysts and xenocrysts and whole-rock major and trace element and Sr-Nd-Pb isotope data are presented for the most primitive mafic magmatic rocks (MgO > 6 wt%) from west to east. The studied samples are characterized by high MgO (6.28-15.75 wt%), K₂O (4.76-8.89 wt%), SiO₂ (46.44-59.74 wt%), Ba (1368-14076 ppm), Th (69-336 ppm) and Ni (106-527 ppm) contents. Chondrite-normalised rare earth element (REE) patterns show enrichment in light rare earth elements (LREE), flat heavy REE (HREE) patterns and negative Eu anomalies. These REE patterns have a very distinctive inverted “spoon shape”, which appears to be a common characteristic of collision-related ultrapotassic magmas. Primitive mantle-normalised incompatible trace element patterns exhibit strong enrichments in large ion lithophile elements (LILE) relative to high field strength elements (HFSE) and strong negative Ta-Nb-Ti anomalies, which are typical of subduction-related magmas. The ultrapotassic magmatic rocks studied have extremely radiogenic initial Sr isotopic compositions (0.712379-0.737616) and low (¹⁴³Nd/¹⁴⁴Nd)_i(0.511662-0.511984). Combined with their Pb isotope compositions [(²⁰⁶Pb/²⁰⁴Pb)_i = 18.30-18.92; (²⁰⁷Pb/²⁰⁴Pb)_i = 15.65-15.87; (²⁰⁸Pb/²⁰⁴Pb)_i = 39.02-39.76] these data are consistent with the involvement of a subducted continental crustal component in their petrogenesis. The Sr-Nd-Pb isotope compositions exhibit linear trends between depleted MORB-source mantle (DMM) and Indian continental crust. The extreme enrichment of the upper mantle below south Tibet is

1
2 considered to result from the addition of components derived from subducted Indian
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4 continental crust to the overlying mantle wedge during northwards underthrusting of Indian
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6 continental lithosphere beneath the Lhasa terrane since India-Asia collision at ~55 Ma. The
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8 post-collisional K-rich mafic magmas in south Tibet were generated by partial melting of
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10 pyroxenite in a mantle source region which was created by reaction of hydrous fluids and
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12 siliceous melts from subducted granulite-eclogite facies Indian continental crustal rocks with
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14 the surrounding peridotitic mantle. A continuous process from slab roll-back, through
15
16 break-off, to detachment of the slab may have induced partial melting of the pyroxenites.
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18 Cessation of the post-collisional ultrapotassic magmatism at ~ 8Ma may be linked to the onset
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20 of flat slab subduction beneath southern Tibet and the elimination of the wedge of Tibetan
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22 sub-continental lithospheric mantle and underlying asthenosphere; geophysical data indicate
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24 that at the present-day eclogite facies Indian continental crust directly underthrusts the crust
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26 of the Lhasa terrane with no intervening mantle wedge. The proportion of the Indian
27
28 continental crustal component in the mantle source of the ultrapotassic mafic magmas
29
30 decreases eastward, as do the ages and volumes of the magmatic rocks. There are no outcrops
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32 of post-collisional K-rich mafic magmatic rocks ($MgO > 6$ wt%) to the east of 87°E in the
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34 Lhasa terrane which may indicate a change in subduction geometry at this longitude.
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Keywords: Post-collisional ultrapotassic magmatism; Indian continental subduction;
pyroxenite; slab rollback; Lhasa terrane; south Tibet

INTRODUCTION

Post-collisional potassic and ultrapotassic magmatism is widely distributed within the Alpine-Himalayan orogenic belt from Eastern Europe to the Tibetan Plateau. Understanding the petrogenesis of the most primitive K-rich magmatic rocks can provide important constraints on the evolution of the upper mantle in such continent-continent collision zones. However, their petrogenesis remains controversial, despite numerous studies (e.g. Pearce & Mei, 1988; Arnaud *et al.*, 1992; Turner *et al.*, 1996; Williams *et al.*, 2001; Ding *et al.*, 2003; Nomade *et al.*, 2004; Mo *et al.*, 2006; Gao *et al.*, 2007; Zhao *et al.*, 2009; Chen *et al.*, 2010, 2011, 2012; Guo *et al.*, 2013, 2014; Wang *et al.*, 2014; Liu *et al.*, 2014a, b, c).

Within the Lhasa terrane of south Tibet post-collisional, K-rich magmatic rocks are distributed in a linear belt to the north of the Indus-Tsangpo Suture (ITS) (Fig. 1). Three hypotheses have been proposed to explain this magmatism: (1) It is the product of convective removal of previously thickened lithospheric mantle (e.g. Turner *et al.*, 1993, 1996; Williams *et al.*, 2001, 2004; Chung *et al.*, 2005; Zhao *et al.*, 2009); (2) It is associated with northward subduction of Indian continental lithosphere (e.g. Pearce & Mei, 1988; Arnaud *et al.*, 1992; Tapponnier *et al.*, 2001; Ding *et al.*, 2003; Guo *et al.*, 2013) and (3) It is the consequence of the break-off of the subducted Indian continental lithosphere slab (e.g. Miller *et al.*, 1999; DeCelles *et al.*, 2002; Mahéo *et al.*, 2002; Replumaz *et al.*, 2010, 2013, 2014) or slab roll-back (e.g. Guo *et al.*, 2013). Because these previous studies mainly focused on individual volcanic fields it has been difficult to develop a holistic petrogenetic model which has a realistic geodynamic context.

Ultrapotassic mafic magmatism within the Lhasa terrane occurs to the west of 87°E (Fig.

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3 1), whereas more evolved potassic magmatism is distributed throughout the terrane. In
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5 contrast, contemporaneous, post-collisional, ore-bearing, adakite-like intrusives, including the
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7 largest porphyry Cu-Mo deposit in China (Qulong) (e.g. Hou *et al.*, 2004, 2013; Yang *et al.*,
8
9 2009), are mainly distributed in the eastern part of the Lhasa terrane (Fig. 1), although rare
10
11 Miocene adakite-like intrusives do occur further to the west (Guo *et al.*, 2007, 2013; Zhao *et*
12
13 *al.*, 2009). It is significant that all of the post-collisional potassic and ultrapotassic magmatic
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15 rocks and adakite-like intrusives are restricted in their distribution to the north of the
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17 Indus-Tsangpo Suture (ITS) (Fig. 1), whereas contemporaneous Miocene leucogranites (Guo
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19 & Wilson, 2012) occur to the south of the ITS in the Himalayas.

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None of the hypotheses mentioned above can fully explain the distribution patterns of
the post-collisional potassic and ultrapotassic magmatic rocks, adakite-like intrusives and
leucogranites in south Tibet and the Himalayas. Additionally, most of the previously studied
samples were of evolved magmatic rocks ($\text{MgO} < 6 \text{ wt}\%$), the parental magmas of which must
have undergone combined crustal contamination and fractional crystallization (Guo *et al.*,
2006, 2013), making it difficult to constrain the nature of their mantle source. The lack of
detailed field sampling and of petrological, geochronological and geochemical data for the
whole volcanic belt from east to west in the Lhasa terrane (Fig. 1) has precluded further
advances in understanding the petrogenesis of the magmas.

In this study, we report new bulk-rock major and trace element, Sr-Nd-Pb isotope and
olivine phenocryst composition data for the entire, near 1000 km long, post-collisional
ultrapotassic mafic magmatic belt (Fig. 1) in the Lhasa terrane of the southern Tibetan Plateau.
Samples were selected to minimise the effects of shallow-level crustal contamination and
magmatic differentiation. Our samples are representative of all of the post-collisional K-rich

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2 mafic volcanic fields recognized in previous studies (e.g. Miller *et al.*, 1999; Williams *et al.*,
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4 2001; Ding *et al.*, 2003, 2006; Chen *et al.*, 2007; Gao *et al.*, 2007; Guo *et al.*, 2013) and also
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7 from additional outcrops new to our study (Fig. 1). On the basis of our new data and
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10 previously published geochemical and geophysical data, we develop a robust petrogenetic
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12 model to better explain genesis of the post-collisional ultrapotassic mafic magmas, their
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14 mantle source characteristics and geodynamic setting.
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20 **GEOLOGICAL SETTING**

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25 The Tibetan Plateau is a collage of E-W trending allochthonous micro-continent terranes
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27 (Kunlun, Songpan-Ganzi, Qiangtang and Lhasa), which were successively accreted to the
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29 southern margin of the Asian continent since Palaeozoic times (Fig. 1). The Lhasa terrane is
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31 bordered by the Bangong-Nujiang Suture (BNS) to the north and the Indus-Tsangpo Suture
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33 (ITS) to the south (Fig. 1). Collision between the Indian and Asian continents at ~ 55-50Ma
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35 resulted in the formation of the ITS and the Tibetan-Himalayan orogenic belt (e.g. Patriat &
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37 Achache, 1984; Klootwijk *et al.*, 1992; Leech *et al.*, 2005; Zhu *et al.*, 2005; Royden *et al.*,
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39 2008; Najman *et al.*, 2010; Sun *et al.*, 2012).
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45 Extensive magmatic activity has characterized the Lhasa terrane during Cenozoic times,
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47 associated with India-Asia continent-continent collision. Three types of magmatism can be
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49 distinguished based on their temporal-spatial distribution and geochemical characteristics.
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51 The first type includes calc-alkaline volcanism (Linzizong volcanics) and associated
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53 granitoid batholiths (Gangdese batholith) during the period 65 to 40 Ma (Scharer *et al.*, 1984;
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55 Coulon *et al.*, 1986; Zhou *et al.*, 2004; Mo *et al.*, 2005, 2006; Wen *et al.*, 2008; Niu *et al.*,
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2009; Fig. 1), considered to be the consequence of northward subduction of Neo-tethyan oceanic lithosphere beneath the Lhasa terrane. The second type comprises Oligocene-Miocene (26-10 Ma) adakite-like intrusives, which are mainly distributed in the eastern part of the Lhasa terrane (Fig. 1) and have been proposed to be the products of melting of thickened lower crust (e.g. Chung *et al.*, 2003; Hou *et al.*, 2004; Guo *et al.*, 2007; Zheng *et al.*, 2012; Zhang *et al.*, 2014). The third type consists of the ultrapotassic and potassic magmatic rocks which are the focus of this study (Fig. 1); their ages range from 25 to 8 Ma based on $^{40}\text{Ar}/^{39}\text{Ar}$, K-Ar, Rb-Sr and single-crystal zircon U-Pb dating (e.g. Coulon *et al.*, 1986; Yin *et al.*, 1994; Miller *et al.*, 1999; Williams *et al.*, 2001, 2004; Ding *et al.*, 2003; Nomade *et al.*, 2004; Mo *et al.*, 2006; Zhao *et al.*, 2006; 2009; Sun *et al.*, 2008; Chen *et al.*, 2010; Zhou *et al.*, 2010; Guo *et al.*, 2013; Liu *et al.*, 2011, 2014). They outcrop as small-volume lava flows, plugs and dyke swarms within a series of north-south trending rifts bounded by normal faults (Guo *et al.*, 2013; Fig. 1). Recent geophysical studies (e.g. Li *et al.*, 2008; Nabelek *et al.*, 2009; Zhao *et al.*, 2011; Zhang *et al.*, 2014) suggest that at present the Tibetan Plateau crust beneath these volcanic fields is underlain by a flat-lying layer of inferred subducted Indian continental crust, ~ 15-20 km thick, with no intervening mantle wedge. This lower layer of the crust could be made of amphibolite-, granulite- or eclogite- facies metamorphic rocks (Worthington *et al.*, 2013; Zhang *et al.*, 2014).

Ultrapotassic volcanic fields

Thirteen ultrapotassic mafic volcanic fields have been identified which form the basis of this study, extending from the Shiquanhe volcanic field in the west, through the Bongba, Xungba, Jarga, Sailipu, Zabuye, Maiga, Garwa, Yaqian, Mibale, Yiqian and Chazi volcanic

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2 fields to the Pabbai volcanic field in the east (Fig. 1). Exposures range from less than $\sim 1 \text{ km}^2$
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4 to $\sim 600 \text{ km}^2$. Lava flow thicknesses vary from $\sim 30 \text{ m}$ in the Yiqian volcanic field to about 300
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6
7 m in the Zabuye volcanic field. The volume of magmatism appears to decrease from west to
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9 east; the Maiga and Yiqian volcanic fields in the east have the smallest exposed areas of less
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11 than 1 km^2 , whereas the Xungba, Sailipu and Zabuye volcanic fields in the west have the
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13 largest exposed areas of $400\text{-}600 \text{ km}^2$. For ease of reference we have sub-divided the volcanic
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15 fields into three sub-groups based upon longitude: (1) a western sub-group (WSVF; including
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17 the Shiquanhe, Bongba, Xungba, Jarga and Sailipu volcanic fields), (2) a central sub-group
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19 (CSVF; including the Zabuye and Maiga volcanic fields) and (3) an eastern sub-group (ESVF;
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21 including the Garwa, Yaqian, Mibale, Yiqian, Chazi and Pabbai volcanic fields) (Fig. 1).
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30 **Age of the volcanism**

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32 To study the timing of the ultrapotassic mafic magmatism we have compiled all the
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34 available geochronological data from the published literature (Table 1), including Miller *et al.*
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36 (1999), Williams *et al.* (2001, 2004), Nomade *et al.* (2004) and Guo *et al.* (2013). A range of
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38 geochronometers (e.g. $^{40}\text{Ar}/^{39}\text{Ar}$, K-Ar, U-Pb, Rb-Sr) have been used and we have checked
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40 the data quality and compared the ages from the different methods in order to identify any
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42 age trends. Our analysis of the data (Fig. 2 and Table 1) indicates that the age of the
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44 magmatism ranges from 25 Ma to 8 Ma and shows a decreasing trend from north to south
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46 (Fig. 2a) and from west to east (Fig. 2b). This is a very interesting and important observation,
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48 which may be consistent with Indian slab roll-back during the period 25Ma to 8Ma from
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50 north to south combined with slab break-off that propagated eastwards from 25 Ma to 10 Ma,
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52 as proposed by Guo *et al.* (2013) and Replumaz *et al.* (2010, 2013, 2014).
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2 The magmatism in the eastern sub-group (ESVF) ranges widely in age from 25 to 8 Ma
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4 (Guo *et al.*, 2013). Here volcanism youngs from north to south within a single N-S trending
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6 rift and has been interpreted by Guo *et al.* (2013) to reflect magma generation in response to
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8 progressive roll-back of the subducted Indian lithosphere.
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11 12 13 14 15 **PETROGRAPHY**

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20 The post-collisional ultrapotassic mafic magmatic rocks ($\text{MgO} > 6 \text{ wt\%}$) studied have
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22 porphyritic textures with phenocrysts of clinopyroxene, phlogopite, olivine and rare sanidine;
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24 the groundmass includes phlogopite, clinopyroxene, olivine, sanidine, apatite, Fe-Ti oxides,
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26 leucite, zircon and glass. The typical size of the phenocrysts is 1-5mm. The main petrographic
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28 characteristics are summarized in Table 2.
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33 Mantle-derived olivine xenocrysts have been recognized in the ultrapotassic magmatic
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35 rocks of the Xungba and Sailipu volcanic fields (Fig. 1). Some samples from the Sailipu
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37 volcanic field also contain mantle xenoliths, which comprise olivine, orthopyroxene,
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39 clinopyroxene and spinel (Zhao *et al.*, 2009; Liu *et al.*, 2011, 2014a). The olivine xenocrysts
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41 are characterized by disequilibrium textures and reaction rims (Fig. 3a,f). Most of the olivine
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43 phenocrysts in the most primitive magmatic rocks of the Sailipu and Maiga volcanic fields
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45 (e.g. $\text{MgO} \approx 15 \text{ wt\%}$ in sample MG1303; Table 3) have corroded rims (Fig. 3b). Some of the
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47 olivine xenocrysts in the ultrapotassic rocks of the Xungba and Sailipu volcanic fields have
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49 reaction rims of pyroxene (Fig. 3c). Phlogopite and clinopyroxene (Fig. 3d,e) occur as
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51 euhedral phenocrysts and microphenocrysts.
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ANALYTICAL METHODS

All of the analysed samples show no evidence of significant hydrothermal alteration or weathering. Samples 4-5 kg in weight were cut into thin slices which were cleaned three times using deionized water, dried, and then crushed between corundum plates and finally ground into whole-rock powders in an agate mortar in preparation for whole-rock major element, trace element and Sr-Nd-Pb isotope analysis.

Whole-rock major and trace element analyses

Whole-rock major element contents were determined on fused glass discs by X-ray fluorescence (XRF) using an XRF-1500 sequential spectrometer at the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGGCAS). Sample powders (0.6 g) were fused with $\text{Li}_2\text{B}_4\text{O}_7$ (6 g) in a TR-1000S automatic bead fusion furnace at 1100°C for 10 min. Loss on ignition (LOI) was determined by ignition at 1100°C for 10 hours of 2g powders. The analytical precision was better than 2 % relative, including the errors in SiO_2 and Al_2O_3 contents inherited from sample crushing using corundum plates and grinding in an agate mortar. The detailed analytical method follows that given in Guo *et al.* (2006, 2013). Whole-rock major element compositions are reported in Table 3.

Whole-rock trace element contents were determined by inductively coupled plasma mass spectrometry (ICP-MS) using a FINNIGAN MAT II element system at IGGCAS. Whole-rock powders (40 mg) were dissolved in distilled 1 ml 20N HF and 0.5 ml 7.5N HNO_3 in 7 ml Savillex Teflon screw-cap capsules and then were ultrasonically stirred for 15 min. Then, the solutions were evaporated at 150°C to dryness and the residue was digested with 1.5 ml 20N HF and 0.5 ml HNO_3 [HNO_3 : H_2O =1:1 (volume ratio)] in Teflon screw-cap capsules. The

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2 solutions were heated at 130°C for 2 h, then their temperature was gradually increased up to
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4 170°C over 24 h. The solutions were then maintained at 170°C for 10 days, dried and
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6 redissolved in 2 ml HNO₃ [HNO₃: H₂O=1:1 (volume ratio)] in the capsules. The solutions
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8 were heated at 150°C for 5 h and then evaporated, dried and redissolved in 2 ml HNO₃ [HNO₃:
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10 H₂O=1:1 (volume ratio)] and 2 ml 1 % HNO₃ at 150°C for 5 h in screw-cap capsules in order
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12 to ensure that the samples were completely dissolved. The solutions were put into plastic
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14 beakers and then 1 ml 500 ppb In was added as an internal standard. Finally, the solutions
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16 were diluted in 1 % HNO₃ to 50 ml for analysis by ICP-MS. A blank solution was prepared;
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18 the total procedural blanks were < 50 ng for all the trace elements reported in Table 3. During
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20 the analytical runs, frequent standard calibrations were performed to correct for instrumental
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22 signal drift following the procedure of Guo *et al.* (2006). Four replicates and two international
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24 standards (BHVO-1 and AGV-1) were prepared using the same procedure to monitor the
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26 analytical reproducibility. The discrepancy, based on repeated analyses of samples and
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28 international standards, is less than 4 % for the trace elements reported in Table 3. Analyses of
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30 the international standards are in excellent agreement with the recommended values
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32 (Govindaraju, 1994), and deviate less than 5 % from the published values (Table 3). The
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34 detailed analytical procedures follow those of Guo *et al.* (2006, 2013).
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48 **Whole-rock Sr-Nd-Pb isotope analyses**

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50 Whole-rock Sr-Nd-Pb isotope analyses were performed on a Finnigan MAT262 mass
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52 spectrometer at IGGCAS. Whole-rock powders (60 mg) were spiked with mixed isotope
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54 tracers (⁸⁷Rb-⁸⁴Sr for Rb-Sr isotope analyses and ¹⁴⁹Sm-¹⁵⁰Nd for Sm-Nd isotope analyses),
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56 then dissolved with a mixed acid (HF: HClO₄ = 3:1; volume ratio) in Teflon capsules for 7
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2 days at room temperature. Rb and Sr and rare earth element fractions were separated in
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4 solution using AG50W×8 (H⁺) cationic ion-exchange resin columns. Sm and Nd were
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6 separated from the other rare earth element fractions in solution using AG50W×8 (H⁺)
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8 cationic ion-exchange columns and P507 extraction and eluviation resin. The collected Sr and
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10 Nd fractions were evaporated and dissolved in 2 % HNO₃ to give solutions for analysis by
11
12 mass spectrometry. The mass fractionation corrections for Sr and Nd isotopic ratios were
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14 based on ⁸⁶Sr/⁸⁸Sr = 0.1194 and ¹⁴⁶Nd/¹⁴⁴Nd = 0.7219, respectively. The international standard
15
16 NBS987 yielded ⁸⁷Sr/⁸⁶Sr = 0.710254 ± 16 (n=8, 2 sigma) (the recommended value is
17
18 0.710240) and international standard NBS607 yielded ⁸⁷Sr/⁸⁶Sr = 1.20032 ± 30 (n=12, 2
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20 sigma) (the recommended value is 1.20039). The international La Jolla standard gave
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22 ¹⁴³Nd/¹⁴⁴Nd = 0.511862 ± 7 (n=12, 2 sigma) (the recommended value is 0.511859) and
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24 international standard BCR-1 gave ¹⁴³Nd/¹⁴⁴Nd = 0.512626 ± 9 (n=12, 2 sigma) (the
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26 recommended value is 0.512638). The whole procedure blank is less than 200 pg for Rb-Sr
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28 isotopic analysis and 50 pg for Sm-Nd isotopic analysis. Analytical errors for Sr and Nd
29
30 isotopic ratios are reported as 2 sigma (2σ) in Table 4. The initial ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd
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32 ratios were calculated using the average ages of the samples based on ⁴⁰Ar/³⁹Ar, K-Ar and
33
34 U-Pb zircon dating methods (Table 1).

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36 For whole-rock Pb isotope analyses, 150 mg powder was weighed and dissolved in
37
38 Teflon capsules using concentrated HF at 120 °C for 7 days. Pb was separated from the
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40 silicate matrix and purified using AG1×8 anionic ion-exchange columns with dilute HBr as
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42 eluant. The whole procedure blank is less than 1 ng. During the period of analysis repeat
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44 analyses of the international standard NBS981 gave ²⁰⁴Pb/²⁰⁶Pb = 0.059003 ± 0.000084 (n = 6,
45
46 2 sigma) (the certified value is 0.058998), ²⁰⁷Pb/²⁰⁶Pb = 0.91449 ± 0.00017 (n = 6, 2 sigma)
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2 (the certified value is 0.914598) and $^{208}\text{Pb}/^{206}\text{Pb} = 2.16691 \pm 0.00097$ ($n = 6$, 2 sigma) (the
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4 certified value is 2.168099). Pb isotope fractionations were corrected using correction factors
5
6 based on replicate analyses of the international standard NBS981. The Pb isotope data are
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8 reported in Table 5. Detailed sample preparation and analytical procedures for the Sr-Nd-Pb
9
10 isotope measurements follow those of Guo *et al.* (2006, 2013).
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18 **Olivine composition analyses**

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20 Compositions of olivine phenocrysts and xenocrysts were analyzed by electron
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22 microprobe (JEOLJXA-8100) at the IGGCAS. Operating conditions were 15 kV accelerating
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24 voltage, 10-20 nA beam current, 3-5 μm beam diameter and 15-20 s counting time.
25
26 Well-defined natural and synthetic mineral standards were used for calibration. The analytical
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28 errors are generally less than 2%. Olivine compositions are reported in Supplementary Data
29
30 Appendix A. The detailed analytical procedures follow those of Ding *et al.* (2003) and Wang
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60
et al. (2014).

40 **RESULTS**

45 **Major element characteristics**

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47 The ultrapotassic mafic magmatic rocks have relatively high *Mg*-numbers [molar
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49 $\text{Mg} \times 100 / (\text{Mg} + \text{Fe}^{2+}) = 70-88$, calculated assuming $\text{Fe}_2\text{O}_3 / (\text{FeO} + \text{Fe}_2\text{O}_3) = 0.20$], MgO contents
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51 (6.28 -15.75 wt%) and Ni (106 -527 ppm) concentrations (Table 3). The studied samples are
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54 thus considered to have relatively primitive compositions, minimising the likelihood of
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60 significant crustal contamination and modification by magmatic differentiation. Whilst an

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3 $\text{Fe}_2\text{O}_3/(\text{FeO}+\text{Fe}_2\text{O}_3)$ ratio of 0.20 might be considered high, this is consistent with the
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5 observations of Brounce *et al.* (2014) who demonstrated that subduction-related magmas are
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7 more oxidizing.
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10 The whole-rock geochemical data (Table 3) indicate that all the studied samples are
11
12 ultrapotassic, with $\text{K}_2\text{O}>3$ wt %, $\text{MgO}>3$ wt % and $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratios ranging from 2.1 to 8.7
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14 (wt %), based on the criteria of Foley *et al.* (1987) and Woolley *et al.* (1996). The studied
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16 rocks plot almost totally within the basaltic trachyandesite and trachyandesite fields (S2,S3)
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18 in an $\text{Na}_2\text{O}+\text{K}_2\text{O}$ versus SiO_2 classification diagram (Fig. 4a) and in the shoshonitic field in a
19
20 plot of K_2O versus SiO_2 (Fig. 4b), following the criteria of Peccerillo & Taylor(1976). Rare
21
22 silica-poor samples plot in the basanite-tephrite field (U1, U2). Following the classification
23
24 scheme of Foley *et al.* (1987), most of the post-collisional magmatic rocks studied are
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26 lamproitic, although they straddle the boundary between lamproites and Roman Province
27
28 ultrapotassic rocks (Fig. 5).
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35 With decreasing MgO contents, SiO_2 , Al_2O_3 and Na_2O abundances increase, but show
36
37 considerable scatter at $\text{MgO} < 10$ wt%; CaO decreases, whereas $\text{Fe}_2\text{O}_3(\text{T})$ is relatively
38
39 constant (Fig. 6). The data suggest a role for both olivine and clinopyroxene fractionation
40
41 from the most primitive magmas ($\text{MgO}>6$ wt%). Whole-rock K_2O contents are relatively
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43 constant at about 6-8 wt%, regardless of MgO variations, suggesting buffering by phlogopite
44
45 (or amphibole) during partial melting of the mantle source. This is consistent with the
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47 presence of phlogopite phenocrysts in the ultrapotassic magmatic rocks (Table 2 and Fig. 3e).
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55 **Trace element characteristics**

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57 The studied samples have light REE (LREE) enriched but relatively flat heavy REE
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2 (HREE) patterns and distinct negative Eu anomalies (Fig. 7). The REE patterns have a very
3
4 distinctive inverted “spoon shape” which appears to be typical of collision-related
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6 ultrapotassic magmas (Davidson *et al.*, 2013). Primitive mantle - normalized incompatible
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8 element patterns (Fig. 8) are distinguished by significantly negative Nb-Ta-Ti and Sr-Ba-P
9
10 anomalies and positive anomalies in large ion lithophile elements (LILE) (e.g. Rb, Pb, U and
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12 Th), despite the high concentrations of Nb, Ta and Ti (Table 3). One possible interpretation
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14 for the significant Nb-Ta-Ti anomalies would be the presence of residual HFSE-rich mineral
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16 phases (e.g. titanite, rutile) in the mantle source because of their high partition coefficients for
17
18 the HFSE. Similar trace element patterns have been observed in lamproites from SE Spain
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20 (Nixon *et al.*, 1984; Contini *et al.*, 1993; Prelević *et al.*, 2008; Conticelli *et al.*, 2009) and
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22 central Italy (Rogers *et al.*, 1985; Gasperini *et al.*, 2002; Perini *et al.*, 2004; Owen, 2008;
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24 Boari *et al.*, 2009; Conticelli *et al.*, 2009) which formed during the same Alpine-Himalayan
25
26 orogenic cycle. A positive correlation between La/Yb and La (Fig. 9) suggests that the degree
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28 of partial melting of the mantle source decreases from west to east.
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37 Mantle-derived magmas produced by partial melting of phlogopite-bearing source
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39 regions are higher in Rb/Sr and lower in Ba/Rb compared to those in equilibrium with
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41 amphibole-bearing sources (Furman & Graham, 1999). The whole-rock data indicate that
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43 both phlogopite and amphibole may have been present in the mantle source region (Fig. 10),
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45 but that phlogopite dominates.
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50 The studied mafic ultrapotassic rocks have flatter LREE patterns in the west than those in
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52 the east of the Lhasa terrane (Fig. 7). In addition, some trace element ratios (e.g. Th/La,
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54 Sm/La, Ba/La, Th/Yb, Th/Nd) display a decreasing trend from west to east (Fig. 11).
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Sr-Nd-Pb isotopes

The ultrapotassic mafic magmatic rocks have extremely radiogenic ($^{87}\text{Sr}/^{86}\text{Sr}$)_i (0.712379-0.737616) and unradiogenic ($^{143}\text{Nd}/^{144}\text{Nd}$)_i (0.511662-0.511984) compositions relative to Bulk Silicate Earth, and high ($^{207}\text{Pb}/^{204}\text{Pb}$)_i (15.65-15.87) and ($^{208}\text{Pb}/^{204}\text{Pb}$)_i (39.02-39.76) at a given ($^{206}\text{Pb}/^{204}\text{Pb}$)_i (18.30-18.92) compared to the Northern Hemisphere Reference Line (NHRL; Hart, 1984) (Fig. 12). The data exhibit coherent trends between depleted MORB-source mantle (DMM) and Indian continental crust [proxied by the isotopic composition of the Higher Himalayan Crystalline Series (HHCS) (Pan *et al.*, 2004; Richards *et al.*, 2005; Guo *et al.*, 2013); see Pan *et al.* (2004) and Richards *et al.* (2005) and references therein for more detailed discussion] in the Sr-Nd-Pb isotope profiles (Fig. 12). There is a trend of increasing ($^{143}\text{Nd}/^{144}\text{Nd}$)_i and decreasing ($^{87}\text{Sr}/^{86}\text{Sr}$)_i and ($^{206}\text{Pb}/^{204}\text{Pb}$)_i from west to east (Fig. 11). The Sr-Nd-Pb isotope compositions are similar to those of mafic ultrapotassic magmatic rocks from SE Spain (Nelson, 1992; Prelević *et al.*, 2008; Conticelli *et al.*, 2009) and central Italy (Gasperini *et al.*, 2002; Perini *et al.*, 2004; Owen, 2008; Boari *et al.*, 2009; Conticelli *et al.*, 2009) in Europe (Fig. 12), although with more radiogenic Sr isotope compositions.

Olivine compositions

A positive correlation between olivine Fo content (mole %) and whole-rock Mg-number indicates that some of the olivine phenocrysts from the ultrapotassic mafic magmatic rocks are in equilibrium with the host magma, proxied by the bulk-rock composition (Fig. 13), based on the equilibrium Fe-Mg exchange coefficient ($K_D=0.3 \pm 0.03$) between olivine and melt (Roeder & Emslie, 1970). Other olivine phenocrysts, however, have much lower Fo

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2 contents which indicates crystallization from more evolved melts. The compositions of those
3
4 olivine phenocrysts which are in equilibrium with their host melts (bulk-rocks) may be used
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6
7 to provide constraints on the nature of the mantle source of the magmas (Foley *et al.*, 2013).
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10 Previous studies (e.g. Prelevic *et al.*, 2013) have indicated that mantle-derived olivine
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12 xenocrysts are characterized by low CaO contents (less than 0.1 wt%). In the Lhasa terrane
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14 ultrapotassic rocks the olivine xenocrysts exhibit distinct reaction rims with abrupt
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17 compositional changes between core and rim (Fig. 14), suggesting disequilibrium between the
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19 olivine xenocrysts and the host magma. The mantle olivine xenocrysts have high Fo [Fo
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21 (mole%)~ 90] and low CaO (less than 0.1 wt%) contents in their cores (Fig. 14c), mantled by
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23
24 low Fo (less than 80) and high CaO (more than 0.15 wt %) rims (Fig. 14c).
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27 The olivine phenocrysts have higher CaO (more than 0.1%) and MnO contents compared
28
29 to the xenocrysts; Ni contents are more variable (Fig. 15). The olivine xenocrysts plot within
30
31 a very restricted compositional range (CaO=0.06~0.10 wt%, NiO=0.31~0.51 wt%,
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33 MnO=0.10~0.21 wt%). Some of the olivine phenocrysts in the most primitive ultrapotassic
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36 magmas (e.g. samples MG1303, SLP1314; Table 3) are characterized by extremely high NiO
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38 contents (0.60-0.70 wt %) (Fig. 15c). This, together with low MnO and CaO contents (Figs.
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41 15a,b), is consistent with an inference of pyroxenite as the mantle source of the host melts
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44 from which the olivine crystallized, as proposed in a number of recent studies (e.g. Sobolev *et*
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47 *al.*, 2005, 2007; Herzberg, 2011; Prelevic *et al.*, 2013).
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51 52 53 **DISCUSSION**

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58 Previous studies have shown that the post-collisional potassium-rich magmatism of the
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3 Tibetan Plateau can provide important constraints for understanding the geodynamic setting
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5 and evolution of the Plateau (e.g. Coulon *et al.*, 1986; Pearce & Mei, 1988; Arnaud *et al.*,
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7 1992; Turner *et al.*, 1993, 1996; Miller *et al.*, 1999; Williams *et al.*, 2001, 2004; Ding *et al.*,
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9 2003; Nomade *et al.*, 2004; Mo *et al.*, 2006; Gao *et al.*, 2007; Zhao *et al.*, 2009; Chen *et al.*,
10
11 2010; Wang *et al.*, 2012; Guo *et al.*, 2013, 2014). Primary magmas generated by equilibrium
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13 partial melting of peridotite in the upper mantle should have high Mg-numbers (more than 70)
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15 and high Ni concentrations (Frey *et al.*, 1978). The magmatic rocks in our study are
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17 considered to be candidates for primitive magmas because of their high MgO contents
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19 (6.28-15.75 wt%), high Mg-numbers (70-88), high Ni concentrations (106-527 ppm) (Table 3),
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21 and the presence of mantle-derived olivine xenocrysts (Figs. 3 and 14) and rare mantle
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23 xenoliths (e.g. Liu *et al.*, 2011, 2014a). Thus, their whole-rock major and trace element and
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25 Sr-Nd-Pb isotope compositions can be used to provide important constraints on the nature of
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27 their mantle source.
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38 **Characteristics of the mantle source region of the primitive magmas**

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40 The ultrapotassic mafic magmatic rocks studied have distinct negative Nb-Ta-Ti
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42 anomalies and positive Pb anomalies (Fig. 8), consistent with a subduction-related origin.
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44 Relatively high contents of LILE and LREE (Figs. 7 and 8) and high ($^{87}\text{Sr}/^{86}\text{Sr}$)_i and low
45
46 ($^{143}\text{Nd}/^{144}\text{Nd}$)_i (Fig. 12) indicate that their parental magmas were derived from an enriched
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48 mantle source. Additionally, their Sr-Nd-Pb isotope compositions exhibit coherent trends
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50 between likely ambient asthenospheric mantle (MORB-source mantle) and Indian continental
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52 crust (Fig. 12), consistent with a subduction-related origin for the mantle source enrichment
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54 related to addition of subducted Indian continental crustally-derived materials to their mantle
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2 source. The presence of a northward subducted Indian continental lithosphere slab beneath
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4 the Lhasa terrane in south Tibet, identified in a number of geophysical studies (e.g. Li *et al.*,
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6 2008; Nabelek *et al.*, 2009, Zhao *et al.*, 2010, 2011; Chen *et al.*, 2015), supports this
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8 inference. The decreases in Sr and Pb isotope initial ratios (Figs. 11 and 12) in the
9
10 ultrapotassic magmatic rocks from west to east may suggest decreasing involvement of the
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12 crustal component towards the east.
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17 Nabelek *et al.* (2009) constructed a high resolution, 800 km long, N-S cross section of
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19 the lithosphere through the Lhasa terrane based on seismic data obtained as part of the
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21 Hi-CLIMB seismic experiment (Fig.1). They identified a flat lying, ~ 15 km thick layer
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23 above the seismic Moho which they interpreted as subducted Indian lower continental crust.
24
25 Such subducted Indian lower crust does not appear to extend further north than 31 °N beneath
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27 Tibet. They suggested that this form of flat slab subduction must have developed in less than
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29 20-25 Myr, based upon current rates of underthrusting of India beneath the Himalayas, and
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31 that before this Indian plate subduction was much steeper. Using the same seismic data set
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33 Wittlinger *et al.* (2009) constrained the characteristics of this Indian lower crustal layer
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35 (thickness 19.3 ± 2.4 km, V_s 4.73 ± 0.27 kms⁻¹, κ 1.69 ± 0.06) which they attributed to
36
37 eclogite facies rocks. Schulte-Pelkum *et al.* (2005) demonstrated fast P-wave velocities of > 7
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39 kms⁻¹ in the lower crust of this region which they attributed to partial eclogitisation (~30%).
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41 Wittlinger *et al.* (2009) located the base of this high grade metamorphic layer at ~ 80 km
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43 depth and showed that it extends up to 150 km north of the ITS (~ 31°N) and more than 1000
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45 km laterally beneath the Lhasa terrane.
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55 The seismic cross-section of Nabelek *et al.* (2009) may provide one of the best images
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57 we have of the present configuration of subduction beneath the Lhasa terrane. The
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1
2 geophysical recognition of an eclogite facies crustal layer within the subducted Indian
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4 continental slab provides further support for our contention that the enrichment of the source
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6 of the ultrapotassic magmas originated from mixing between subducted Indian continental
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8 crustal materials and mantle wedge peridotite. The spatial distribution of the eclogite facies
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10 lower crustal layer overlaps with that of the post-collisional potassic and ultrapotassic
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12 magmatic rocks in the Lhasa terrane (Fig. 1).
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17 Rogers *et al.* (1985) proposed that abundances of the first row transition elements in
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19 post-collisional mafic ultrapotassic and potassic magmas are not affected by metasomatic
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21 enrichment of their mantle source region and may thus reflect the original protolith.
22
23 Consequently, we use these transition elements to constrain the compositional characteristics
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25 of the mantle source region of the Tibetan ultrapotassic mafic magmas prior to its subsequent
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27 enrichment. Chondrite-normalised patterns of the first row transition elements (Fig. 16) are
28
29 comparable with those of MORB, suggesting a DMM-like mantle protolith prior to its
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31 continental subduction-related enrichment. Moreover, the composition of the mantle source
32
33 region of the ultrapotassic mafic magmas prior to this enrichment appears to be constant (i.e.
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35 MORB-source mantle, DMM) from west to east in south Tibet. This confirms that the
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37 decreases in Sr and Pb isotope initial ratios and trace-element ratios (Figs. 11 and 12) of the
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39 magmas from west to east must result from a decreasing crustal component in the mantle
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41 source region of the magmas towards the east.
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50 To constrain the composition of the crustal component in the mantle source region of the
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52 post-collisional mafic K-rich magmas we corrected the major element compositions of the
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54 most primitive mafic magmatic rocks to Mg-number $[100\text{Mg}/(\text{Mg}+\text{Fe}^{2+})]=72$ for those
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56 samples with Mg-numbers less than 72, using an olivine-melt K_D of 0.30 (cf Roeder & Emslie,
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1
2 1970), based on the approach of Niu *et al.* (1999) and Humphreys & Niu (2009). The
3
4 compositional correction allows comparison of the major element compositions normalized to
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6 the same Mg-number (=72) for the different volcanic fields in south Tibet, thus compensating
7
8 for any effects of olivine fractionation and/or accumulation on the whole-rock composition of
9
10 the magmatic rocks. Compared to MORB the K-rich magmatic rocks in south Tibet are
11
12 characterized by higher SiO₂ and Ni contents, higher Sr and Pb isotope ratios, and lower CaO
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14 and MgO contents and Nd isotope ratios (Fig. 17). The correlations between Sr-Nd-Pb isotope
15
16 and major element compositions indicate the presence of a component in the mantle source of
17
18 the K-rich magmas which is enriched in SiO₂, Al₂O₃, Ni, ⁸⁷Sr/⁸⁶Sr, and depleted in CaO, MnO
19
20 and MgO. This component is consistent with the composition of a partial melt of eclogite
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22 facies continental crust, generated from the crustal layer at the top of the subducted Indian
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24 continental lithosphere. This melt would have infiltrated the overlying mantle wedge causing
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26 metasomatic enrichment. This inference is consistent with seismological observations of the
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28 presence of eclogite facies crustal materials on the top of the subducted Indian continental
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30 lithospheric slab beneath south Tibet (e.g. Nabelek *et al.*, 2009; Chen *et al.*, 2015).
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43 *A pyroxenite source for the magmas*

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45 A number of recent studies (Sobolev *et al.*, 2005, 2007; Herzberg, 2011; Prelevic *et al.*,
46
47 2013) have suggested that partial melts of subducted eclogite facies crustal rocks may react
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49 with surrounding mantle peridotite to yield pyroxenite via reaction of olivine to form
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51 pyroxene. Partial melts derived from subsequent melting of such pyroxenites are
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53 characterized by higher Ni and SiO₂, and lower MgO contents, compared with melts of upper
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55 mantle peridotite (Sobolev *et al.*, 2005, 2007; Straub *et al.*, 2008; Herzberg, 2011). The
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compositional characteristics (e.g. high SiO₂ and Ni concentrations and relatively low MgO and CaO contents) of the mafic K-rich magmas in south Tibet are consistent with those of partial melts produced by pyroxenite melting (Fig. 18).

Olivine phenocrysts crystallized from partial melts of such mantle pyroxenites have lower Ca, Mn and Mg concentrations compared with those crystallized from melts of upper-mantle peridotite (Sobolev *et al.*, 2005, 2007; Herzberg, 2011). The olivine phenocrysts in the mafic K-rich magmas in south Tibet display similar compositional characteristics to those crystallized from partial melts of pyroxenite (Fig.19).

Two types of pyroxenite can be recognized in the mantle source of the mafic K-rich magmas of the Lhasa terrane based upon the compositional characteristics of their olivine phenocrysts: (1) olivine-free pyroxenite (shown by the black-blue-green symbols in Fig. 19) and (2) olivine-bearing pyroxenite (shown by the red symbols in Fig. 19). Olivine phenocrysts crystallized from partial melts of olivine-free pyroxenite are lower in Mg and Ca, higher in Mn, and lower in Fe/Mn compared to those crystallized from partial melts of olivine-bearing pyroxenite. The host melts produced from the olivine-bearing pyroxenite are higher in MgO (14-15.8 wt%) and lower in SiO₂ (46-47 wt%) (e.g. samples MG1303 and SLP1314 in Table 3) compared to those derived from the olivine-free pyroxenite (Fig. 18). However, partial melts derived from these two types of pyroxenite appear to show semi-continuous compositional variation trends (Fig. 18). The formation of the olivine-free pyroxenite in the mantle source region requires a higher proportion of partial melt from subducted Indian eclogite facies crust reacting with a peridotite protolith compared with the olivine-bearing pyroxenite.

Recent studies (e.g. Sobolev *et al.*, 2005, 2007; Straub *et al.*, 2008; Prelevic *et al.*, 2013)

1
2 indicate that abundances of Ni in upper mantle peridotite-derived primary magmas are mainly
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4 governed by olivine, because this mineral has the highest solid-melt partition coefficient for
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6 Ni (e.g. $Kd_{Ni}^{olivine}=10$; Straub *et al.*, 2008). The extremely high Ni concentrations in the
7
8 olivine phenocrysts of the mafic K-rich magmas in south Tibet may be interpreted in terms of
9
10 replacement of olivine in their mantle source by orthopyroxene through reaction with partial
11
12 melts of subducted Indian crustal materials. The Ni previously preserved in the olivine of the
13
14 peridotite protolith must now be contained in orthopyroxene. The orthopyroxene in the
15
16 reaction-formed pyroxenite has a lower solid-melt partition coefficient for Ni (e.g.
17
18 $Kd_{Ni}^{orthopyroxene}=2$; Straub *et al.*, 2008; Prelevic *et al.*, 2013) compared with the original olivine
19
20 in the peridotite, suggesting that the pyroxenite-derived primary magmas would have about
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22 five times the Ni concentration compared to melts derived from peridotite. Olivine
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24 phenocrysts crystallized from such Ni rich primary magmas would have also high Ni
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26 concentrations.
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35 The olivine-melt partition coefficients for Ni ($Kd_{Ni}^{olivine}$) in low MgO and high SiO₂
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37 magmas are significantly higher than those in melts with MgO>10 wt % (e.g. Kinzler *et al.*,
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39 1990; Beattie *et al.*, 1991; Straub *et al.*, 2008). Because magmas derived from pyroxenite
40
41 melting are lower in MgO and higher in SiO₂ than those formed from peridotite melting (Fig.
42
43 18), the olivine-melt partition coefficient for Ni ($Kd_{Ni}^{olivine}$) during pyroxenite melting will be
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45 significantly higher (up to 35; Beattie *et al.*, 1991; Straub *et al.*, 2008). Consequently, olivine
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47 phenocrysts crystallized from the Ni-rich, SiO₂-rich and MgO-poor magmas resulting from
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49 pyroxenite melting should have extremely high Ni contents.
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55 The compositions of the olivine phenocrysts in the Lhasa terrane ultrapotassic magmatic
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57 rocks appear to define three distinct curved trends (B1, B2 and B3) in a plot of Ni content in
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2
3 olivine versus Fo (mole %) (Fig. 20), shown by the black, blue and red symbols, respectively.
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5 They display concave down trends. Previous studies (e.g. Prelevic *et al.*, 2013; Straub *et al.*,
6
7 2008) have indicated that such concave down trends may be explained by fractional
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9 crystallization of olivine from pyroxenite-derived parent magmas. Modelled variations in
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11 olivine composition induced by fractional crystallization of the host melt (see Supplementary
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13 Data Appendix B for more details of the modeling calculation) fit the B1, B2 and B3 trends of
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15 the olivine phenocrysts very well (Fig. 20). The different compositions of the parent magmas
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17 of each trend are interpreted to reflect varying proportions of eclogite partial melt involved in
18
19 the formation of their pyroxenite sources. The B1 and B2 magma sources are olivine-poor to
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21 olivine-free pyroxenites whereas the B3 source is an olivine-rich pyroxenite. The proportion
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23 of the eclogite melt component in the pyroxenite source increases from B3 to B1 (shown by
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25 the blue arrow in Fig. 20).
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33 The transition element Ni is enriched in the upper mantle, whereas it is strongly depleted
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35 in the continental and oceanic crust (e.g. Rudnick & Gao, 2002). Moreover, Ni is neither
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37 fluid-mobile, nor mobile in low-MgO melts of continental crust (e.g. Straub *et al.*, 2008).
38
39 Thus, siliceous melts from subducted Indian eclogite facies continental crust would have low
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41 Ni contents. Olivine phenocrysts defining the compositional trend B1 have lower Fo contents
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43 than the B2 and B3 trends, but similarly high Ni concentrations in the most forsteritic olivines
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45 (olivine composition field labeled “A” in Fig. 20), which can only reasonably be interpreted
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47 in terms of a greater amount of melt derived from subducted Indian eclogite facies continental
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49 crustal materials (as shown by the blue arrow in Fig.20) in the generation of their pyroxenite
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51 source, compared to those of the B2 and B3 trends. We subsequently refer to their source as a
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53 SiO₂-rich pyroxenite.
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3 The olivine phenocrysts which define the B3 trend are consistent with those crystallized
4 from melts of olivine-bearing pyroxenite (Fig. 19). However, they could also reflect
5 crystallization from mixtures of melts of olivine-free pyroxenite (“R” in Fig. 20) and melts of
6 peridotite (“P” in Fig. 20). The corroded rims of the olivine phenocrysts (Fig. 3b) and reaction
7 rims on xenocrystic olivine (Fig. 3a,c) are consistent with mixing of melts from such a
8 hybridized peridotite-pyroxenite source. The hybrid melts have the highest MgO (14-15.8
9 wt%), moderate Ni (272-452 ppm) and relatively low SiO₂ contents (46-47 wt%) (e.g.
10 samples MG1303 and SLP1314 in Table 3). These samples may indicate a more important
11 contribution for peridotite in the mantle source.
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25 The correlation between the Ni and Fo contents of olivine phenocrysts (Fig. 20) appears
26 to be highly sensitive in distinguishing between different mantle source regions and their
27 partial melts. Three distinct source components are recognized from west to east in south
28 Tibet: SiO₂-rich pyroxenite (compositional trend B1 in Fig. 20), olivine-free pyroxenite
29 (compositional trend B2 in Fig. 20) and olivine-bearing pyroxenite (compositional trend B3 in
30 Fig. 20). In detail, the mantle source region of the magmas in the western and central
31 sub-groups (Sailipu, Xungba, Zabuye and Maiga volcanic fields; Figs. 1 and 20) contains all
32 three types of pyroxenite, whereas that in the eastern sub-group (Mibale volcanic field; Figs. 1
33 and 20) only contains olivine-bearing pyroxenite. In Fig. 21 we illustrate our understanding of
34 the changing characteristics of the mantle source of the ultrapotassic magmas from west to
35 east in the Lhasa terrane.
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53 On the basis of the above discussion we propose that the mantle source region of the
54 post-collisional mafic K-rich magmas in south Tibet has been metasomatised by partial melts
55 of subducted Indian eclogitic facies continental crust which infiltrated the peridotite mantle
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1
2 wedge. These partial melts reacted to varying degrees with the surrounding peridotite to yield
3
4 olivine-bearing pyroxenite, olivine-free pyroxenite and SiO₂-rich pyroxenite, subsequent
5
6 partial melting of which formed the most primitive K-rich mafic magmas in south Tibet (Figs.
7
8 20 and 21). Partial melting of olivine-bearing pyroxenites resulted in more SiO₂-poor and
9
10 higher MgO magmas in the eastern sub-group volcanic fields (Fig. 21). An increasing
11
12 proportion of subducted Indian crust in the mantle source of the magmas westward resulted in
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14 more SiO₂-rich, K-rich magmas in the central-western sub-group volcanic fields (e.g. samples
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16 MG1301 in the Maiga volcanic field; Figs. 20 and 21).
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25 ***Trace element and Sr-Nd-Pb isotope indicators of the role of subducted continental crust***

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27 Previous studies (e.g. Tommasini *et al.*, 2011; Prelevic *et al.*, 2013) have suggested that
28
29 Th (and Sm) enrichment in K-rich magmas may be attributed to the presence of continental
30
31 crustal material in their mantle source. Zoisite-epidote and lawsonite in continentally-derived
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33 metasedimentary materials (products of blueschist facies metamorphism of subducted
34
35 sediments) are likely repositories of Sr, Pb, U, Th, and LREE (e.g. Brunsmann *et al.*, 2000;
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37 Spandler *et al.*, 2003; Frei *et al.*, 2004; Usui *et al.*, 2006; Feineman *et al.*, 2007; Tommasini *et*
38
39 *al.*, 2011) based on their high partition coefficients for these elements. Partial melting of
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41 metasediments containing zoisite-epidote and lawsonite could thus lead to Th (and Sm)
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43 enrichment of the source of the K-rich magmas. The ultrapotassic mafic magmatic rocks of
44
45 south Tibet exhibit such Th and Sm enrichments (Figs. 8 and 22a).
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53 Enrichment in LILE and the Sr-Nd-Pb isotope compositions of subduction-related
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55 magmas have been widely ascribed to the migration of fluids or melts from the subducted
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57 slab into their mantle source (e.g. Gill, 1981; Pearce, 1982; Hawkesworth *et al.*, 1997;
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1
2 Johnson & Plank, 1999). Geochemical distinctions have been recognized between
3
4 subduction-related magmatic rocks whose mantle sources have been modified by fluids and
5
6 those that have been enriched by slab-derived melts (e.g. Hawkesworth *et al.*, 1997; Johnson
7
8 & Plank, 1999; Class *et al.*, 2000; Woodhead *et al.*, 2001; Guo *et al.*, 2014). Slab-derived
9
10 fluids, which carry little REE and HFSE, can introduce significant amounts of LILE (e.g. K,
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12 Rb, Cs, Sr, Ba, U, Pb) from the subducting slab into the mantle source, whilst slab-derived
13
14 melts are characterized by high Th and LREE contents.
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20 We use the ratio of fluid-mobile Ba to fluid-immobile La to investigate the input of
21
22 subduction-related fluids and Th/Nd to reflect the input of slab-derived melts because neither
23
24 Ba/La nor Th/Nd is likely to be fractionated during partial melting of the mantle source.
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26 Within the Lhasa terrane the eastward decreases in Ba/La and Th/Nd in the post-collisional
27
28 K-rich mafic magmatic rocks (Figs. 11d,e) could be explained by a decreasing flux of
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30 subducted Indian continental crust derived fluid and melt components into their mantle
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32 source region from west to east. This is consistent with the conclusions reached above based
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34 on the composition of olivine phenocrysts.
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40 The ultrapotassic mafic magmatic rocks define a diffuse linear trend in a ($^{143}\text{Nd}/^{144}\text{Nd}$)_i
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42 versus Th/La diagram (Fig. 22b), plotting between depleted MORB-source mantle (DMM)
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44 and an inferred Indian continental crustal sediment melt component with low ($^{143}\text{Nd}/^{144}\text{Nd}$)_i
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46 and high Th/La, suggesting that their mantle source region comprises two end-members: (1)
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48 depleted MORB-source mantle (cf., Workman & Hart, 2005) and (2) a melt derived from
49
50 subducted Indian continental crust with ($^{143}\text{Nd}/^{144}\text{Nd}$)_i < 0.5117 and high Th/La (>6). Because
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52 there is likely to be a wide variation of Th/La in the subducted Indian continental crustal
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54 component (Inger & Harris, 1993; Ahmad *et al.*, 2000; Richards *et al.*, 2005; Rehman *et al.*,
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2008; Guo & Wilson, 2012), this can explain the dispersion in the data in Fig. 22b. Similarly, the ultrapotassic rocks display a negative correlation of $(^{143}\text{Nd}/^{144}\text{Nd})_i$ versus Ba/La (Fig. 22c) plotting between a depleted MORB-source mantle (DMM) component and an Indian slab-derived fluid with low $(^{143}\text{Nd}/^{144}\text{Nd})_i$ and high Ba/La. This suggests the presence of an Indian slab-derived fluid component in the mantle source region. The decreases in some trace element ratios (e.g. Th/La, Ba/La; Figs. 11 and 22) and Sr-Pb initial isotope compositions (Fig. 11), together with increasing $(^{143}\text{Nd}/^{144}\text{Nd})_i$, from west to east (Fig. 11), suggest an eastward decrease in the proportion of the Indian continental crustal components (including sediment melt and aqueous fluid) in the mantle source of the mafic K-rich magmas.

Dy/Dy* variations and Eu anomalies

Davidson *et al.* (2013) have recently proposed the use of the parameter Dy/Dy* in the recognition of mantle source components and petrogenetic processes in the generation of subduction-related magmas. We have adapted their approach to show the role of subducted Indian continental crust in the petrogenesis of the post-collisional mafic ultrapotassic magmas in the Lhasa terrane (Fig. 22d). The relatively high Dy/Yb ratios of the magmatic rocks, combined with Dy/Dy* values less than 1.0, are distinct from those of MORB, OIB and arc magmas. We interpret this distinctive signature to reflect the presence of an eclogite facies crustal component derived from Indian continental lithosphere subduction in the mantle source of the ultrapotassic mafic magmas in south Tibet. The decreasing Dy/Dy* values of the magmas from west to east, combined with an increase in Dy/Yb ratios (Fig. 22d), may indicate a decrease in the degree of partial melting of the mantle source toward the east. The Dy/Yb ratios are higher in the western and central subgroup volcanic fields than in the eastern

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2 subgroup volcanic fields (Fig. 22d), suggesting that the proportion of the eclogite facies
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5 crustal component derived from Indian continental crust subduction in the source of the
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8 magmas decreases from west to east.

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10 The most primitive ultrapotassic rocks exhibit negative Eu anomalies in their
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12 chondrite-normalized REE patterns (Fig. 7). Whilst such negative Eu anomalies are typically
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14 attributed to plagioclase feldspar fractionation (or to the presence of residual plagioclase in
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16 the magma source) we consider that negative Eu anomalies in the Tibetan ultrapotassic
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18 magmatic rocks are an intrinsic feature inherited from their pyroxenite mantle source.

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20 The diffuse negative correlation between $^{143}\text{Nd}/^{144}\text{Nd}$ and Dy/Dy* (Fig. 22e) may also
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22 indicate that the proportion of the eclogite facies crustal component in the mantle source of
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24 the magmas decreases from west to east. The role of the Indian continental crustal component
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26 in the source of the magmas is also clearly demonstrated in a plot of Ce/Pb versus Nb/La (Fig.
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22f).

Role of a subduction mélangé

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41 Marschall & Schumacher (2012) have proposed a physical model by which components
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43 from a subducting slab and its overlying mantle wedge can be transported into the mantle
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45 source of subduction-related magmas. This involves formation of a mélangé zone in a
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47 subduction channel on the top surface of the slab in which hydrated mantle rocks are mixed
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49 with material derived from the subducting slab, including trench sediments. The
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51 trace-element characteristics of exhumed metamorphic mélangé rocks are similar to those of
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53 subduction-related magmas (Marschall & Schumacher, 2012), with distinctive enrichment in
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55 LILE and significant depletion in HFSE (e.g. Nb, Ta and Ti), indicating that subducted
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3 mélange rocks in the subduction channel could provide an important source component for
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5 arc magmas. Partial melting of well-mixed mélange rocks underplated below the base of the
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7 lithosphere via diapiric upwelling from the subduction channel is considered to generate
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9 subduction-related magmas in the Marschall & Schumacher (2012) model. Geophysical
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11 imaging of such underplated mélanges would clearly be extremely helpful in validating such
12
13 a model. We propose that the eclogite facies lower crustal layer on the top surface of the
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15 subducted Indian continental slab, recognized by seismic tomography (e.g. Nabelek *et al.*,
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17 2009), might be the geophysical image of the final stage in the evolution of such a subduction
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19 channel beneath the Lhasa terrane, in which the mantle lithosphere of the Tibetan plate has
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21 effectively been removed by flat slab subduction. The seismic data clearly indicate the
22
23 presence of eclogite-granulite facies rocks in the crustal layer of the subducted Indian
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25 lithosphere. The sub-continental lithospheric mantle (SCLM) removed from the base of the
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27 Tibetan lithosphere during flat slab subduction likely represents the original mantle protolith
28
29 involved in the petrogenesis of the ultrapotassic magmas. The similarity of this protolith to
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31 MORB-source mantle (Fig. 16) suggests that it may have been relatively recently accreted
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33 rather than ancient sub-continental lithospheric mantle. The termination of ultrapotassic
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35 magmatism at ca 8Ma could reflect the timing of complete erosion of the Tibetan lithospheric
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37 mantle.
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48 The primitive mantle-normalized incompatible trace element pattern of average
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50 subduction channel mélange rocks reported by Marschall & Schumacher (2012) is similar to
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52 those of the post-collisional ultrapotassic mafic magmas in south Tibet, characterized by
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54 enrichment in LILE and significant depletion in Nb, Ta and Ti (Fig. 8), although overall
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56 abundances are lower in the mélange rocks. In addition, the heavy REE abundances, together
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2 with some trace element ratios (e.g. Ce/Pb versus Nb/La; Fig. 22f), are similar between
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4 average subduction channel mélangé rocks (Marschall & Schumacher, 2012) and the
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6 ultrapotassic magmas in south Tibet (Fig. 7).
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10 11 ***Quantifying the proportion of the subducted Indian crustal component***

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15 Based on our discussions of the mineral chemistry of olivine (Figs. 15, 19, 20) and
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17 whole-rock geochemistry (Figs. 11, 12, 17, 22), together with the evidence from seismic
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19 tomography (e.g. Nabelek *et al.*, 2009), we conclude that mantle source region of the
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21 ultrapotassic magmas comprises two components: (1) depleted MORB-source mantle (DMM)
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23 and (2) an subducted Indian eclogite facies continental crustal component.
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28 We have attempted to quantify the proportions of the subducted Indian eclogite facies
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30 crustal component which contributes to the mantle source of the magmas from west to east in
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32 the Lhasa terrane, following the approach used in our earlier simulations (Guo *et al.*, 2005,
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34 2006; Guo & Wilson, 2012). The modeling is subdivided into two steps: (1) simulation of
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36 partial melting of a subducted Indian eclogite facies continental crustal component; (2)
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38 two-component mixing between this eclogite facies continental crustal partial melt and the
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40 bulk peridotite of the pre-existing mantle wedge (i.e. depleted MORB-source mantle). Such a
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42 bulk-mixing scenario is clearly an over-simplification of the actual processes involved.
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48 We selected the bulk-rock composition of an Indian eclogite from Rehman *et al.* (2008)
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50 as a proxy for the composition of the subducted continental crustal component in our
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52 modeling simulations. Considering the constraints on the compositional characteristics of the
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54 pre-existing mantle protolith of the source of the mafic K-rich magmas (Fig. 16), we used the
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56 composition of depleted MORB-source mantle (DMM; Workman & Hart, 2005) as a proxy
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2 for the mantle protolith. The mineral-melt partition coefficients used are given in Table 6. The
3
4 model melting curve for partial melting of the eclogite (the red solid curve with filled rhombs)
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6 and the DMM composition are shown in a plot of Dy/Yb versus La/Yb (Fig. 23). The Dy/Yb
7
8 ratio of the eclogite melt is approximately constant (~ 7) for degrees of melting up to 50%.
9
10 Two-component mixing between the eclogite-derived crustal partial melt and the pre-existing
11
12 bulk mantle protolith (DMM) are shown by the black dashed lines with tick marks labeled
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14 *M1*, *M2* and *M3* (Fig. 23). The results of the modeling calculation are summarized in Tables 7
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16 and 8.
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22 The results of modeling the Dy/Yb vs La/Yb variations suggest that the whole-rock REE
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24 compositions of the ultrapotassic magmas could be explained by derivation from a mantle
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26 source produced by mixing between an eclogite facies crust-derived melt and a pre-existing
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28 bulk mantle peridotite protolith (depleted MORB-source mantle) (Fig. 23), assuming that the
29
30 Dy/Yb and La/Yb ratios of the ultrapotassic magmas are not significantly fractionated from
31
32 those of their source. The degree of melting of the eclogite component in the western-central
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34 subgroup volcanic fields is 10 to 20%, whereas in the eastern subgroup volcanic field it is 2 %
35
36 (Fig. 23). The proportions of the eclogite-derived melt in the mixed source is 10-50% for the
37
38 magmas of the western and central subgroup volcanic fields and 5-25% for the eastern
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40 subgroup volcanic field (Fig. 23). This is consistent with our inference of a decrease in the
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42 amount of the subducted Indian eclogite crustal component in the mantle source region
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44 eastward in south Tibet, as indicated by the composition of olivine phenocrysts (Figs. 20, 21).
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55 **Petrogenesis of the ultrapotassic mafic magmas in south Tibet**

56 The mineral chemistry of olivine phenocrysts combined with whole-rock geochemical
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2 and Sr-Nd-Pb isotopic data suggest that the post-collisional ultrapotassic magmas in the
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4 Lhasa terrane are the melting products of partial melting of pyroxenites which were formed
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6 by reaction of partial melts and fluids derived from northward subducted Indian eclogite
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8 facies continental crustal materials with mantle peridotite which likely resided within the
9
10 overlying mantle wedge, including the Tibetan sub-continental lithospheric mantle (SCLM).
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12 The significant Ta-Nb-Ti troughs and enrichment in LILE in the mantle-normalised trace
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14 element patterns of the most primitive ultrapotassic magmatic rocks (Fig. 8), combined with
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16 their similar Ce/Pb and Nb/La ratios to those in average subduction channel mélanges (Fig.
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18 22f), further support their relationship to continental subduction. Previous studies (e.g. Li *et*
19
20 *al.*, 2008; Zhao *et al.*, 2010; Zhao *et al.*, 2011; Chen *et al.*, 2015) suggested that onset of the
21
22 northward underthrusting of Indian continental lithosphere beneath south Tibet immediately
23
24 followed India-Asia continent-continent collision at ~ 55Ma. However, the age (25-8Ma) of
25
26 the post-collisional K-rich magmatism in south Tibet (Table 1 and Fig. 2) post-dates
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28 India-Asia collision by about 30 Myr. The clear delay in the onset of K-rich magmatism and
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30 the finite duration of the magmatism (25-8 Ma) implies an existence of other factors in their
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32 petrogenesis in addition to the northward underthrusting of the Indian continental lithospheric
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34 slab. If partial melting of the mantle wedge was directly caused by the northward
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36 underthrusting of the Indian continental lithosphere, this should lead to decreasing ages for
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38 the resultant K-rich magmas from south to north, rather than the southward decreasing ages
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40 which are actually observed (Fig. 2).
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53 On the basis of the above discussion, we conclude that northward underthrusting of the
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55 Indian continental lithosphere since India-Asia collision at ~55 Ma led to metasomatic
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57 enrichment of the overlying mantle wedge (including asthenospheric mantle and SCLM) and
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2 the formation of pyroxenites which provided the two-component (crustal eclogite melt from
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4 the Indian continental slab reacting with MORB-source mantle) source region of the
5
6 post-collisional K-rich magmas in south Tibet. Seismic tomography studies (e.g. Zhou &
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8 Murphy, 2005; Li *et al.*, 2008; Zhao *et al.*, 2010; Zhao *et al.*, 2011; Chen *et al.*, 2015) have
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10 indicated the continuous presence of a northward dipping subducted slab of Indian
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12 continental lithosphere which increases in depth from south to north, suggesting the potential
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14 for continuous formation of pyroxenite in the mantle wedge since the onset of continental
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16 subduction. However, it is less clear as to what geodynamic event or process caused melting
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18 of the pyroxenites in the metasomatised mantle wedge uniquely during the period 25-8Ma.
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20 There must have been a critical geodynamic event between 25 and 8Ma which resulted in the
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22 partial melting of the pyroxenites in the mantle wedge beneath south Tibet. The increasing
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24 age of the K-rich magmas from south to north in the Lhasa terrane (Table 1 and Fig. 2),
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26 combined with a northward decline in components derived from the subducted Indian
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28 continental lithosphere in the mantle source of the K-rich magmas in the eastern sub-group
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30 (Guo *et al.*, 2013), suggests that partial melting of the mantle wedge pyroxenites can be
31
32 explained by the progressive roll-back of the northward underthrusting Indian continental
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34 lithospheric slab during the period 25-8Ma. The decreasing trend in the age of the magmatism
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36 from west to east (Fig. 2b) may be consistent with break-off of the Indian continental slab
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38 that propagated eastwards from 25 Ma to 8 Ma (e.g. Replumaz *et al.*, 2010, 2013, 2014; Chen
39
40 *et al.*, 2015). Such slab break-off combined with slab roll-back during continental lithosphere
41
42 subduction may have triggered asthenospheric mantle upwelling and caused thermal erosion
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44 of the Tibetan SCLM. Detached fragments of the Indian continental lithosphere may have
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46 undergone further melting as they subsided into the asthenosphere.
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SUMMARY

On the basis of the compositions of olivine phenocrysts, whole-rock major element, trace element, Sr-Nd-Pb isotope and seismic tomography data, we propose a two-stage model to account for the petrogenesis of the post-collisional ultrapotassic mafic magmatism during the period 25-8Ma in south Tibet. The first stage (55-25Ma) involved the subduction of eclogite facies continental crustal materials on the top of a subducted Indian continental lithospheric slab beneath south Tibet since India-Asia collision at ~55 Ma; these crustal materials could have undergone fluid release and partial melting, resulting in metasomatism of the overlying mantle wedge, forming pyroxenites. Alternatively the subducted Indian crustal materials may have underplated the base of the Tibetan lithosphere (Fig. 24a) forming a subduction channel *mélange*. At this stage (55-25 Ma) there is no ultrapotassic magmatism within the Lhasa terrane and thus no evidence for pyroxenite melting (or indeed pyroxenite formation).

The second stage of the model (25-8Ma) involves partial melting of pyroxenites caused by reaction between silica-rich melts derived from partial melting of subducted eclogite facies crustal materials and the surrounding mantle peridotite in the mantle wedge and Tibetan SCLM (Fig. 24b). Partial melting of these pyroxenites was likely induced by upwelling of hot asthenospheric mantle in response to Indian slab roll-back and break-off, resulting in the generation of the ultrapotassic magmas (Fig. 24b). Indian slab roll-back has been linked to a sharp decrease in the convergence rate between India and Asia between 25 and 8Ma (Lee & Lawver, 1995). There may be further metasomatism of the Tibetan lithospheric mantle during uprise of the ultrapotassic magmas (Fig. 24b).

The post-collisional ultrapotassic mafic magmatism in south Tibet is older than 8 Ma

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3 (Table 1 and Fig. 2). The absence of ultrapotassic mafic magmatism from 8Ma to the present
4
5 can be explained by flat slab subduction of the Indian continental lithosphere since 8 Ma, as
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7 indicated by the recent geophysical studies (e.g. Lee & Lawver, 1995; Nabelek *et al.*, 2009;
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9 Zhao *et al.*, 2011; Chen *et al.*, 2015). This flat slab subduction has eliminated the mantle
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11 wedge above the subducted Indian lithosphere, preventing the upwelling of hotter
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13 asthenospheric mantle through slab tears (Fig. 24c).
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17 The compositions of olivine phenocrysts in the mafic K-rich magmas in the Lhasa
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19 terrane suggest a progressively increasing participation of peridotite in the mantle source
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21 region from west to east (Fig. 21). Previous studies (e.g. Hirschmann & Stolper, 1996) have
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23 shown that mantle peridotite has a higher temperature solidus and lower melt productivity
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25 relative to pyroxenite at a given pressure. Thus, pyroxenite in the mantle melts preferentially,
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27 and may produce a larger volume of melt. The eastward decrease in lava volumes and partial
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29 melting degree in south Tibet (Figs. 1 and 9) is consistent with a reduction in the contribution
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31 of components from the subducted Indian continental crust to the upper-mantle wedge from
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33 west to east.
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41 No ultrapotassic mafic magmatism (e.g. MgO>6wt%) occurred to the east of 87°E
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43 during the period 25-8Ma in the Lhasa terrane (Figs. 1 and 21). This may indicate the absence
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45 of mantle pyroxenite to the east of 87°E, reflecting the reduction in components derived from
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47 subducted Indian continental crust. The Indian slab may also be subducting more steeply
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49 beneath the Lhasa terrane to the east of 87°E compared to further west (Zhou & Murphy,
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51 2005; Negredo *et al.*, 2007; Chen *et al.*, 2015). Compared to west of 87°E, the lack of
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53 significant recycling of subducted Indian continental crustal materials during the northward
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55 underthrusting of the Indian slab may have led to a relatively hot and dry mantle wedge to the
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2 east of 87°E. This is consistent with the estimated temperatures of up to 1300 °C and
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4 pressures of 24 kbar for granulite facies crustal xenoliths entrained in a Miocene K-rich dyke
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6
7 intruded into the eastern Lhasa terrane (Chan *et al.*, 2009).
8
9

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Figure Captions

Fig. 1. Simplified map showing the distribution of Cenozoic magmatic rocks in southern Tibet (modified from Pan *et al.*, 2004; Ding *et al.*, 2006; Guo *et al.*, 2007, 2013; Hou *et al.*, 2013). The red dashed line labeled A-A1 represents the location of the line of the Hi-CLIMB seismic array at ~85 °E from Nabelek *et al.* (2009).

BNS, Bangong-Nujiang Suture; ITS, Indus-Tsangpo Suture. Numbers indicate the individual volcanic fields (or intrusive bodies): 1: Shiquanhe; 2: Chajiasi; 3: Bongba; 4: Xungba; 5: Gegar; 6: Manasarowar; 7: Yare; 8: Puridazong; 9: Mayum; 10: Jarga; 11: Sailipu; 12: Zabuye; 13: Maiga; 14: Daggyai; 15: Saga; 16: Garwa; 17: Yaqian; 18: Konglongxiang; 19: Mibale; 20: Yiqian ; 21: Chazi ; 22: Pabbai; 23: Sangsang; 24: Zhuno; 25: Kuday; 26: Nanmuqie; 27: Xigaze; 28: Namlin; 29: Chongjiang; 30: Majiang; 31: Nimu; 32: Tinggong; 33: Pagu; 34: Yangying; 35: Nanmu; 36: Qulong; 37: Lakang'e; 38: Jiama; 39: Tangbula; 40: Linzhi.

Fig. 2. Ages of the post-collisional ultrapotassic magmatic rocks from north to south (a) and from west to east (b) in south Tibet. WSVF, western sub-group volcanic fields; CSVF, central sub-group volcanic fields; ESVF, eastern sub-group volcanic fields. Data for the ages of the magmatic rocks are from Table 1.

Fig. 3. Representative photomicrographs (a-e) of the ultrapotassic mafic volcanic rocks studied: (a) Shows a complex reaction rim around a mantle olivine xenocryst from sample (XB1307) from the Xungba volcanic field; this includes both orthopyroxene (Ca-poor

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2
3 pyroxene) and clinopyroxene (Ca-rich diopside) (cross-polarized light). This may be a
4
5 fragment of the source olivine-bearing pyroxenite. Black circles represent points of electron
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7 microprobe analyses. (b) Corroded olivine phenocryst in sample (MG1303) from the Maiga
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9 volcanic field (cross-polarized light). (c) Reaction rim of clinopyroxene on an olivine
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11 xenocryst in sample (XB1307) from the Xungba volcanic field (cross-polarized light). (d)
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13 Phenocrysts of clinopyroxene and olivine in sample (MG1301) from the Maiga volcanic field
14
15 (cross-polarized light). (e) Phenocrysts of clinopyroxene and phlogopite in sample (XB1307)
16
17 from the Xungba volcanic field (cross-polarized light). (f) Back-scattered electron (BSE)
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19 image of the olivine grain in (a) clearly showing the xenocrystic olivine core (darker grey).
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25 Abbreviations: Cpx, clinopyroxene; Ol, olivine; Opx, orthopyroxene; Phl, phlogopite.
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30 **Fig. 4. (a)** $K_2O + Na_2O$ (wt %) vs SiO_2 (wt %) for the post-collisional ultrapotassic mafic
31
32 igneous rocks of the Lhasa terrane. All data plotted have been recalculated to 100 wt % on a
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34 volatile-free basis (Table 3). Classification boundaries are from Le Bas *et al.* (1986) and Le
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36 Maitre *et al.* (1989). Filled and open symbols represent, respectively, data from this study and
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38 the published data of Williams *et al.* (2001, 2004), Miller *et al.* (1999), Zhao *et al.* (2009),
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40 Tian *et al.* (2012), Wang *et al.* (2008), Sun *et al.* (2007), Chen *et al.* (2012), Jiang *et al.*
41
42 (2003), Ding *et al.* (2003, 2006), Gao *et al.* (2007) and Guo *et al.* (2013). Classification fields
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44 shown by letters are as follows. S2: basaltic trachyandesite; S3: trachyandesite; U1: tephrite;
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46 U2: phonotephrite; U3: tephriphonolite. **(b)** K_2O (wt %) vs SiO_2 (wt %) diagram for the same
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48 samples plotted in (a). The dividing lines show the classification boundaries from Peccerillo
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50 & Taylor (1976) and Rickwood (1989). Data sources and symbols are as in (a).
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3 **Fig. 5. (a)** CaO (wt %) vs Al₂O₃ (wt %); **(b)** CaO (wt %) vs MgO (wt %) for the Lhasa
4 terrane ultrapotassic magmatic rocks. Classification boundaries are from Foley *et al.* (1987).
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6 Data have been normalized to 100 wt % volatile-free as indicated in Table 3. Data sources are
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8 as in Fig. 4.
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15 **Fig. 6.** Selected major elements vs MgO (wt %) indicating the compositional range of the
16 post-collisional ultrapotassic igneous rocks in south Tibet. (a) SiO₂; (b) TiO₂; (c) Al₂O₃; (d)
17 TFe₂O₃; (e) CaO; (f) Na₂O; (g) K₂O; (h) P₂O₅. Data have been normalized to 100 wt %
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19 volatile-free as indicated in Table 3. Data sources and symbols are as in Fig. 4.
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21 TFe₂O₃: all Fe reported as Fe₂O₃.
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30 **Fig. 7.** Chondrite-normalized rare earth element (REE) patterns for the western (a), central (b)
31 and eastern (c) sub-group volcanic fields of the Lhasa terrane in south Tibet; normalization
32 factors are from Sun & McDonough (1989). The mélange average composition is from
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34 Marschall & Schumacher (2012). Data sources are as in Fig. 4.
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43 **Fig. 8.** Primitive mantle-normalized trace element patterns for the western (a), central (b) and
44 eastern (c) sub-group volcanic fields of the Lhasa terrane in south Tibet; normalization factors
45 are from Sun & McDonough (1989). The mélange average composition is from Marschall &
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47 Schumacher (2012). Data sources are as in Fig. 4.
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55 **Fig. 9.** La/Yb vs La (ppm) for the Lhasa terrane ultrapotassic magmatic rocks. Data sources
56 and symbols are as in Fig. 4.
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5 **Fig. 10.** Rb/Sr vs Ba/Rb for the Lhasa terrane ultrapotassic magmatic rocks. Amphibole and
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7 phlogopite on the arrows refer to these phases as residual phases in the mantle source region
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9 of the magmas. Data sources and symbols are as in Fig. 4.
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15 **Fig. 11.** Variation in Sr-Nd-Pb isotope and trace element ratios of the post-collisional
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17 ultrapotassic mafic magmatic rocks from west to east in the Lhasa terrane. Data are from
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19 Tables 3-5. Symbols are as in Fig. 4.
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25 **Fig. 12.** (a) $(^{143}\text{Nd}/^{144}\text{Nd})_i$ vs $(^{87}\text{Sr}/^{86}\text{Sr})_i$ in the post-collisional ultrapotassic mafic magmatic
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27 rocks; (b) enlargement of the field of data in (a); (c) $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs $(^{206}\text{Pb}/^{204}\text{Pb})_i$; (d)
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29 $(^{143}\text{Nd}/^{144}\text{Nd})_i$ vs $(^{206}\text{Pb}/^{204}\text{Pb})_i$; (e) $(^{207}\text{Pb}/^{204}\text{Pb})_i$ vs $(^{206}\text{Pb}/^{204}\text{Pb})_i$; (f) $(^{208}\text{Pb}/^{204}\text{Pb})_i$ vs
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31 $(^{206}\text{Pb}/^{204}\text{Pb})_i$. Arrows (or outlined fields) indicate the composition of DMM (Workman &
32
33 Hart, 2005) and Indian continental crust [see Inger & Harris (1993), Ahmad *et al.* (2000),
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35 Richards *et al.* (2005) and Guo & Wilson (2012) for more detailed discussion]. The NHRL
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37 (Northern Hemisphere Reference Line; Hart, 1984), fields for OIB and BSE (Wilson, 1989;
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39 Hofmann, 1997) are shown for reference. Fields labeled Italy and Spain denote the
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41 compositions of Italian and Spanish lamproites, respectively, based on data from Gasperini *et*
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43 *al.* (2002), Perini *et al.* (2004), Owen (2008), Prelevic *et al.* (2008), Boari *et al.* (2009) and
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45 Conticelli *et al.* (2009). BSE: Bulk Silicate Earth. Data are from Tables 4 and 5. Symbols are
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47 as in Fig. 4.
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58 **Fig. 13.** Fo content (mole %) in olivine phenocrysts vs Mg# in the whole-rocks. Solid and
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1
2 dashed curve lines show the Fe-Mg exchange equilibrium for $K_{D(\text{Fe-Mg})} = 0.3 \pm 0.03$ between
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4 olivine and melt (Roeder & Emslie, 1970). The data for olivine phenocrysts in the Mibale
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6 volcanic rocks are from Yu *et al.* (2012). Because olivine phenocrysts from different samples
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8 in the same volcanic field were analysed, we use different symbols to distinguish these. Data
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10 are from Table 3 and Supplementary Data Appendix A.
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16 **Fig. 14.** Disequilibrium in a mantle olivine xenocryst. (a) Photomicrograph showing a
17
18 disequilibrium texture between the olivine xenocryst and the surrounding melt
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20 (cross-polarized light; sample XB1307). Black circles indicate the locations of electron
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22 microprobe analyses. (b) Back-scattered electron image of the olivine xenocryst showing the
23
24 location of the electron microprobe profile A-B. The xenocrystic core of the olivine is a
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26 darker grey. The lighter rim formed by reaction with the host melt. (c) Variation of Fo and
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28 CaO content along the profile A-B. The low CaO (less than 0.1 wt %; shown by the solid blue
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30 line) content of the core of the olivine crystal, combined with its high Fo (> 85 %; shown by
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32 the solid red line), is consistent with the core being xenocrystic (Prelevic *et al.*, 2013). The
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34 rim of the olivine crystal is compositionally similar to the olivine phenocrysts in the sample.
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44 **Fig. 15.** Variation of CaO, MnO and NiO versus MgO (wt%) in olivine phenocrysts and
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46 xenocrysts. (a) CaO (wt %) vs MgO (wt %); (b) MnO (wt %) vs MgO (wt %); (c) NiO (wt %)
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48 vs MgO (wt %). The fields outlined by green dashed lines indicate the compositions of
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50 olivine phenocrysts in MORB (Sobolev *et al.*, 2007). The fields outlined with the black solid
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52 lines denote the compositions of olivine phenocryst in partial melts of peridotite (Sobolev *et*
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54 *al.*, 2007). The fields outlined by pink dashed lines indicate the compositions of olivine in
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56 mantle xenoliths from south Tibet (Liu *et al.*, 2011). The arrow in (b) denotes olivine
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2 fractional crystallization (Sobolev *et al.*, 2007). Data are from Supplementary Data Appendix
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10 **Fig. 16.** Chondrite-normalised transition element variations in the western (a), central (b) and
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12 eastern (c) sub-group volcanic fields in south Tibet; normalization factors are from Langmuir
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14 *et al.* (1977). The grey fields indicate the composition of MORB (Rogers *et al.*, 1985;
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16 Hofmann, 1997; Workman & Hart, 2005; GERM website: <http://www.earthref.org/>). Data
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18 sources are as in Fig. 4.
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25 **Fig. 17.** Correlations between whole-rock Sr-Nd-Pb isotope initial ratios and selected major
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27 and trace element contents corrected to $Mg\#=72$ after Niu *et al.* (1999) and Humphreys &
28
29 Niu (2009). Compositions of N-MORB (Presnall & Hoover, 1987; Rollinson, 1993; Hofmann,
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31 1997; Workman & Hart, 2005) are shown for comparison. Data are from Tables 3, 4 and 5.
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38 **Fig. 18. (a)** CaO (wt %) vs MgO (wt %) variations in the whole-rock compositions of the
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40 ultrapotassic magmatic rocks from the Lhasa terrane indicating their origin as partial melts of
41
42 pyroxenite. The red dashed line in (a) separates peridotite-derived melts (field above the line)
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44 from pyroxenite-derived melts (field below the line) (after Herzberg, 2011). **(b)** Ni (ppm) vs
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46 MgO (wt %). **(c)** Ni (ppm)/ MgO (wt %) vs SiO₂ (wt %). The red dashed curves in (b) and (c)
47
48 separate peridotite-derived melts (field below the curves) from pyroxenite-derived melts
49
50 (field above the curves) (after Sobolev *et al.*, 2005). Compositional fields of
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52 pyroxenite-derived (orange field) and peridotite-derived (green field) partial melts in (b) and
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54 (c) are based on Hawaiian data (Sobolev *et al.*, 2005). Most of the ultrapotassic magmatic
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2 rocks plotted exhibit higher Ni concentrations than peridotite-derived melts in (b) and (c),
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4 suggesting a pyroxenite mantle source (see text for details). The ultrapotassic magmatic rocks
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6 from the Lhasa terrane have higher SiO₂ and lower MgO contents compared to those of
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8 partial melts of Hawaiian pyroxenite (Sobolev *et al.*, 2005), consistent with a contribution
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10 from a subducted continental crustal component to their mantle source. Data and symbols are
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12 as in Fig. 4 (a).
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20 **Fig. 19.** Comparison of the olivine phenocryst compositions in the ultrapotassic magmatic
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22 rocks in south Tibet with those crystallized from partial melts of mantle peridotite and
23
24 pyroxenite (after Herzberg, 2011). (a) Ca (ppm) vs Fo (mole %); (b) Mn (ppm) vs Fo
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26 (mole %); (c) Fe (ppm)/Mn (ppm) vs Fo (mole %). The olivine phenocrysts from the Lhasa
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28 terrane magmatic rocks are characterized by lower Ca and Mn contents compared to those
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30 crystallized from partial melts of mantle peridotite. This indicates a mantle pyroxenite source
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32 for the magmas, following the criteria of Herzberg (2011). We have further subdivided the
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34 pyroxenite source into olivine-bearing pyroxenite (red symbols) and olivine-free pyroxenite
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36 (other colour symbols) mantle sources according to the differences in the Ca, Mn and Fe/Mn
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38 of the olivine phenocrysts. Blue dashed lines, which separate the fields of olivine phenocrysts
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40 crystallized from partial melts of peridotite and pyroxenite, are from Herzberg (2011). The
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42 arrow labeled by “Ol” denotes olivine fractional crystallization (Sobolev *et al.*, 2007). Data
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44 are from Supplementary Data Appendix A.
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55 **Fig. 20.** Variation of Ni (wt %) vs Fo (mole %) in olivine phenocrysts in the post-collisional
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57 ultrapotassic magmas in south Tibet. The red modeling curves A-X, B-Y and C-Z simulate
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3 the variation trends (B1, B2 and B3) in olivine composition as a result of olivine fractionation
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5 from parental magmas A, B and C, respectively. The numbers in percent shown on the tick
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7 marks of the three red modeling curves (i.e. A-X, B-Y and C-Z) denote proportions of the
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9 olivine fractionation (See Supplementary Data Appendix B for the modeling calculation
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11 details). The blue arrow indicates that the parental magmas (A, B, C) were derived from
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13 pyroxenite sources which contain an increasing component of partial melt from subducted
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15 Indian eclogite facies crust from C to A. The field labeled “R” represents the composition of
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17 olivine crystallized from partial melts of olivine-free pyroxenite. The field labeled “P”
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19 represents the composition of olivine crystallized from partial melt of mantle peridotite. Data
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21 for the compositions of olivine phenocrysts in the Tibetan ultrapotassic magmatic rocks are
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23 from Supplementary Data Appendix A. The sample numbers shown in the legend are
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25 consistent with those in Table 3.
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35 **Fig. 21.** Cartoon showing the mantle source components of the post-collisional mafic
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37 ultrapotassic magmas in south Tibet. A: silica-rich pyroxenite; B: olivine-free pyroxenite; C:
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39 olivine-bearing pyroxenite. Abbreviations: WSVF: western sub-group volcanic fields; CSVF:
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41 central sub-group volcanic fields; ESVF: eastern sub-group volcanic fields. Sailipu: Sailipu
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43 volcanic field; Xungba: Xungba volcanic field; Zabuye: Zabuye volcanic field; Maiga: Maiga
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45 volcanic field; Mibale: Mibale volcanic field.
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52 **Fig. 22.** (a) Th/La vs Sm/La. (b) $(^{143}\text{Nd}/^{144}\text{Nd})_i$ vs Th/La. (c) $(^{143}\text{Nd}/^{144}\text{Nd})_i$ vs Ba/La. (d)
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54 Dy/Dy* vs Dy/Yb. (e) $(^{143}\text{Nd}/^{144}\text{Nd})_i$ vs Dy/Dy*. (f) Ce/Pb vs Nb/La.
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58 The mélange average in (f) is from Marschall & Schumacher (2012). (d) is modified from
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Davidson *et al.* (2013); the definition of Dy/Dy^* in (d) and (e) is from Davidson *et al.* (2013). Data for OIB and MORB are from Wilson (1989) and Sun & McDonough (1989). The field of arc basalts is from Plank (2005). Data for the Linzizong volcanic rocks are from Dong (2002), Mo *et al.* (2008) and Lee (2009, 2012). The compositions of upper continental crust, lower continental crust and bulk continental crust are from Rudnick & Gao (2003). The composition of Indian continental crust is based on data from Inger & Harris (1993), Ahmad *et al.* (2000), Richards *et al.* (2005), Rehman *et al.* (2008) and Guo & Wilson (2012). Data for the Lhasa terrane ultrapotassic rocks are from Tables 3 and 4. Symbols are as in Fig. 4.

Fig. 23. Dy/Yb vs La/Yb variations in the ultrapotassic magmatic rocks indicating a decreasing proportion of the eclogite partial melt component in the mantle source of the magmas from west to east in south Tibet. The numbers in percent shown on the filled rhombs of the solid red modeling curve denote the degree of partial melting of the crustal eclogite. The bulk composition of the pre-existing mantle peridotite protolith (i.e. DMM) is from Workman & Hart (2005). The numbers in percent shown on the filled rhombs of the black dashed modeling curves (M1, M2 and M3) represent the proportions of the eclogite-derived melt in the two-component mixture between the eclogite-derived melt and the mantle peridotite protolith (i.e. DMM). Data and symbols are as in Fig. 4.

Fig. 24. Petrogenetic model for the post-collisional ultrapotassic mafic magmatism from 25-8Ma in south Tibet.

(a) **55-25 Ma:** Northward subduction of Indian continental lithosphere beneath the Lhasa terrane during this early stage resulted in compressional deformation in south Tibet, the occurrence of the eclogite facies crustal materials on the top of the subducted Indian slab and

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2 formation of pyroxenites within the mantle wedge and in the Tibetan SCLM as a consequence
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4 of reaction of partial melts and fluids derived from subducted Indian eclogite facies
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7 continental crustal materials with ambient mantle peridotite.
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10 **(b) 25-8 Ma:** Generation of post-collisional leucogranites in the Himalayas and
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12 ultrapotassic mafic magmas in the Lhasa terrane of south Tibet. Slab roll-back and break-off
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14 induced upper plate extension and asthenospheric upwelling as a counter flow. This is the
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16 period in which the pyroxenites which formed in the metasomatised SCLM and
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18 asthenospheric mantle wedge partially melted to produce the ultrapotassic magmas. The
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20 petrogenesis of the Himalayan leucogranites is discussed in Guo & Wilson (2012).
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25 **(c) 8 Ma-the present:** Flat-slab subduction of Indian continental lithosphere during this
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27 late stage has eliminated the wedge of upwelling asthenospheric mantle and Tibetan SCLM,
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29 terminating the mafic ultrapotassic magmatism. The cross section of the India-Asia collision
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31 zone during the period 8Ma to the present is based on geophysical constraints (e.g. Fig. 2D of
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33 Nabelek *et al.*, 2009). The eclogite facies Indian lower continental crust (green shading)
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35 underplates the Himalayas and Tibet up to 31°N.
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39 Abbreviations are as follows. BNS: Bangong-Nujiang suture; ITS: Indus-Tsangpo suture;
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41 MBT: Main Boundary thrust; MCT: Main Central thrust; MHT: Main Himalayan Thrust;
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43 STDS: south Tibetan detachment system. Filled red triangles represent the locations of
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45 post-collisional potassium-rich mafic magmatism in south Tibet. The white arrow denotes the
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47 detachment of the Indian continental slab.
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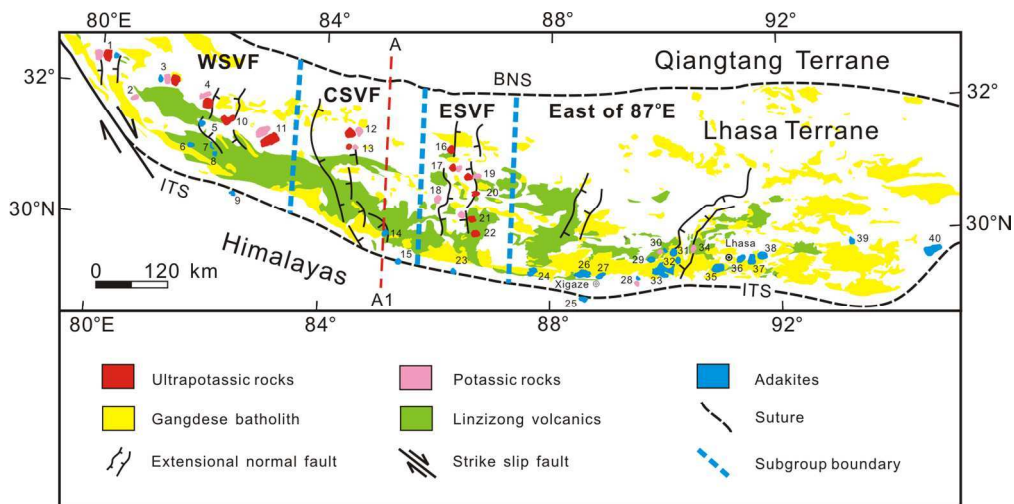


Figure 1

Fig.1

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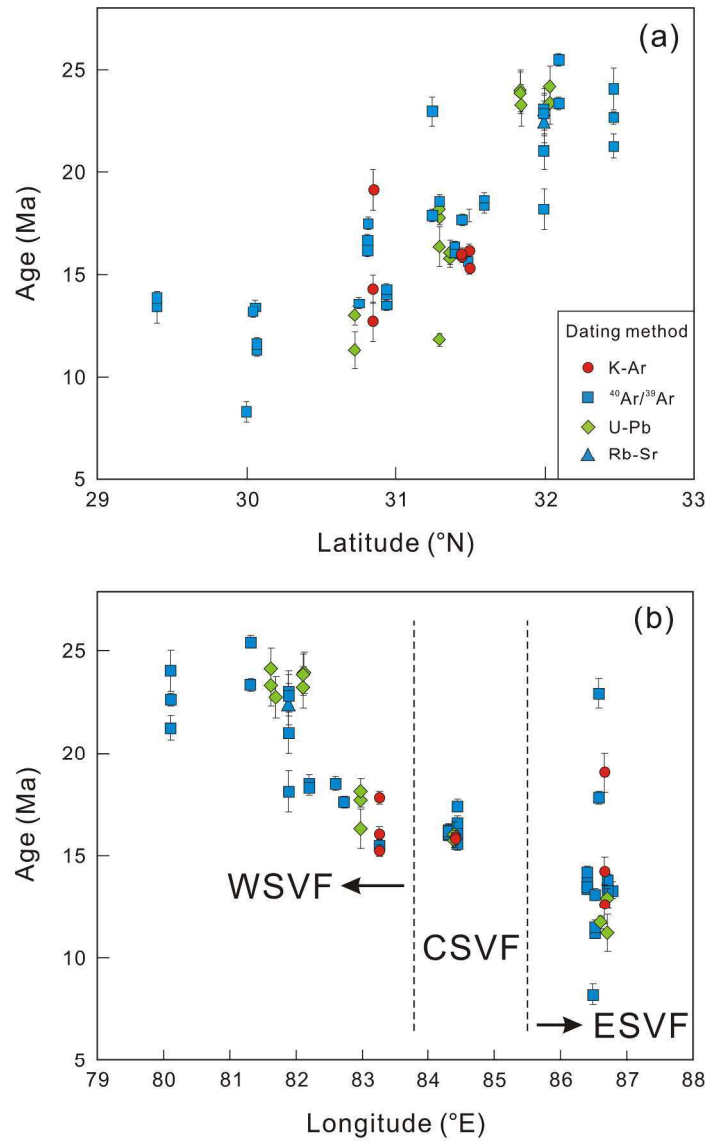


Figure 2

Fig.2

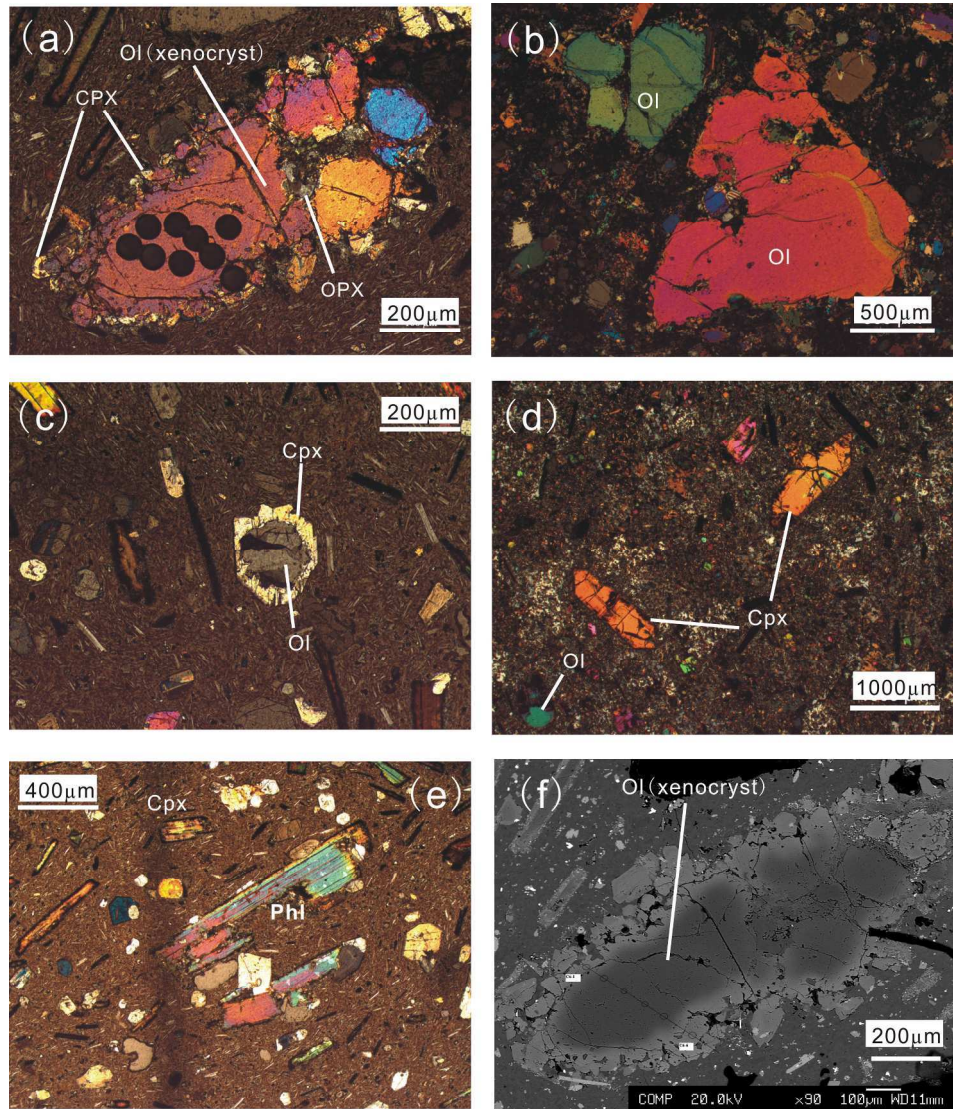


Figure 3

Fig.3

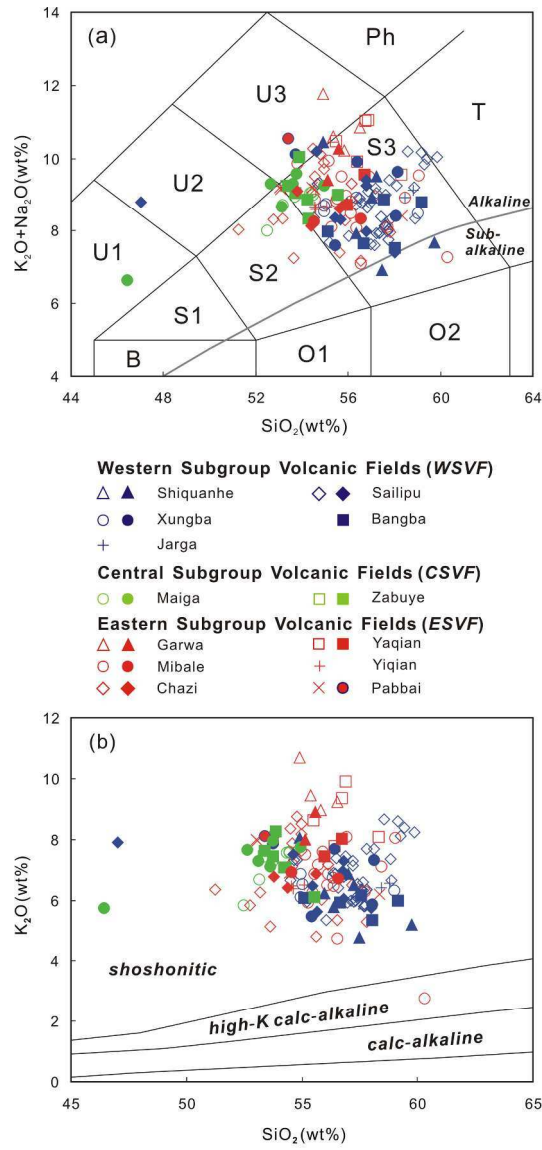


Fig. 4

Fig.4

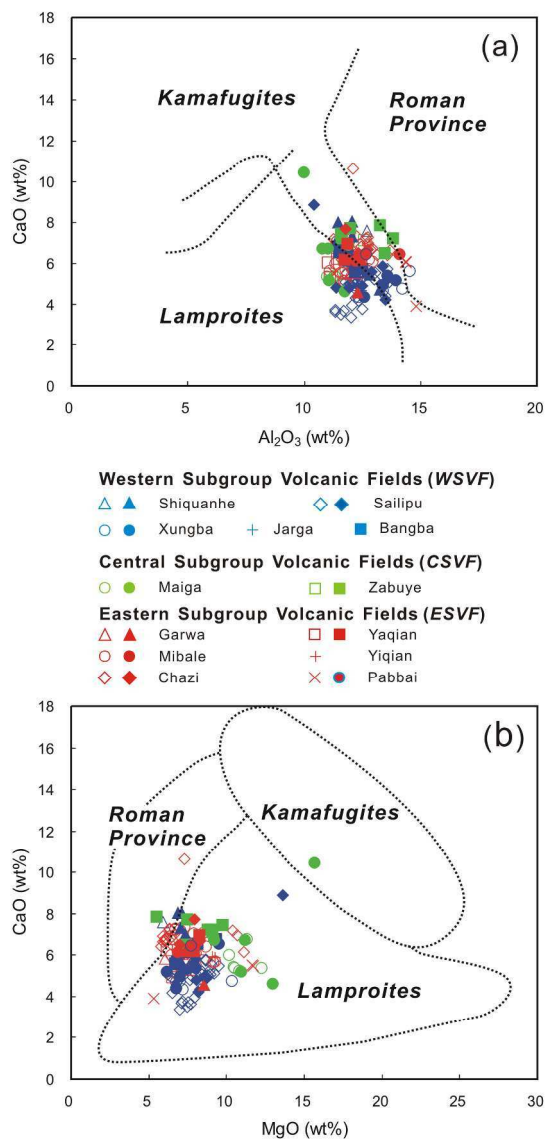


Figure 5

Fig.5

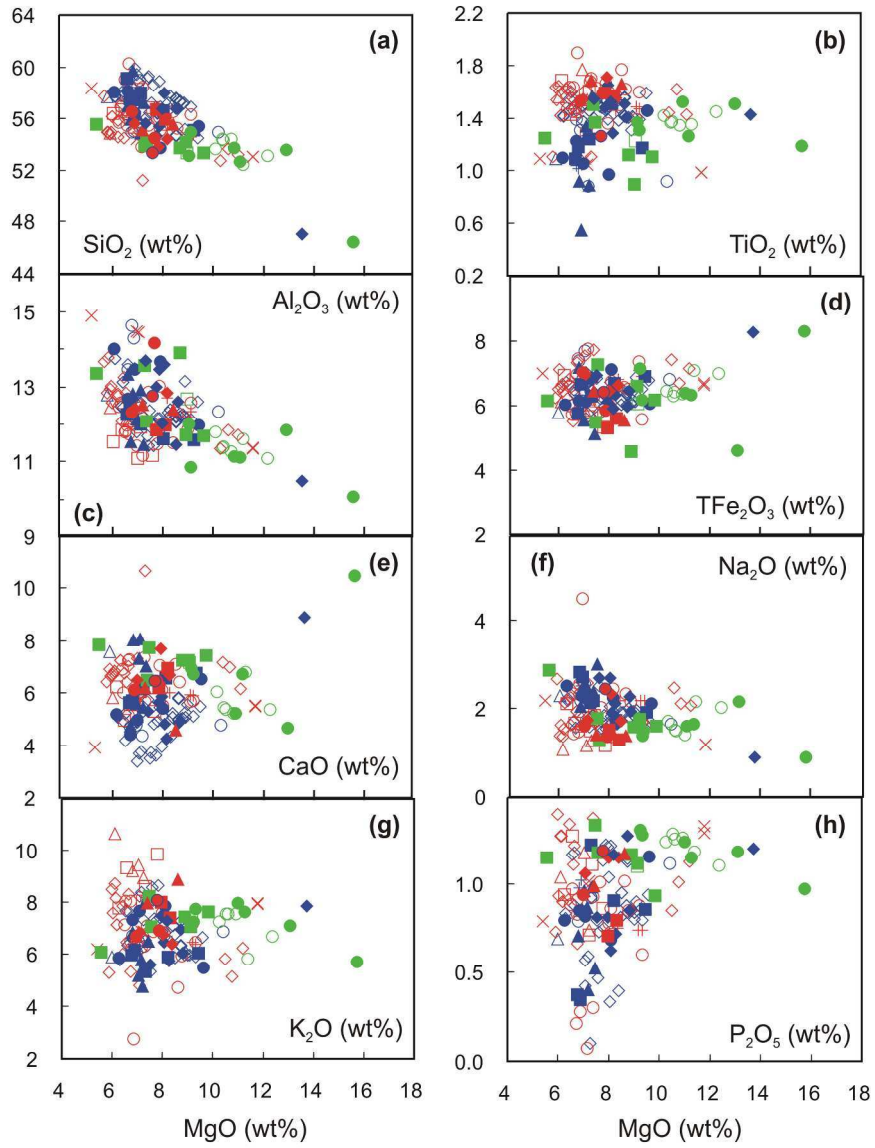


Figure 6

Fig.6

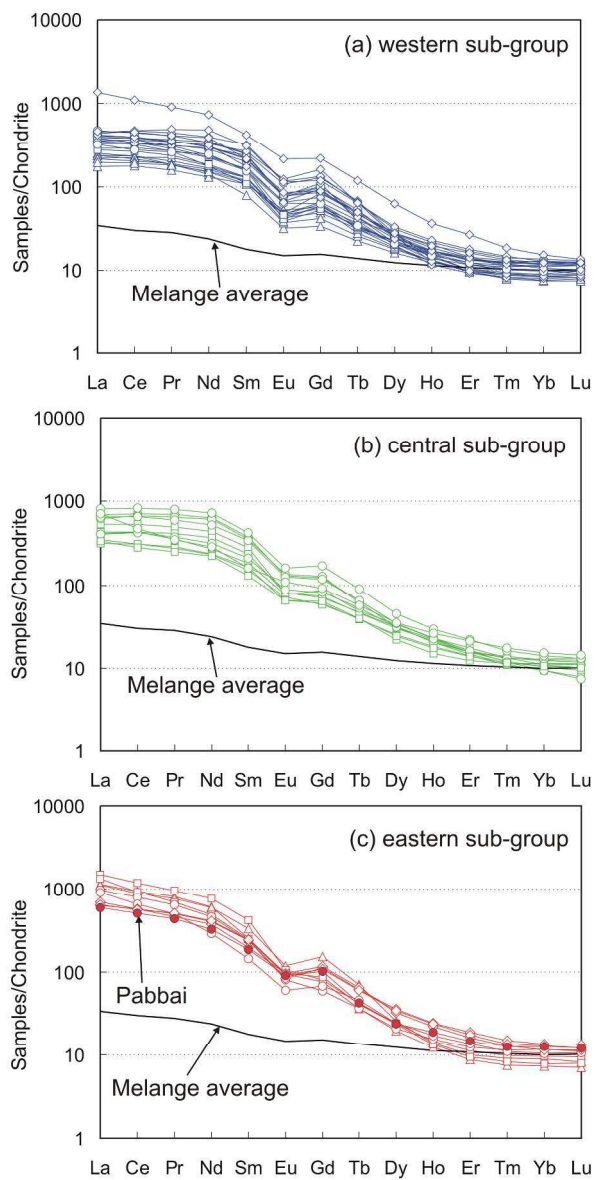


Fig. 7

Fig.7

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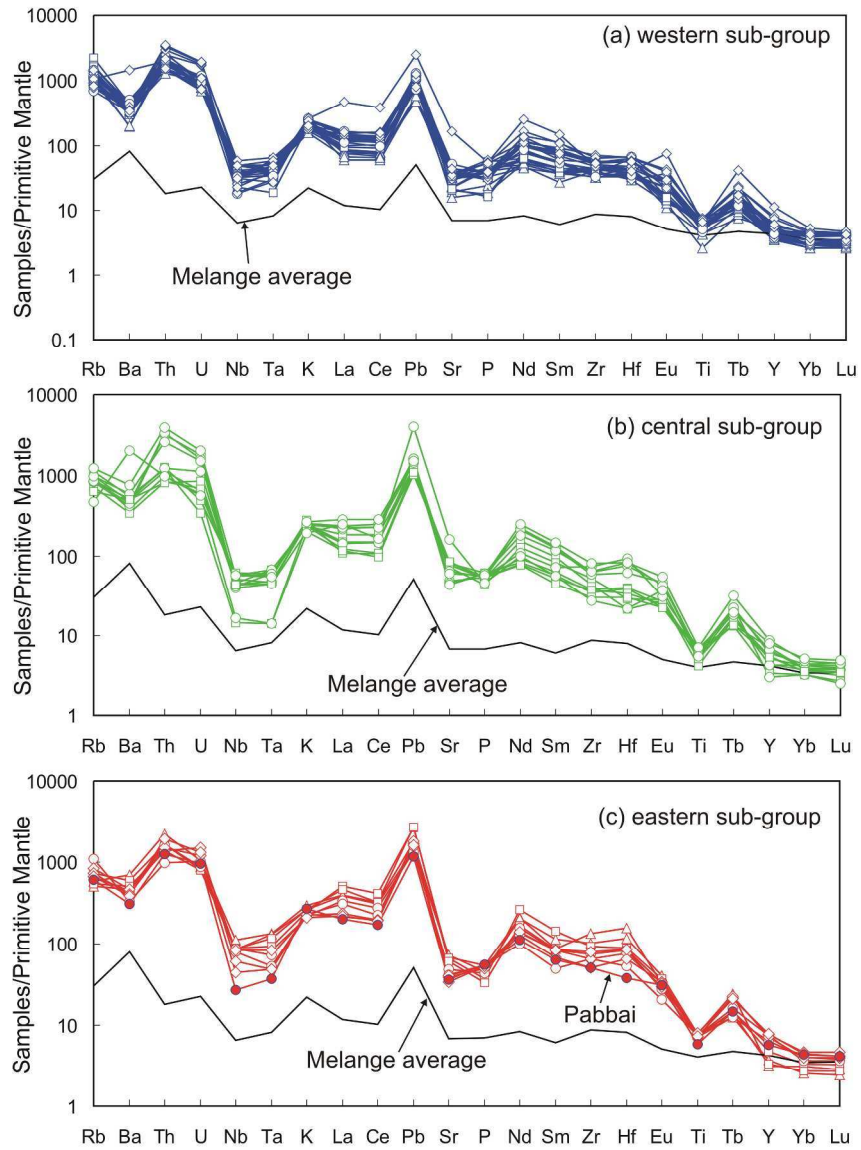


Fig. 8

Fig.8

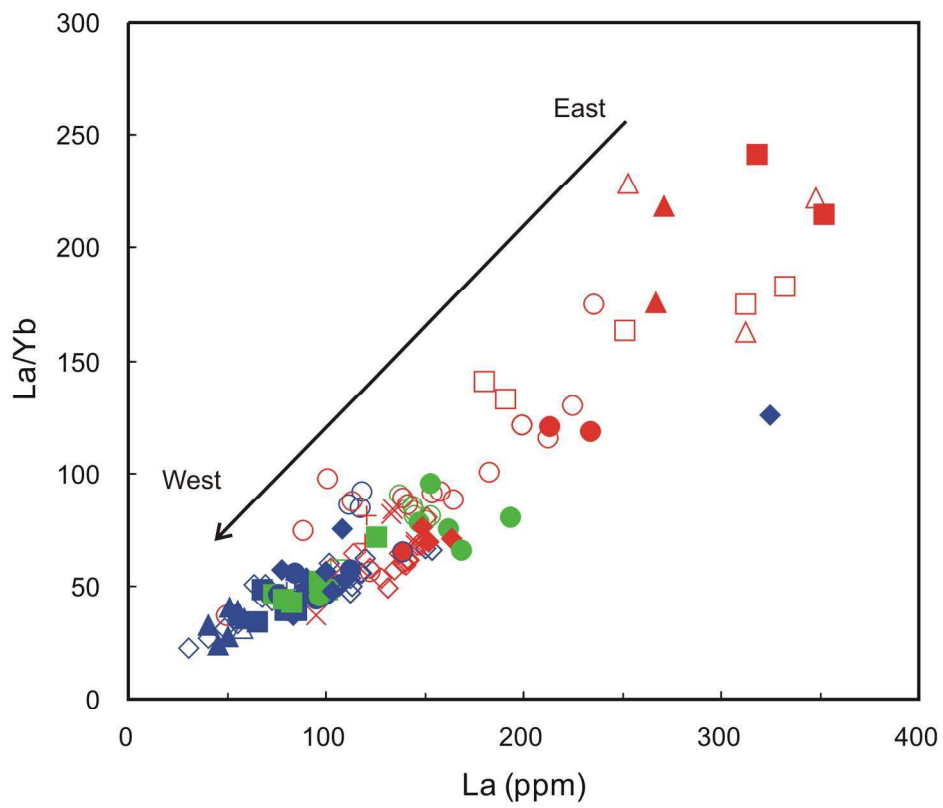


Figure 9

Fig.9



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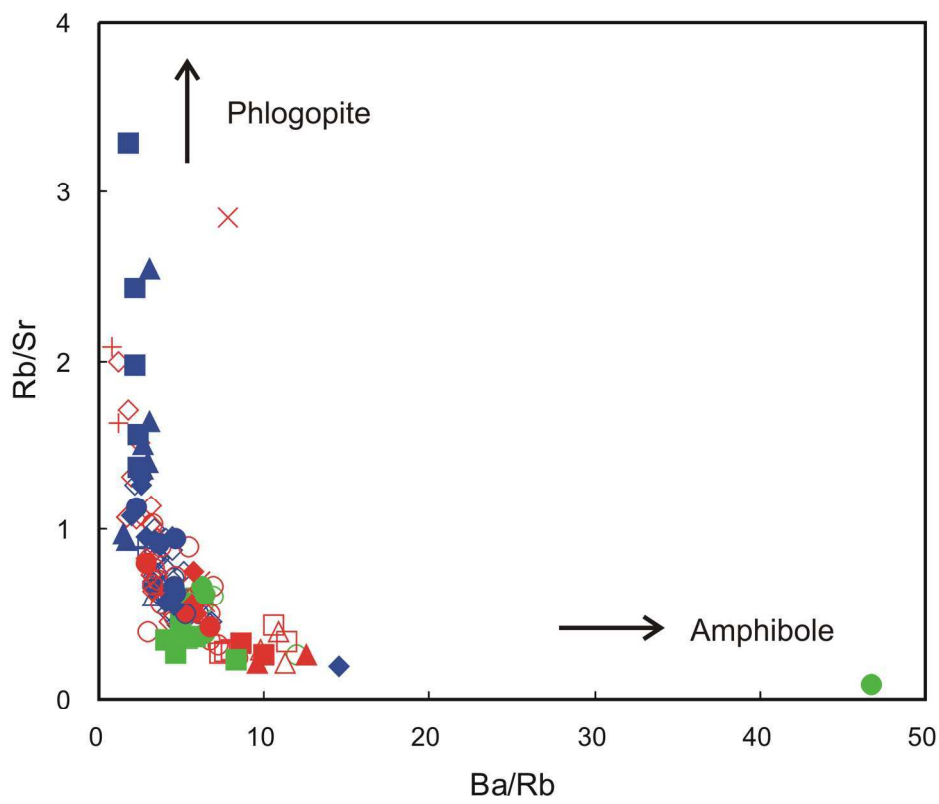


Figure 10

Fig.10



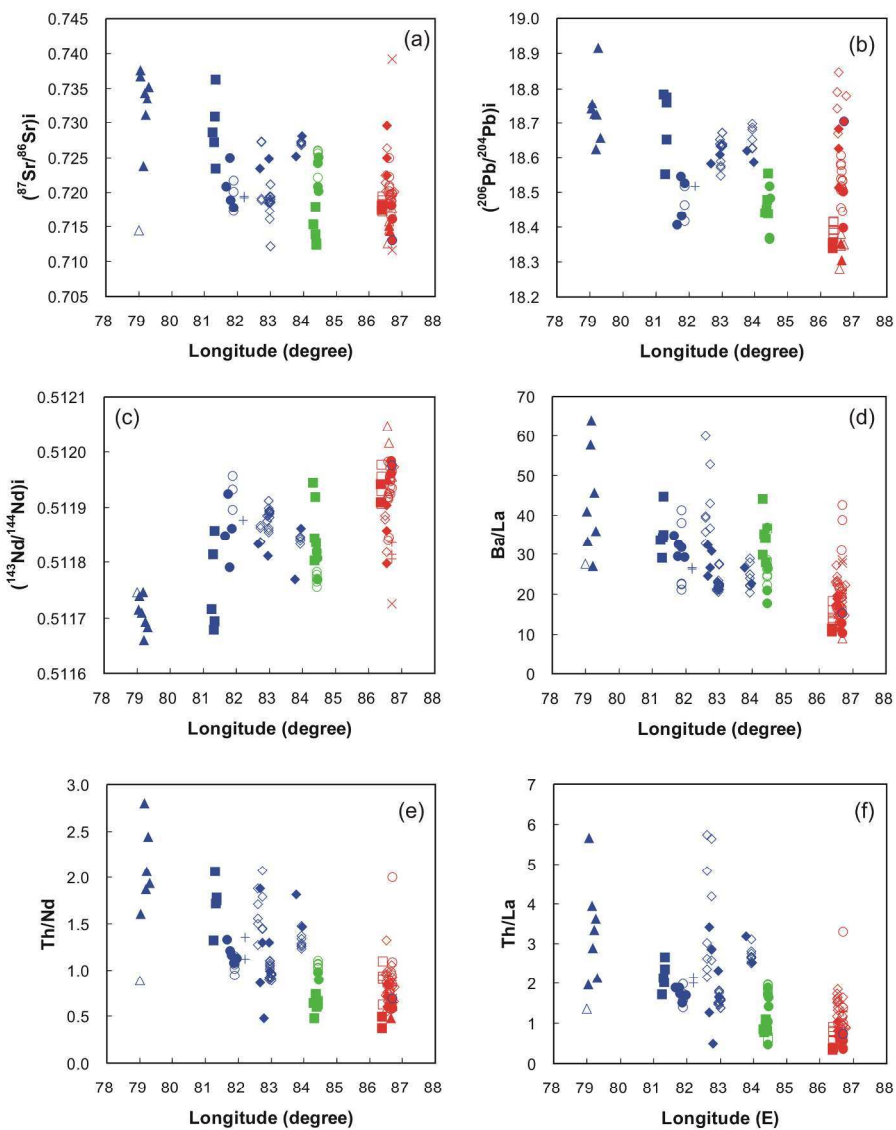


Figure 11

Fig.11

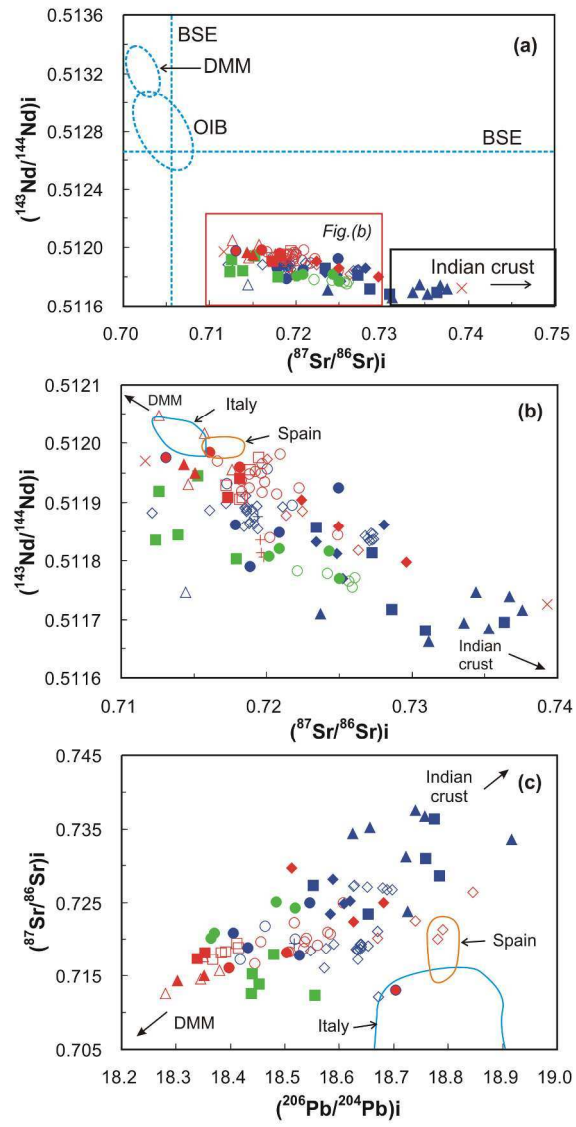


Figure 12(a)

Fig.12a

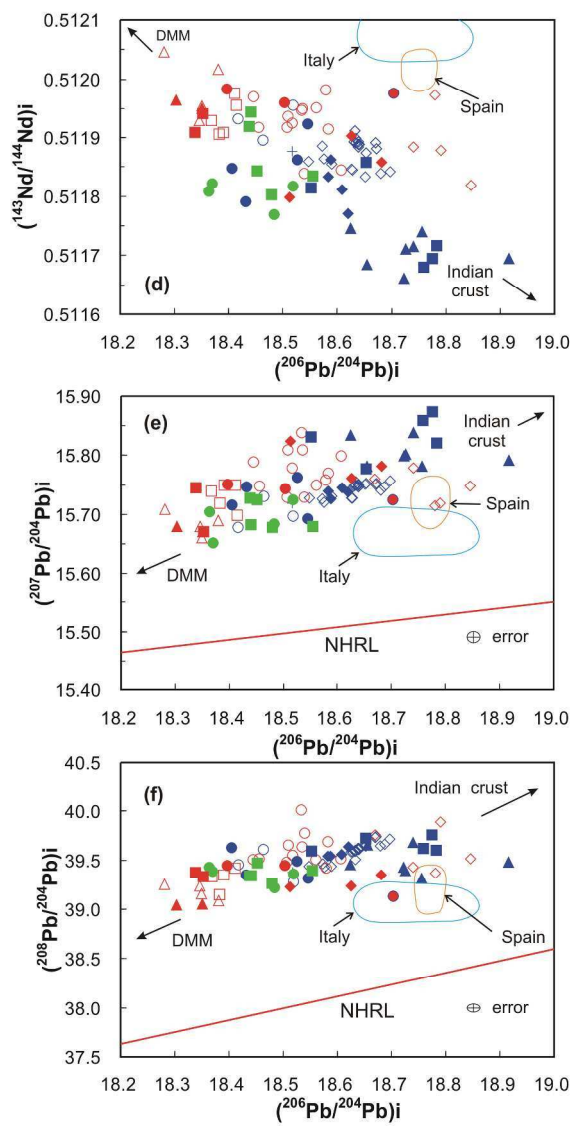
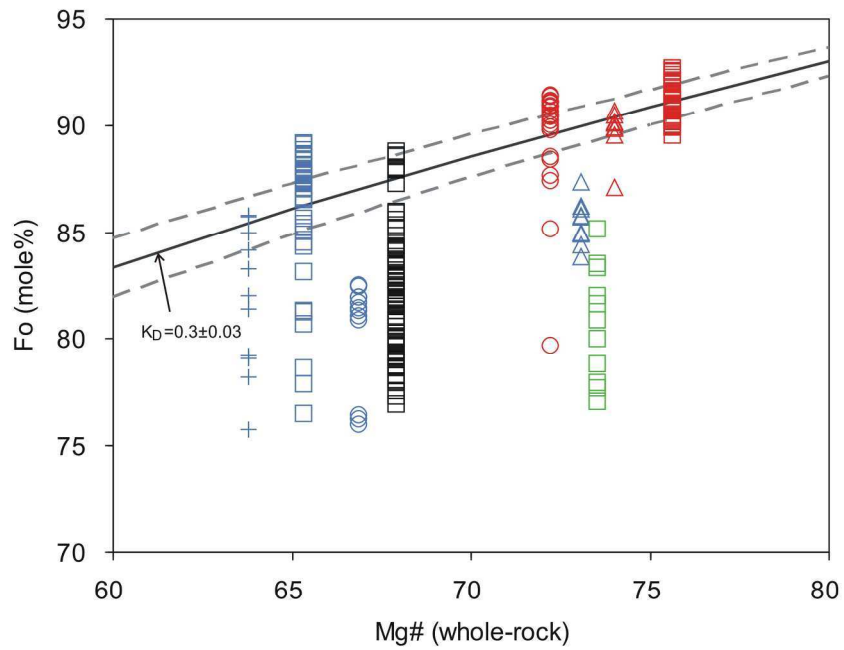


Fig.12b



Western Subgroup Volcanic Fields (WSVF)

- Xungba(XB1307)
- △ Sailipu (SLP1320)
- Sailipu (SLP1314)
- Sailipu (SLP1302)
- Sailipu (SLP1308)

Central Subgroup Volcanic Fields (CSVF)

- + Zabuye (BGS1306)
- Maiga (Mg1301)
- Maiga (Mg1303)

Eastern Subgroup Volcanic Fields (ESVF)

- △ Mibale

Figure 13

Fig.13

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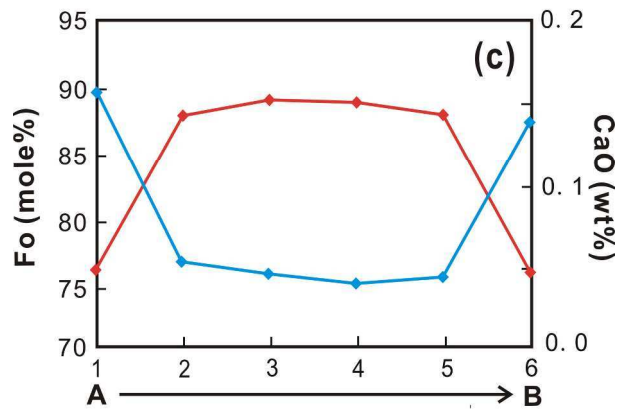
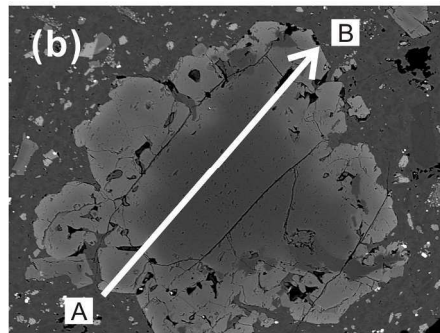
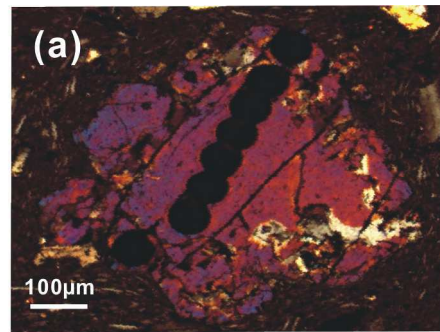


Figure 14

Fig.14

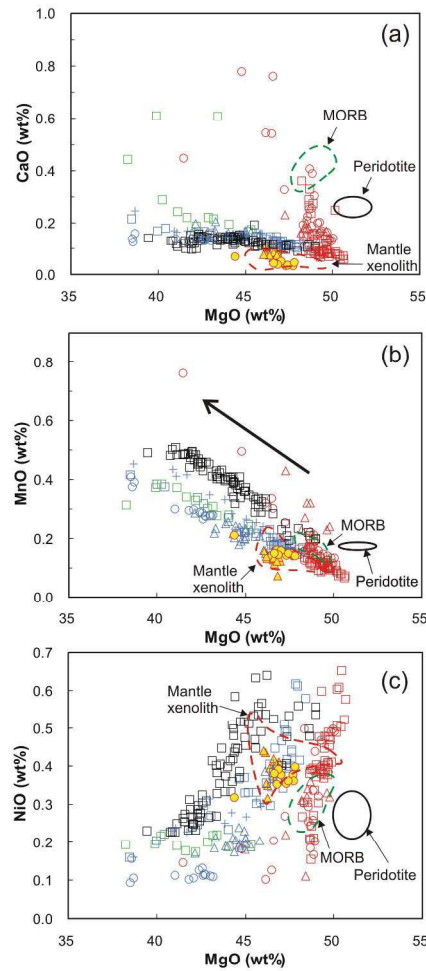


Figure 15

Fig.15

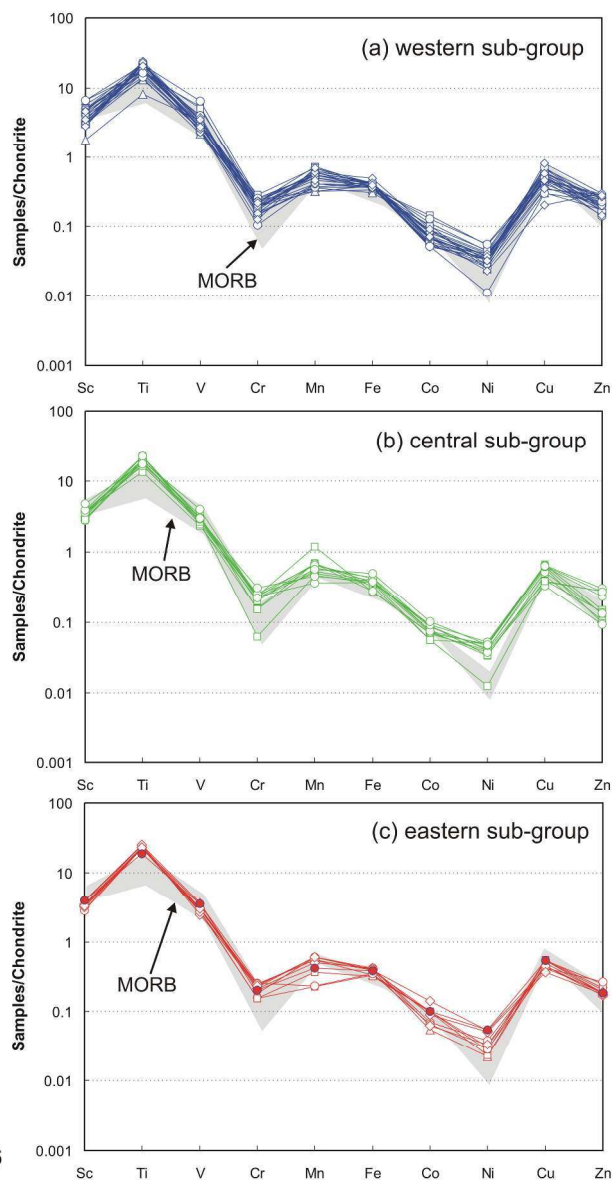


Fig. 16

Fig.16

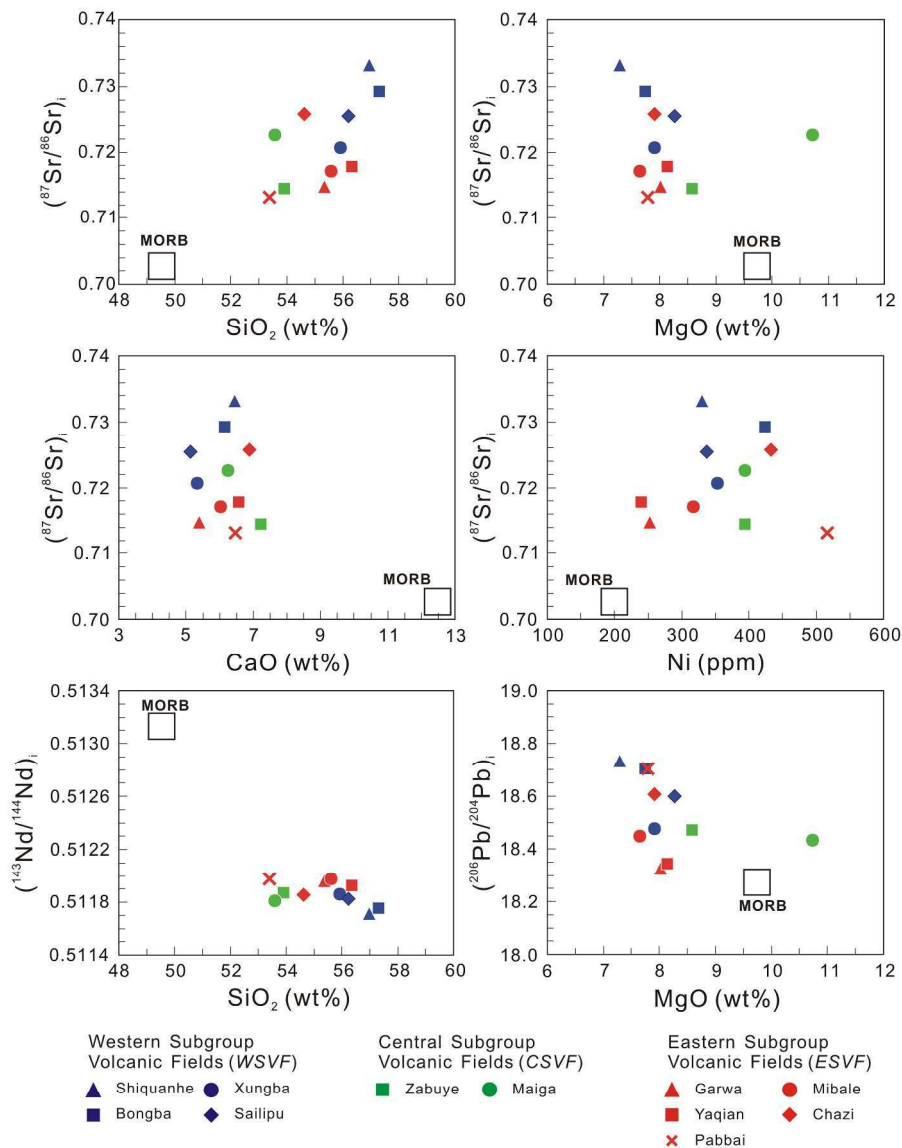


Figure 17

Fig.17

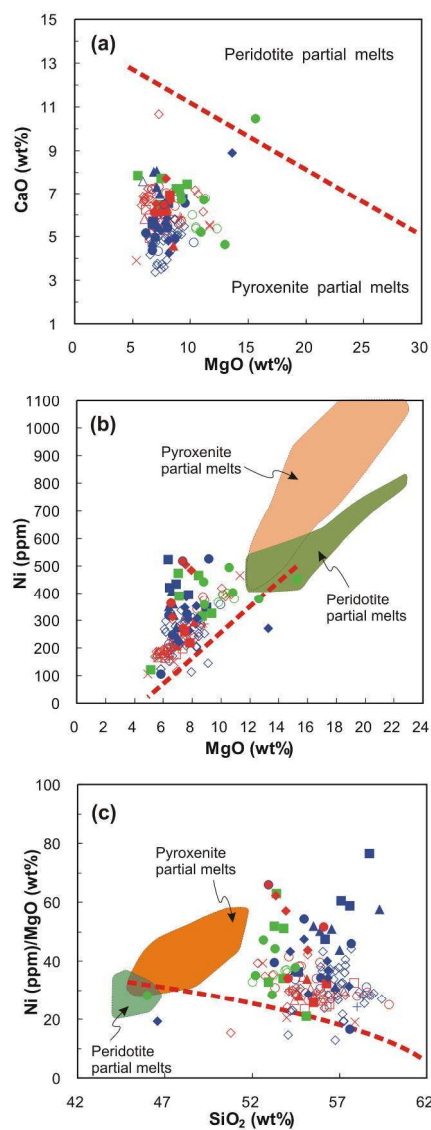


Figure 18

Fig.18

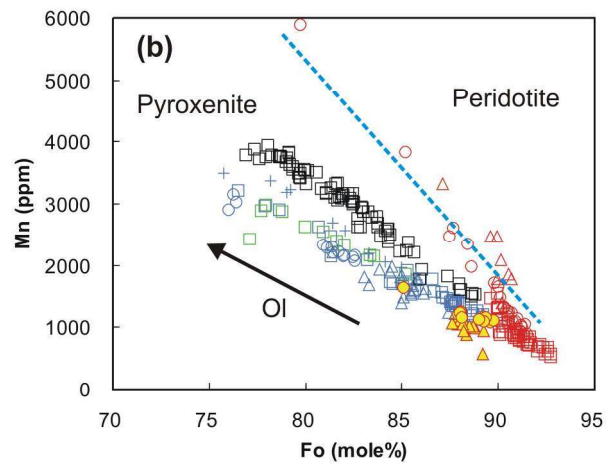
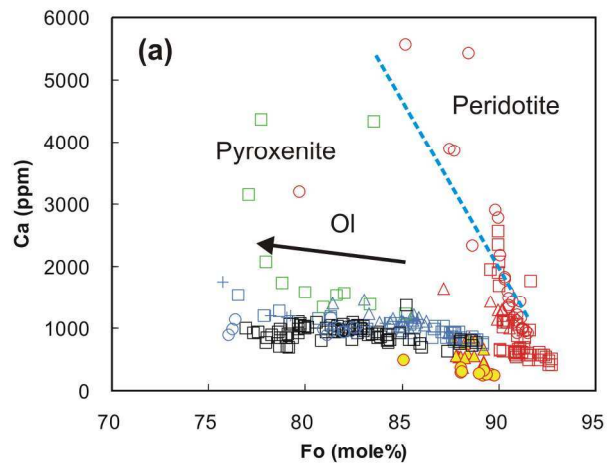


Fig. 19(a)

Fig. 19a

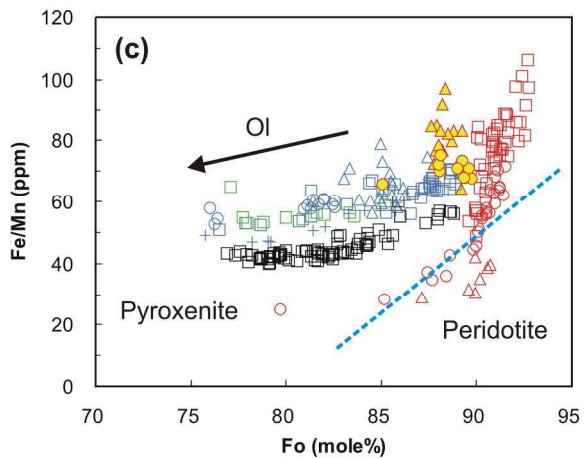


Fig. 19(b)

Fig.19b

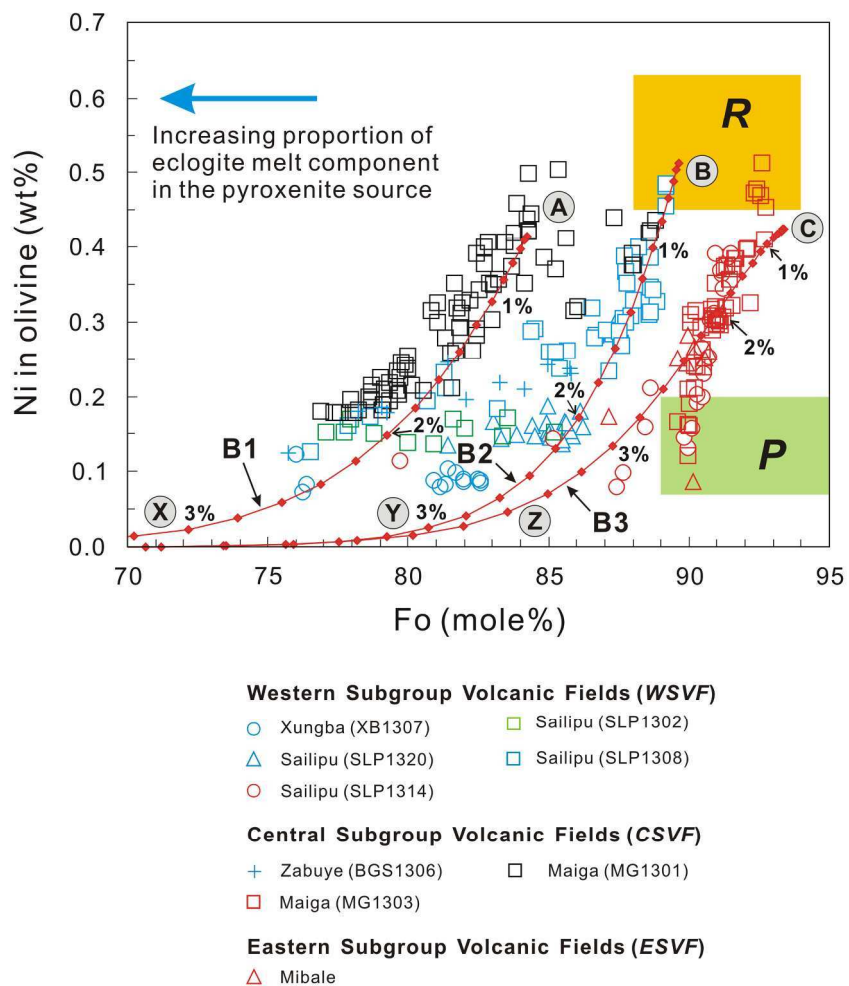


Figure 20

Fig.20

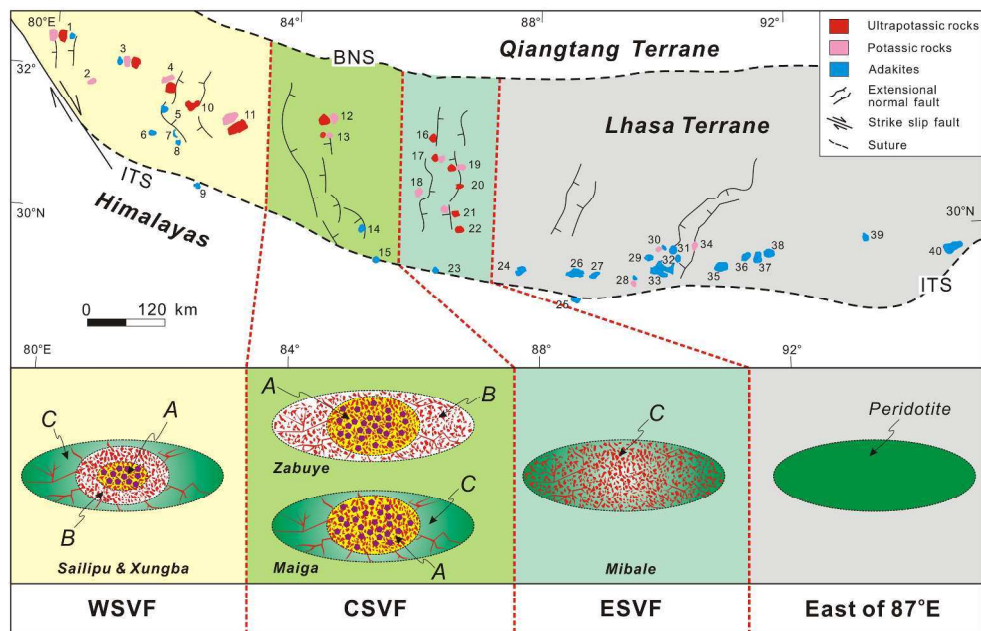


Fig. 21

Fig. 21

Review

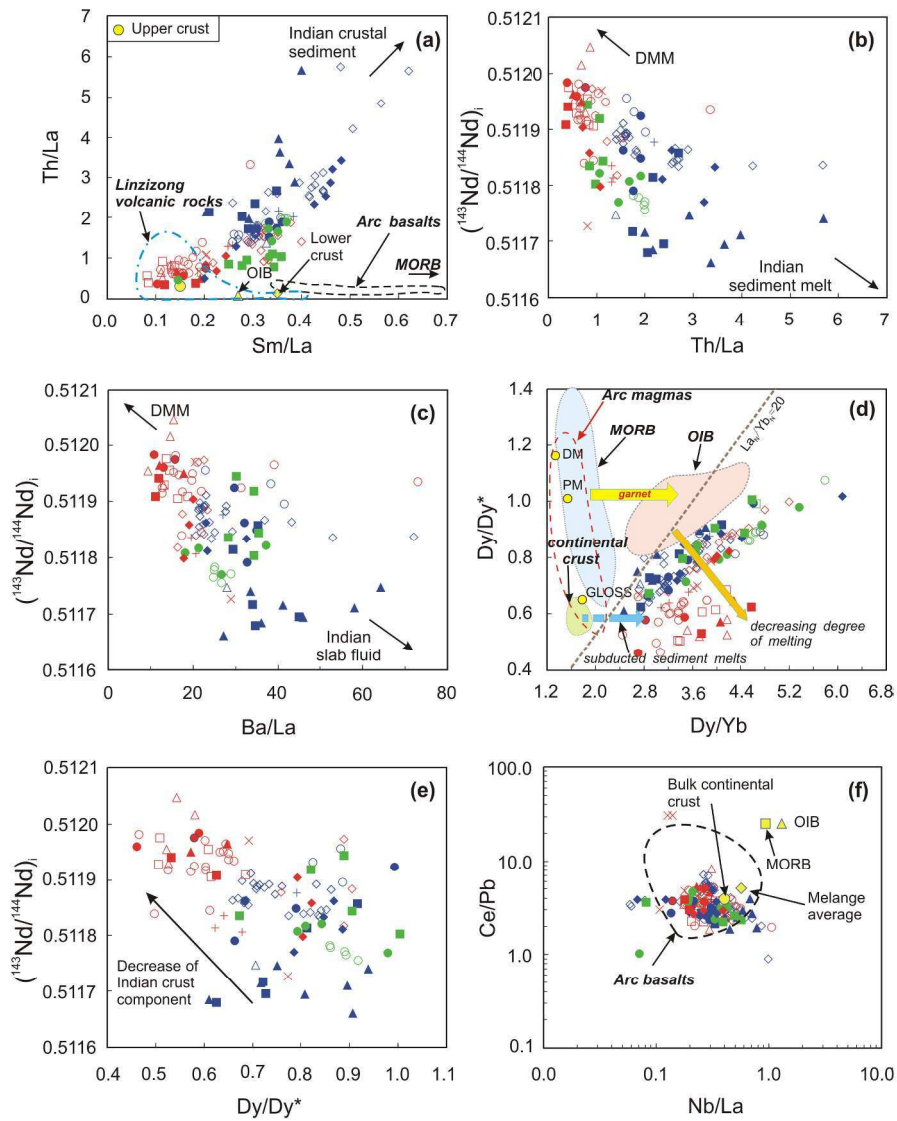


Figure 22

Fig.22

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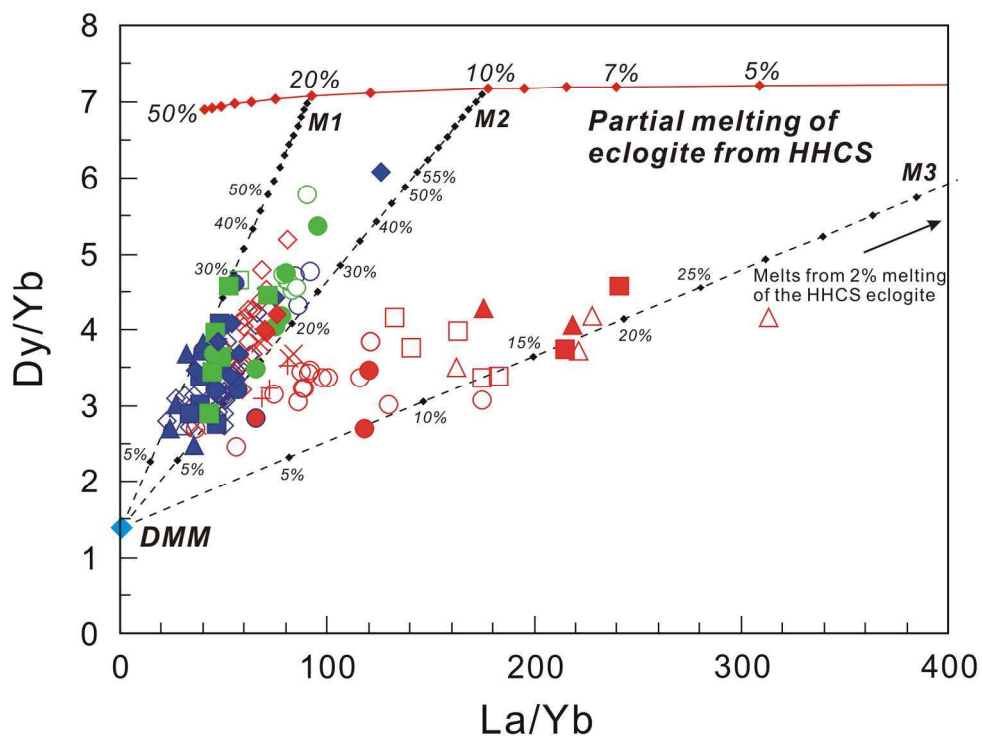


Figure 23

Fig.23

view

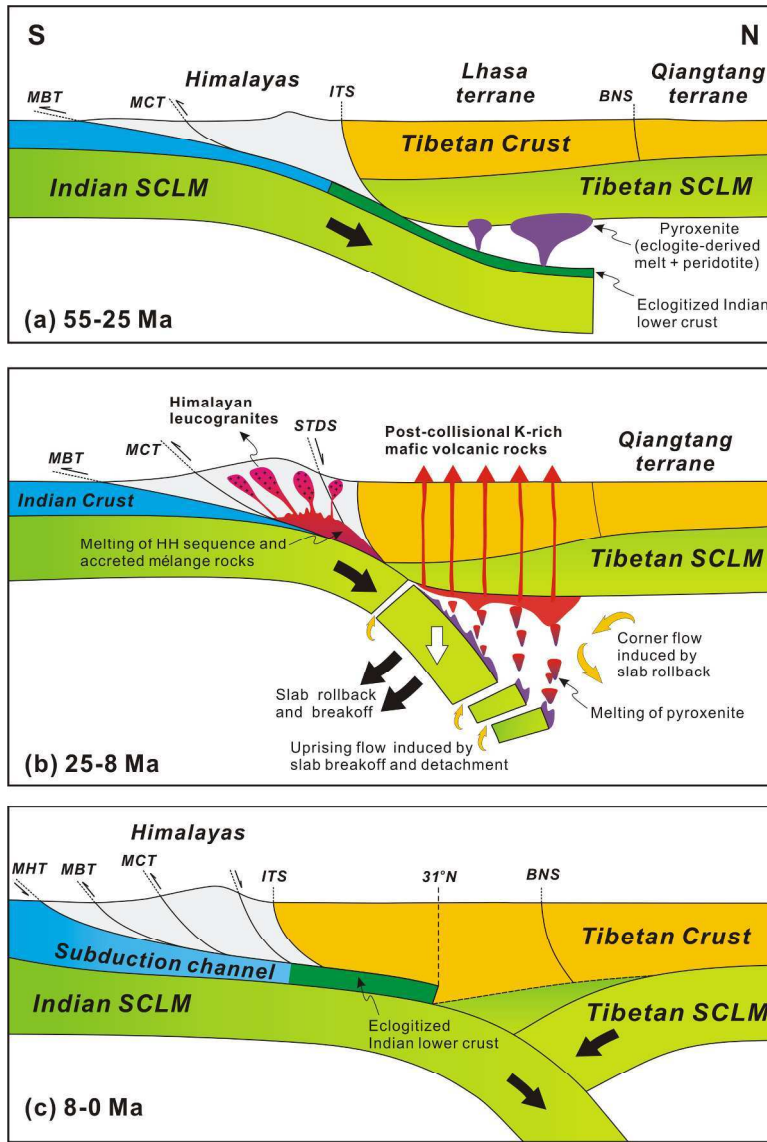


Fig. 24

Fig.24

Table 1: Ages of the post-collisional K-rich magmatism in south Tibet

Field no.	Sample no.	Field name	Dating method	Materials dated	Age (Ma)	Data sources
1	97-7-16-2pk	Shiquanhe	$^{40}\text{Ar}/^{39}\text{Ar}$	Biotite	22.6±0.1 (a)	[1]
1	JPT24C	Shiquanhe	$^{40}\text{Ar}/^{39}\text{Ar}$	Phl	24.0±1.0 (b)	[2]
1	JPT24B	Shiquanhe	$^{40}\text{Ar}/^{39}\text{Ar}$	Phl	22.3±4.7 (b)	[2]
1	JPT23	Shiquanhe	$^{40}\text{Ar}/^{39}\text{Ar}$	Phl	21.2±0.6 (b)	[2]
3	TE148/93	Bongba	$^{40}\text{Ar}/^{39}\text{Ar}$	WR	23.3±0.2 (c)	[3]
3	TE150/93	Bongba	$^{40}\text{Ar}/^{39}\text{Ar}$	WR	25.4±0.2 (c)	[3]
4	TE011/93	Xungba	$^{40}\text{Ar}/^{39}\text{Ar}$	Phl	23.0±0.3 (d)	[3]
4	TE011/93	Xungba	Rb-Sr	Phl-WR	22.4±0.3 (b)	[3]
4	TE011/93	Xungba	$^{40}\text{Ar}/^{39}\text{Ar}$	Groundmass	21.0±0.4 (c)	[3]
4	TE138/93	Xungba	$^{40}\text{Ar}/^{39}\text{Ar}$	Phl	18.1±0.3 (d)	[3]
4	TE025/93	Xungba	$^{40}\text{Ar}/^{39}\text{Ar}$	WR	22.8±0.2 (d)	[3]
4	10XB01	Xungba	U-Pb	Zircon	24.1±0.3 (e)	[4]
4	10XB06	Xungba	U-Pb	Zircon	23.3±0.4 (e)	[4]
4	10YR06	Xungba	U-Pb	Zircon	23.9±0.6 (e)	[4]
4	08YR05	Xungba	U-Pb	Zircon	23.2±0.4 (e)	[5]
4	10YR06	Xungba	U-Pb	Zircon	23.8±1.0 (e)	[5]
4	10XB10	Xungba	U-Pb	Zircon	22.7±1.3 (e)	[5]
10	TE117/93	Jarga	$^{40}\text{Ar}/^{39}\text{Ar}$	Phl	18.5±0.4 (d)	[3]
10	TE118/93	Jarga	$^{40}\text{Ar}/^{39}\text{Ar}$	Phl	18.3±0.4 (d)	[3]
11	Not mentioned	Sailipu	$^{40}\text{Ar}/^{39}\text{Ar}$	WR	15.47±0.30 (c)	[6,7]
11	Not mentioned	Sailipu	K-Ar	WR	15.24±0.30 (f)	[6]
11	Not mentioned	Sailipu	K-Ar	WR	16.07±0.32 (f)	[6]
11	Not mentioned	Sailipu	K-Ar	WR	17.8±0.3 (f)	[6,7]
11	SL0619	Sailipu	U-Pb	Zircon	17.01±3.99 (e)	[8]
11	SL0625	Sailipu	U-Pb	Zircon	16.28±0.95 (e)	[8]
11	SL0628	Sailipu	U-Pb	Zircon	17.67±0.33 (e)	[8]
11	05SLP-8-1	Sailipu	$^{40}\text{Ar}/^{39}\text{Ar}$	Phl	17.58±0.19 (d)	[9]
11	08SLP4-2	Sailipu	$^{40}\text{Ar}/^{39}\text{Ar}$	Phl	18.48±0.34 (d)	[10]
11	SLP1101	Sailipu	U-Pb	Zircon	18.1±0.6 (e)	[5]
12	Not mentioned	Zabuye	K-Ar	WR	15.8±0.3 (f)	[11]
12	Not mentioned	Zabuye	K-Ar	WR	15.9±0.3 (f)	[11]
12	ZB1	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Sanidine	16.16±0.12 (d)	[12]
12	ZB4	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Sanidine	16.12±0.12 (d)	[12]
12	ZB10	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Sanidine	16.02±0.17 (d)	[12]
12	ZB10	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Biotite	16.1±0.2 (d)	[12]
12	ZB12	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Sanidine	16.01±0.12 (d)	[12]
12	ZB12	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Biotite	16.11±0.16 (d)	[12]
12	CQ1-04-02	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	WR	15.56±0.07 (d)	[13]
12	ZB1	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Sanidine	16.23±0.05 (d)	[13]
12	ZB4	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Sanidine	16.17±0.04 (d)	[13]
12	ZB10	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Sanidine	16.02±0.03 (d)	[13]
12	ZB10	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Biotite	16.19±0.14 (d)	[13]
12	ZB12	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Sanidine	16.07±0.03 (d)	[13]
12	ZB12	Zabuye	$^{40}\text{Ar}/^{39}\text{Ar}$	Biotite	16.21±0.13 (d)	[13]

12	ZB1102	Zabuye	U-Pb	Zircon	15.7±0.4 (e)	[5]
12	ZB1106	Zabuye	U-Pb	Zircon	16.0±0.6 (e)	[5]
13	2003T405	Maiga	⁴⁰ Ar/ ³⁹ Ar	Biotite	17.4±0.1 (d)	[14]
13	CQ01	Maiga	⁴⁰ Ar/ ³⁹ Ar	WR	16.10±0.12 (c)	[15,16]
13	CQ02	Maiga	⁴⁰ Ar/ ³⁹ Ar	WR	16.59±0.13 (c)	[15,16]
16	99T60	Garwa	⁴⁰ Ar/ ³⁹ Ar	Sanidine	22.9±0.7 (d)	[17]
16	99T62	Garwa	⁴⁰ Ar/ ³⁹ Ar	Sanidine	17.8±0.3 (c)	[17]
17	DR01-01	Yaqian	⁴⁰ Ar/ ³⁹ Ar	Biotite	13.4±0.2 (d)	[18]
17	DR01-02	Yaqian	⁴⁰ Ar/ ³⁹ Ar	Sanidine	13.7±0.2 (d)	[18]
17	DR03	Yaqian	⁴⁰ Ar/ ³⁹ Ar	Sanidine	14.2±0.3 (d)	[18]
17	DR04	Yaqian	⁴⁰ Ar/ ³⁹ Ar	Sanidine	13.5±0.2 (d)	[18]
19	8030-5	Mibale	K-Ar	WR	19.04±0.97 (f)	[19, 20]
19	8030-5	Mibale	K-Ar	WR	12.60±0.97 (f)	[19, 21]
19	8030-18	Mibale	K-Ar	WR	14.22±0.68 (f)	[19]
20	2003T536	Yiqian	⁴⁰ Ar/ ³⁹ Ar	Biotite	13.5±0.1 (d)	[14]
20	DR1104	Yiqian	U-Pb	Zircon	12.9±0.5 (e)	[5]
20	DR1114	Yiqian	U-Pb	Zircon	11.2±0.9 (e)	[5]
21	99T132	Chazi	⁴⁰ Ar/ ³⁹ Ar	Phl	13.3±0.4 (d)	[17]
21	99T145	Chazi	⁴⁰ Ar/ ³⁹ Ar	Sanidine	8.2±0.5 (d)	[17]
21	99T154	Chazi	⁴⁰ Ar/ ³⁹ Ar	Sanidine	13.1±0.3 (d)	[17]
21	XR01-03	Chazi	⁴⁰ Ar/ ³⁹ Ar	Biotite	11.2±0.3 (a)	[18]
21	XR01-3	Chazi	⁴⁰ Ar/ ³⁹ Ar	Biotite	11.5±0.2 (d)	[15]
21	G8	Chazi	U-Pb	Zircon	11.70±0.15 (g)	[22]
22	JPT7	Pabbai	⁴⁰ Ar/ ³⁹ Ar	Phl	18.3±2.7 (b)	[23]
22	T2A	Pabbai	⁴⁰ Ar/ ³⁹ Ar	Phl	13.8±3.5 (b)	[23]
22	T3B	Pabbai	⁴⁰ Ar/ ³⁹ Ar	Biotite	13.8±0.3 (b)	[23]
22	T5A	Pabbai	⁴⁰ Ar/ ³⁹ Ar	Phl	13.3±0.8 (b)	[23]

Field no. refers to number of the volcanic field in Fig. 1. Field name corresponds to that in Fig.1.

Sources of the age data for the magmatic rocks are as follows: [1] Kapp *et al.* (2003); [2] Williams *et al.* (2004); [3] Miller *et al.* (1999); [4] Liu *et al.* (2011); [5] Liu *et al.* (2014); [6] Geological Survey Bureau (2006) ; [7] Sun *et al.* (2007); [8] Sun *et al.* (2008); [9] Wang *et al.* (2008); [10] Chen *et al.* (2012); [11] Ma *et al.* (2002); [12] Nomade *et al.* (2004); [13] Zhao *et al.* (2009a); [14] Ding *et al.* (2006); [15] Zhao *et al.* (2009b); [16] Mo *et al.* (2006); [17] Ding *et al.* (2003); [18] Zhao *et al.* (2006); [19] Xie *et al.* (2004); [20] Gao *et al.* (2007); [21] Liao *et al.* (2002); [22] Guo *et al.* (2013); [23] Williams *et al.* (2001) Age categories of the K-rich magmatic rocks, which are shown by letters in parentheses, are as follows: (a) inverse isochron age; (b) isochron age; (c) total gas age; (d) plateau age; (e) weighted mean age of LA-ICP-MS; (f) K-Ar age; (g) isochron age of SHRIMP.

Phl: Phlogopite; WR: Whole-rock.

Table 2: Phenocryst and groundmass minerals of the K-rich mafic magmatic rocks in south Tibet

Field no.	Sample no.	Field name	Faces	Mg-no.	Phenocrysts	Groundmass
1	GZ-S05	Shiquanhe	lava flow	0.72	Cpx+Ol+Phl	Phl+Cpx+Sa+Fe-Ti+G
1	GZ-S06	Shiquanhe	lava flow	0.72	Phl+Cpx+Sa	Cpx+Phl+Sa+G
1	GZ-S12	Shiquanhe	lava flow	0.76	Ol+Phl+Cpx+Fe-Ti	Cpx+Phl+Sa+Fe-Ti+G
1	GZ-S15	Shiquanhe	lava flow	0.70	Phl+Cpx+Fe-Ti	Phl+Cpx+Sa+G
1	GZ-S27	Shiquanhe	lava flow	0.72	Ol+Cpx+Fe-Ti	Cpx+Ol+Phl+Sa+Fe-Ti
1	SQH-08	Shiquanhe	lava flow	0.78	Ol+Phl+Cpx	Ol+Phl+Cpx+Sa+Fe-Ti+G
1	SQH-15	Shiquanhe	lava flow	0.74	Ol+Phl+Cpx+Fe-Ti	Cpx+Sa+Phl+Ap
3	16YS	Bongba	lava flow	0.75	Ol+Phl+Cpx+Sa	Phl+Cpx+Sa+Ap
3	20YS	Bongba	lava flow	0.74	Phl+Ol+Cpx+Ap	Cpx+Phl+Sa+Fe-Ti+G
3	23YS	Bongba	lava flow	0.72	Ol+Phl+Cpx+Ap	Cpx+Sa+Phl+Fe-Ti+Ap+G
3	25YS	Bongba	lava flow	0.77	Ol+Cpx+Phl	Ol+Cpx+Phl+Sa+Ap+Pl+G
3	JS07	Bongba	lava flow	0.74	Cpx+Ol+Phl	Cpx+Sa+Phl+Ap+Fe-Ti
4	99-B06	Xungba	lava flow	0.80	Ol+Cpx+Phl+Fe-Ti	Sa+Cpx+Phl+Fe-Ti
4	99-B09	Xungba	lava flow	0.73	Ol+Cpx+Phl	Phl+Ol+Sa+Cpx+G
4	99-B23	Xungba	lava flow	0.74	Ol+Cpx+Phl	Cpx+Sa+Phl+Fe-Ti+Ap
4	99-C12	Xungba	lava flow	0.74	Ol+Phl+Cpx	Cpx+Ol+Sa+Pl+Fe-Ti+Ap
11	99-C15	Sailipu	lava flow	0.76	Ol+Cpx+Phl	Ol+Phl+Cpx+G
11	99-C28	Sailipu	lava flow	0.76	Ol+Cpx+Sa	Ol+Phl+Cpx+Sa
11	XB-06	Sailipu	lava flow	0.76	Ol+Cpx+Phl	Ol+Phl+Cpx+Sa
11	XB-24	Sailipu	lava flow	0.78	Cpx+Ol+Phl	Ol+Phl+Cpx+Ap
12	BG015	Zabuye	lava flow	0.83	Ol+Cpx+Phl+Sa	Ol+Sa+Cpx+Ap+G
12	BG026	Zabuye	lava flow	0.77	Ol+Phl+Sa+Cpx	Sa+Cpx+Phl+G
12	BG037	Zabuye	lava flow	0.77	Cpx+Ol+Phl+Sa	Cpx+Ol+Sa+Fe-Ti
12	BG201	Zabuye	lava flow	0.72	Sa+Cpx+Phl	Sa+Cpx+Pl+Ap
12	BG203	Zabuye	lava flow	0.80	Ol+Cpx+Sa+Phl	Ol+Cpx+Sa+Ap
13	MG-008	Maiga	lava flow	0.88	Ol+Phl	Ol+Sa+G
13	MG-012	Maiga	plug	0.81	Ol+Phl	Cpx+Sa+Ap+Fe-Ti
13	MG-015	Maiga	plug	0.79	Ol+Phl	Ol+Cpx+Phl+G
13	MG-016	Maiga	dyke	0.76	Ol+Cpx	Cpx+Sa+Fe-Ti
16	GZF3	Garwa	lava flow	0.79	Phl+Cpx+Lc+Ol	Sa+Cpx+Ap+Fe-Ti
16	GZF9	Garwa	lava flow	0.74	Ol+Phl+Cpx+Lc	Cpx+Sa+Ap+Fe-Ti
17	G-029	Yaqian	lava flow	0.79	Phl+Cpx	Cpx+Sa+Fe-Ti
17	GH03	Yaqian	lava flow	0.79	Cpx+Ol	Ol+Cpx+Ap+Lc+G
19	GH05	Mibale	lava flow	0.71	Ol+Cpx+Phl+Lc	Phl+Sa+Cpx+Lc
19	GH09	Mibale	lava flow	0.77	Phl+Cpx+Sa	Sa+Cpx+Fe-Ti
21	ZFC004	Chazi	lava flow	0.75	Phl+Cpx	Phl+Sa+G+Fe-Ti
21	ZFC008	Chazi	lava flow	0.72	Ol+Cpx	Phl+Cpx+Ol+G
21	ZFC009	Chazi	lava flow	0.76	Phl+Cpx+Lc	Cpx+Ol+Lc
22	GB036	Pabbai	dyke	0.75	Phl+Ol+Cpx	Phl+Cpx+Ol+Sa
4*	XB1307	Xungba	lava flow	0.72	Cpx+Ol+Phl	Cpx+Ol+Phl+Sa+Fe-Ti
11*	SLP1314	Sailipu	lava flow	0.80	Ol+Phl+Cpx	Ol+Phl+Cpx+Fe-Ti
11	SLP1302	Sailipu	lava flow	0.77	Ol+Phl+Cpx	Phl+Cpx+Sa+Ap
11	SLP1308	Sailipu	lava flow	0.73	Phl+Cpx+Ol	Phl+Cpx+Ap+Fe-Ti
11	SLP1320	Sailipu	lava flow	0.77	Ol+Phl+Cpx	Ol+Phl+Cpx+Lc+G
12	BGS1306	Zabuye	lava flow	0.69	Phl+Cpx	Phl+Cpx+Ap+Fe-Ti
13*	MG1301	Maiga	plug	0.81	Ol+Cpx	Ol+Cpx+Ap+Fe-Ti
13*	MG1303	Maiga	lava flow	0.82	Ol+Cpx	Ol+Cpx+Ap

Field no. refers to number of the volcanic field in Fig. 1. Field name corresponds to that in Fig. 1.

Ap: Apatite; Cpx: clinopyroxene; Fe-Ti: Fe-Ti oxides; G: glass; Lc: Leucite; Ol: olivine; Phl: phlogopite; Pl: plagioclase; Sa: sanidine.

Samples (XB1307, SLP1314, SLP1302, SLP1308, SLP1320, BGS1306, MG1301, MG1303) have been analysed for olivine phenocryst compositions (Appendix A).

*Samples contain xenocrystic olivine.

Table 3: Major and trace element analyses of the K-rich mafic magmatic rocks in south Tibet

Field no.:	1	1	1	1	1	1	1	3
Sample no.:	GZ-S05	GZ-S06	GZ-S12	GZ-S15	GZ-S27	SQH-08	SQH-15	16YS
Field name:	Shiquanhe	Shiquanhe	Shiquanhe	Shiquanhe	Shiquanhe	Shiquanhe	Shiquanhe	Bongba
Latitude(N)	32°28' 12"	32°28' 33.6"	32°47' 27.6"	32°31' 8.4"	32°38' 52.8"	32°48' 57.6"	32°28' 1.2"	32°6' 3.6"
Longitude(E)	80°6' 0"	80°12' 36"	80°10' 48"	80°7' 1.2"	80°3' 36"	80°8' 49.2"	80°16' 44.4"	81°19' 51.6"
Age (Ma):	22.5	22.5	22.5	22.5	22.5	22.5	22.5	24.4
SiO ₂	55.97	56.35	59.74	57.01	54.90	57.23	57.47	56.65
TiO ₂	0.93	1.26	0.56	1.17	0.90	1.58	1.35	1.52
Al ₂ O ₃	11.55	12.14	12.81	13.35	12.92	11.45	12.15	11.61
TFe ₂ O ₃ *	6.77	6.84	5.55	7.19	6.92	5.14	6.37	6.69
MnO	0.14	0.12	0.07	0.08	0.13	0.08	0.07	0.16
MgO	6.96	7.16	7.04	6.83	7.34	7.47	7.21	8.23
CaO	8.03	7.32	5.59	4.75	5.45	7.01	8.05	6.57
Na ₂ O	2.57	2.15	2.50	2.05	2.39	3.01	2.16	1.74
K ₂ O	6.21	5.78	5.18	6.87	8.05	6.51	4.76	5.92
P ₂ O ₅	0.87	0.88	0.97	0.71	0.99	0.53	0.41	0.92
LOI	1.69	2.15	1.08	0.69	1.07	1.34	0.89	1.47
Mg-no.	0.72	0.72	0.76	0.70	0.72	0.78	0.74	0.75
La	59.1	55.4	56.3	40.7	51.6	50.8	46.2	68.2
Ce	142	143	131	108	135	124	116	179
Pr	17.4	19.2	17.4	14.9	21.3	18.0	17.1	25.7
Nd	65.3	78.4	69.0	60.8	83.2	75.2	71.3	103
Sm	12.1	19.6	16.4	16.3	19.4	18.1	17.9	23.8
Eu	1.84	2.45	2.15	3.06	2.73	2.21	2.52	2.38
Gd	6.89	10.1	8.58	11.7	12.6	10.8	11.9	12.0
Tb	0.83	1.17	0.95	1.04	1.16	1.27	1.35	1.19
Dy	4.10	5.32	4.47	4.65	4.88	5.63	5.19	5.75
Ho	0.70	0.75	0.73	0.72	0.73	0.91	0.87	0.82
Er	1.74	1.63	1.71	1.58	1.60	2.24	1.92	1.69
Tm	0.26	0.22	0.24	0.20	0.21	0.30	0.28	0.23
Yb	1.66	1.43	1.52	1.26	1.28	1.87	1.93	1.41
Lu	0.25	0.21	0.23	0.19	0.20	0.29	0.30	0.22
Sc	17.6	33.3	10.3	38.8	21.6	28.4	19.1	23.5
V	108	139	168	146	125	153	171	114
Cr	390	473	563	453	422	362	443	513
Co	35.3	39.2	42.4	32.5	27.7	50.4	42.3	26.2
Ni	364	362	408	350	273	236	319	395
Cu	53.9	69.1	64.4	76.9	58.9	35.1	54.1	46.8
Zn	61.0	84.7	69.4	92.6	63.3	61.3	103	68.3
Ga	22.2	24.2	24.4	23.6	18.8	21.1	23.4	19.8
Rb	683	1030	674	719	643	672	841	917
Sr	503	683	481	735	691	411	331	464
Y	23.4	22.0	17.9	15.7	16.6	21.9	28.6	17.9
Zr	461	505	450	374	495	355	476	495
Nb	24.2	28.8	25.2	28.7	34.9	29.7	35.8	22.6
Ba	2138	3216	2320	1369	1407	2335	2969	2409
Hf	12.6	13.0	9.16	10.2	13.5	12.3	11.7	13.1
Ta	1.78	1.57	1.88	1.54	1.63	1.86	1.79	1.43
Pb	56.1	57.3	70.6	39.3	34.3	42.0	59.8	61.1
Th	128	221	112	232	174	185	135	184
U	14.5	21.4	20.9	18.4	23.0	16.8	16.5	21.8

Table 3: continued

Field no.:	3	3	3	3	4	4	4	4
Sample no.:	20YS	23YS	25YS	JS07	99-B06	99-B09	99-B23	99-C12
Field name:	Bongba	Bongba	Bongba	Bongba	Xungba	Xungba	Xungba	Xungba
Latitude(N)	32°8'27.6"	32°12'14.4"	32°4'1.2"	32°4'55.2"	32°0.72'	32°3'39.6"	32°0.24'	32°2'27.6"
Longitude(E)	81°22'19.2"	81°27'54"	81°21'50.4"	81°25'4.8"	81°47'27.6"	81°54'14.4"	81°52'33.6"	81°54'3.6"
Age (Ma):	24.4	24.4	24.4	24.4	22.6	22.6	22.6	22.6
SiO ₂	59.17	57.55	55.12	58.03	55.43	58.14	56.41	53.73
TiO ₂	1.10	1.19	1.19	1.25	1.47	1.24	1.06	0.98
Al ₂ O ₃	12.27	12.59	11.59	12.00	11.99	12.67	13.47	13.67
TFe ₂ O ₃ *	5.76	6.68	6.91	6.30	6.05	6.15	6.11	7.14
MnO	0.11	0.12	0.12	0.15	0.12	0.08	0.12	0.13
MgO	6.79	6.90	9.45	7.33	9.62	6.84	7.09	8.11
CaO	5.62	5.73	6.78	6.15	6.55	4.39	4.98	5.41
Na ₂ O	2.83	2.72	1.92	2.18	2.12	2.30	2.24	2.27
K ₂ O	5.97	6.16	6.07	5.36	5.48	7.34	7.69	7.86
P ₂ O ₅	0.38	0.35	0.86	1.23	1.16	0.86	0.83	0.69
LOI	1.28	0.63	0.92	1.45	1.59	1.64	1.87	0.91
Mg-no.	0.74	0.72	0.77	0.74	0.80	0.73	0.74	0.74
La	90.4	85.2	80.0	65.4	84.7	97.5	113	95.4
Ce	209	226	198	163	204	245	273	234
Pr	27.2	29.3	25.2	22.4	29.8	33.4	39.1	33.6
Nd	106	111	82.7	86.4	132	138	158	143
Sm	25.1	24.7	16.8	19.9	30.5	32.5	37.7	33.2
Eu	3.82	2.77	2.71	2.66	4.47	4.39	4.38	4.69
Gd	19.0	17.2	13.4	11.2	17.6	18.5	19.8	20.3
Tb	1.62	1.79	1.48	1.24	1.84	1.74	1.95	1.82
Dy	5.34	6.58	6.96	5.62	7.05	7.02	6.34	6.21
Ho	0.89	0.96	1.04	0.96	0.85	1.15	0.92	0.95
Er	2.04	2.16	2.35	2.24	1.68	2.39	2.25	2.26
Tm	0.29	0.32	0.31	0.29	0.23	0.33	0.32	0.31
Yb	1.94	2.19	2.06	1.94	1.53	1.99	1.98	2.17
Lu	0.32	0.30	0.31	0.32	0.25	0.29	0.29	0.32
Sc	28.9	29.0	21.1	27.8	38.3	19.2	26.2	23.5
V	267	246	136	148	318	198	119	137
Cr	459	563	584	655	584	352	419	527
Co	68.5	54.2	32.3	48.3	61.4	25.6	27.1	29.4
Ni	523	421	347	434	527	317	247	324
Cu	51.4	76.9	63.8	71.5	61.1	62.5	57.9	82.0
Zn	83.1	70.4	72.2	65.2	50.1	58.3	69.2	73.9
Ga	22.8	25.6	21.5	23.9	21.1	19.8	20.8	21.5
Rb	1428	1098	840	1025	916	675	719	759
Sr	435	452	615	657	809	716	1079	829
Y	20.1	22.0	21.4	18.3	16.6	23.1	21.1	25.2
Zr	401	386	480	416	538	718	580	635
Nb	26.3	21.6	16.1	21.7	25.3	24.0	24.2	12.9
Ba	3128	2899	2358	2944	2520	3418	3625	3124
Hf	11.0	13.4	11.5	19.3	13.5	20.4	13.7	17.9
Ta	1.96	2.51	0.77	2.41	1.68	1.81	1.58	1.13
Pb	79.3	81.4	58.6	76.5	86.7	93.8	73.3	85.3
Th	185	149	173	156	162	186	174	168
U	18.7	18.9	24.4	19.1	24.9	21.4	22.2	18.3

Table 3: continued

Field no.:	11	11	11	11	12	12	12	12
Sample no.:	99-C15	99-C28	XB-06	XB-24	BG015	BG026	BG037	BG201
Field name:	Sailipu	Sailipu	Sailipu	Sailipu	Zabuye	Zabuye	Zabuye	Zabuye
Latitude(N)	31°18'6.13"	31°20'46.51"	31°24'50.29"	31°20'13.42"	31°26'24"	31°23'48"	31°24'24"	31°24'18"
Longitude(E)	82°44'50.35"	82°58'28.4"	82°45'23.43"	82°52'37.16"	84°23'54"	84°18'6"	84°18'12"	84°18'18"
Age (Ma):	16.97	16.97	16.97	16.97	16.02	16.02	16.02	16.02
SiO ₂	55.45	58.02	54.64	56.78	53.73	54.23	53.87	54.29
TiO ₂	1.66	1.29	1.49	1.52	1.13	0.91	1.52	1.38
Al ₂ O ₃	13.48	13.61	12.99	11.47	13.91	11.73	13.57	12.08
TFe ₂ O ₃ *	6.35	6.34	6.14	5.98	4.58	6.62	5.50	7.28
MnO	0.11	0.09	0.12	0.13	0.26	0.12	0.15	0.13
MgO	8.08	8.28	7.97	8.75	8.91	9.14	7.49	7.58
CaO	5.86	4.25	5.62	4.85	7.24	7.25	6.50	7.73
Na ₂ O	1.90	1.65	2.71	2.29	1.59	1.78	1.78	1.27
K ₂ O	6.47	5.75	7.50	6.96	7.46	7.09	8.27	7.07
P ₂ O ₅	0.63	0.73	0.82	1.28	1.18	1.13	1.34	1.19
LOI	0.92	1.63	1.85	1.79	1.83	0.76	1.68	1.84
Mg-no.	0.76	0.76	0.76	0.78	0.83	0.77	0.77	0.72
La	90.2	78.4	109	84.1	97.5	127	74.1	82.9
Ce	237	202	287	225	262	324	189	190
Pr	34.9	29.2	42.3	33.2	33.8	46.6	27.0	26.3
Nd	161	138	187	152	125	204	111	108
Sm	38.4	36.4	49.0	40.6	28.0	43.4	25.2	23.1
Eu	6.18	4.51	6.73	7.06	4.47	4.63	3.78	4.12
Gd	26.2	22.6	27.2	32.9	17.6	14.3	12.3	11.9
Tb	2.52	1.48	2.43	2.37	1.83	1.74	1.43	1.46
Dy	6.84	5.04	6.41	7.92	8.52	7.89	6.39	5.57
Ho	0.78	0.66	0.72	1.22	1.28	1.24	1.02	0.85
Er	1.92	1.53	1.63	2.78	2.87	2.55	2.31	2.03
Tm	0.26	0.21	0.23	0.36	0.34	0.31	0.28	0.29
Yb	1.68	1.37	1.45	2.29	1.86	1.77	1.61	1.93
Lu	0.23	0.21	0.22	0.33	0.26	0.24	0.20	0.29
Sc	19.8	24.2	17.3	16.3	21.6	21.0	19.2	17.1
V	172	153	164	183	126	116	119	159
Cr	538	332	429	358	553	493	361	348
Co	40.5	25.9	24.5	39.2	33.2	35.1	26.8	41.2
Ni	387	263	347	353	465	315	475	390
Cu	42.2	39.6	52.0	66.6	76.8	53.3	72.0	68.3
Zn	69.5	101	50.3	91.7	52.7	37.2	46.2	34.3
Ga	23.2	24.6	23.9	26.4	21.1	19.4	20.9	24.2
Rb	867	643	827	709	621	710	583	457
Sr	801	672	637	758	1638	1619	1696	1718
Y	18.5	16.9	19.1	36.9	20.1	25.3	15.8	32.7
Zr	637	569	613	784	421	483	386	415
Nb	18.3	41.4	13.3	16.4	36.6	10.2	42.0	41.0
Ba	2113	2110	2508	2739	3351	3817	2634	2351
Hf	9.03	14.9	19.8	20.3	12.2	6.56	12.1	10.4
Ta	1.51	2.67	1.24	1.65	1.89	0.58	2.33	2.75
Pb	71.1	84.2	73.8	79.7	86.3	89.2	80.4	72.9
Th	212	253	279	289	95.2	102	83.5	70.2
U	36.2	37.4	39.3	41.2	13.5	7.21	12.4	14.6

Table 3: continued

Field no.:	12	13	13	13	13	16	16	17
Sample no.:	BG203	MG-008	MG-012	MG-015	MG-016	GZF3	GZF9	G-029
Field name:	Zabuye	Maiga	Maiga	Maiga	Maiga	Garwa	Garwa	Yaqian
Latitude(N)	31°24'24"	30°48'45.27"	30°48'45.27"	30°49'22.31"	30°49'22.31"	31°16'4.8"	31°15'14.4"	30°56'40.8"
Longitude(E)	84°18'18"	84°25'47.59"	84°25'47.59"	84°25'20.46"	84°25'20.46"	86°33'57.6"	86°34'48.36"	86°23'33.46"
Age (Ma):	16.02	16.7	16.7	16.7	16.7	20.35	20.35	13.70
SiO ₂	53.40	53.62	52.65	54.95	53.11	55.59	55.13	55.97
TiO ₂	1.12	1.52	1.27	1.32	1.38	1.67	1.70	1.58
Al ₂ O ₃	11.69	11.85	11.12	10.86	12.01	12.40	12.54	11.97
TFe ₂ O ₃ *	6.17	4.62	6.35	6.20	7.16	5.55	6.47	5.60
MnO	0.15	0.15	0.11	0.08	0.14	0.12	0.13	0.05
MgO	9.83	13.10	11.29	9.33	9.26	8.63	7.43	8.33
CaO	7.45	4.65	6.73	6.72	6.94	4.59	6.19	6.96
Na ₂ O	1.61	2.17	1.66	1.50	1.40	1.39	1.42	1.30
K ₂ O	7.63	7.12	7.65	7.75	7.28	8.89	7.99	7.43
P ₂ O ₅	0.94	1.19	1.16	1.29	1.32	1.18	1.00	0.81
LOI	1.24	1.68	0.84	1.57	1.96	1.38	1.85	1.83
Mg-no.	0.80	0.88	0.81	0.79	0.76	0.79	0.74	0.79
La	101	96.8	163	148	153	271	267	353
Ce	263	257	435	412	398	574	545	735
Pr	38.9	35.5	66.3	61.5	56.8	76.5	72.9	91.6
Nd	173	148	297	284	244	285	280	359
Sm	36.3	32.2	57.2	54.6	51.9	38.8	51.2	64.4
Eu	4.68	5.16	7.48	7.76	7.19	5.49	6.84	6.18
Gd	15.0	16.8	25.1	26.0	24.3	15.6	31.6	16.6
Tb	1.71	1.95	2.44	2.28	2.43	1.36	2.63	1.58
Dy	7.63	7.84	8.73	7.89	8.64	5.03	6.51	6.15
Ho	1.19	1.16	1.28	1.16	1.26	0.70	0.78	0.84
Er	2.58	2.57	2.67	2.44	2.65	1.44	1.67	1.86
Tm	0.34	0.35	0.33	0.29	0.30	0.19	0.23	0.25
Yb	2.09	2.13	2.16	1.89	1.61	1.24	1.52	1.64
Lu	0.30	0.32	0.34	0.28	0.19	0.18	0.21	0.24
Sc	19.8	18.3	21.6	16.4	20.9	22.4	23.1	17.0
V	136	147	137	196	150	149	178	138
Cr	458	543	574	539	592	557	419	360
Co	42.8	32.7	40.7	34.0	35.8	47.9	26.2	46.8
Ni	327	379	401	357	441	296	211	218
Cu	45.9	44.2	59.6	38.7	72.3	61.4	56.0	62.2
Zn	49.2	97.3	47.0	33.0	82.4	87.3	65.3	71.5
Ga	22.8	26.8	22.4	19.4	25.8	19.4	24.5	25.9
Rb	553	521	532	537	610	371	324	463
Sr	1527	1339	1027	949	913	1448	1529	1420
Y	23.6	19.7	27.4	30.1	13.9	16.8	14.2	21.4
Zr	587	831	695	658	717	1515	1143	1038
Nb	43.4	32.1	31.6	28.4	29.7	80.3	61.1	63.0
Ba	3476	3603	2943	3153	4095	4820	3242	4179
Hf	8.98	28.7	23.1	18.9	25.7	49.0	36.3	27.4
Ta	2.44	2.77	1.95	1.81	2.45	5.51	5.42	3.82
Pb	82.1	109	113	115	101	125	102	189
Th	107	103	272	283	221	186	140	140
U	10.3	23.7	37.7	34.1	31.3	21.9	18.4	17.3

Table 3: continued

Field no.:	17	19	19	21	21	21	22	4
Sample no.:	GH03	GH05	GH09	ZFC004	ZFC008	ZFC009	GB036	XB1307
Field name:	Yaqian	Mibale	Mibale	Chazi	Chazi	Chazi	Pabbai	Xungba
Latitude(N)	30°56'22.41"	30°50'54.29"	30°51'10.11"	30°50'41.34"	30°34'35.43"	30°07'12.47"	29°23'5.19"	31°58'41.68"
Longitude(E)	86°23'31.04"	86°39'46.37"	86°40'10.27"	86°40'10.30"	86°41'20.41"	86°32'17.18"	86°43'18.35"	81°58'11.02"
Age (Ma):	13.70	15.29	15.29	11.50	11.50	11.50	14.80	22.6
SiO ₂	56.71	56.57	54.53	53.81	55.63	54.39	53.39	58.05
TiO ₂	1.61	1.54	1.60	1.73	1.56	1.57	1.27	1.11
Al ₂ O ₃	11.84	12.33	14.17	11.87	12.44	12.85	12.77	14.02
TFe ₂ O ₃ *	5.32	7.03	5.86	6.46	6.97	6.65	6.44	6.04
MnO	0.08	0.13	0.05	0.12	0.11	0.13	0.09	0.09
MgO	7.97	6.99	7.88	8.03	7.10	8.40	7.79	6.28
CaO	6.20	6.14	6.45	7.71	6.51	6.70	6.48	5.20
Na ₂ O	1.53	1.62	1.36	2.34	1.75	1.73	2.45	2.53
K ₂ O	8.02	6.71	6.91	6.77	6.86	6.41	8.10	5.87
P ₂ O ₅	0.72	0.95	1.19	1.16	1.07	1.16	1.20	0.81
LOI	1.52	1.67	1.89	2.27	1.83	2.36	3.25	1.06
Mg-no.	0.79	0.71	0.77	0.75	0.72	0.76	0.75	0.72
La	319	234	213	164	149	152	139	76.4
Ce	571	486	403	352	347	341	308	170
Pr	65.8	60.8	47.1	47.3	43.8	48.3	41.0	24.8
Nd	226	214	136	186	173	191	152	115
Sm	37.8	36.8	21.9	36.7	30.2	37.2	28.3	23.5
Eu	5.53	4.68	3.48	4.91	5.63	5.42	5.24	3.69
Gd	17.8	12.3	14.0	23.0	24.2	22.1	20.9	12.8
Tb	1.53	1.36	1.72	2.32	2.46	2.28	1.59	1.29
Dy	6.03	5.35	6.15	9.16	8.24	8.76	6.03	5.31
Ho	0.75	0.88	0.98	1.39	1.26	1.34	1.07	0.84
Er	1.55	2.05	2.26	3.16	2.67	2.91	2.43	2.02
Tm	0.21	0.29	0.28	0.38	0.33	0.35	0.32	0.27
Yb	1.32	1.98	1.77	2.31	1.96	2.19	2.13	1.66
Lu	0.20	0.28	0.27	0.34	0.29	0.31	0.31	0.25
Sc	18.3	16.3	19.3	19.8	18.4	19.0	22.7	17.6
V	136	144	171	120	133	148	172	124
Cr	356	515	590	574	486	536	465	244
Co	47.5	33.4	31.0	67.5	29.7	43.2	47.7	25.1
Ni	262	364	271	504	314	483	517	106
Cu	63.3	52.1	48.0	48.4	51.3	41.7	61.2	34.4
Zn	76.0	64.3	92.8	67.1	58.8	62.6	65.7	79.8
Ga	22.5	21.2	32.4	23.5	21.2	23.6	22.7	22.0
Rb	342	427	695	531	464	428	386	443
Sr	1294	1016	873	706	824	855	766	710
Y	15.0	28.7	33.9	34.9	32.3	35.4	25.7	19.3
Zr	890	844	729	729	571	893	564	372
Nb	62.4	62.7	57.2	43.2	31.7	59.4	19.3	23.4
Ba	3557	3069	2298	3271	2850	2713	2185	2268
Hf	25.8	25.6	16.4	23.8	20.0	26.4	11.7	10.9
Ta	4.81	3.02	2.18	1.98	2.01	3.39	1.51	1.40
Pb	191	94.6	110	120	128	115	83.2	53.3
Th	117	138	82.2	116	127	163	108	133
U	16.7	18.4	21.6	31.7	25.2	27.8	20.1	20.2

Table 3: continued

Field no.:	11	11	11	11	12	13	13
Sample no.:	SLP1314	SLP1302	SLP1308	SLP1320	BGS1306	MG1301	MG1303
Field name:	Sailipu	Sailipu	Sailipu	Sailipu	Zabuye	Maiga	Maiga
Latitude(N)	31°15'49.55"	31°26'7.37"	31°22'6.24"	31°17'38.64"	31°24'0.54"	30°48'16.08"	30°49'15.84"
Longitude(E)	82°47'25.73"	82°44'20.23"	82°40'10.76"	82°58'29.64"	84°18'15.6"	84°26'23.94"	84°26'7.62"
Age (Ma):	16.97	16.97	16.97	16.97	16.02	16.7	16.7
SiO ₂	47.04	56.80	55.67	56.80	55.54	53.75	46.44
TiO ₂	1.44	1.54	1.56	1.42	1.26	1.54	1.20
Al ₂ O ₃	10.50	12.03	13.70	12.61	13.36	11.15	10.09
TFe ₂ O ₃ *	8.26	5.93	6.94	6.47	6.14	6.39	8.31
MnO	0.13	0.09	0.15	0.10	0.10	0.10	0.13
MgO	13.74	8.20	7.55	8.82	5.58	11.03	15.75
CaO	8.88	4.83	5.30	4.94	7.86	5.21	10.47
Na ₂ O	0.91	2.13	2.71	1.95	2.89	1.61	0.90
K ₂ O	7.89	7.29	5.61	6.04	6.10	7.97	5.73
P ₂ O ₅	1.21	1.18	0.82	0.86	1.16	1.25	0.98
LOI	1.36	0.86	0.26	0.32	2.08	0.22	2.20
Mg-no.	0.80	0.77	0.73	0.77	0.69	0.81	0.82
La	325	104	101	110	79.1	194	169
Ce	680	289	244	278	172	503	290
Pr	87.2	46.5	33.5	39.2	23.6	75.3	33.3
Nd	347	227	149	184	104	336	134
Sm	65.0	47.6	26.7	33.0	19.8	64.2	24.5
Eu	12.4	6.55	3.77	4.94	3.76	9.25	6.28
Gd	44.7	24.1	15.3	18.0	12.9	35.1	18.9
Tb	4.41	2.38	1.61	1.83	1.47	3.38	2.13
Dy	15.7	8.36	6.00	7.06	6.16	11.5	8.95
Ho	2.06	1.29	0.97	1.11	0.98	1.65	1.48
Er	4.44	2.97	2.32	2.69	2.28	3.61	3.47
Tm	0.47	0.37	0.30	0.34	0.30	0.41	0.44
Yb	2.59	2.18	1.80	2.08	1.79	2.41	2.57
Lu	0.34	0.31	0.26	0.31	0.26	0.33	0.37
Sc	26.6	18.2	16.4	20.2	16.7	22.6	27.6
V	172	112	127	136	131	143	148
Cr	499	460	289	503	144	717	516
Co	43.7	31.9	33.4	34.0	26.7	43.1	49.1
Ni	272	306	220	308	120	493	452
Cu	92.6	40.5	23.6	52.5	45.4	74.5	73.5
Zn	105	92.6	104	97.8	98.7	106	96.2
Ga	21.3	24.0	23.2	23.3	20.6	27.9	15.7
Rb	675	928	503	528	401	775	298
Sr	3476	732	912	931	1765	1264	3389
Y	51.8	33.4	24.4	27.4	24.8	39.5	36.4
Zr	538	722	618	598	340	908	311
Nb	22.1	35.2	34.9	35.2	31.0	41.0	11.9
Ba	10115	2795	2501	2400	3514	5323	14076
Hf	12.3	20.6	17.9	17.4	9.33	25.7	6.77
Ta	1.11	2.09	2.36	2.00	1.82	2.22	0.59
Pb	176	78.9	89.2	57.0	76.8	106	286
Th	172	299	131	186	69.0	336	83.4
U	16.0	40.6	15.6	23.2	17.8	42.5	11.9

Major element oxide contents are normalized to 100 wt % on a volatile-free basis. Field no. refers to number of the volcanic field in Fig. 1. Age (Ma) is average value of the ages of the volcanic fields calculated from all of age data shown in Table 1. Mg-no. = Mg/(Mg + Fe²⁺), calculated assuming Fe₂O₃/(FeO + Fe₂O₃) = 0.20. *Total iron is given as Fe₂O₃. Samples (XB1307, SLP1314, SLP1302, SLP1308, SLP1320, BGS1306, MG1301, MG1303) have been analysed for olivine phenocryst compositions (Appendix A).

Table 4: Sr and Nd isotope compositions of the K-rich mafic magmatic rocks in south Tibet

Field no.	Sample no.	Field name	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr} \pm 2\sigma$	$(^{87}\text{Sr}/^{86}\text{Sr})_i$	$\epsilon\text{Sr}(i)$	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd} \pm 2\sigma$	$(^{143}\text{Nd}/^{144}\text{Nd})_i$	$\epsilon\text{Nd}(i)$
1	GZ-S05	Shiquanhe	3.9444	0.736547±13	0.735287	437.4	0.1117	0.511701±7	0.511685	-18.0
1	GZ-S06	Shiquanhe	4.3681	0.725218±11	0.723822	274.7	0.1510	0.511734±9	0.511712	-17.5
1	GZ-S12	Shiquanhe	4.0669	0.738916±10	0.737616	470.5	0.1439	0.511737±6	0.511716	-17.4
1	GZ-S15	Shiquanhe	2.8370	0.737682±12	0.736775	458.5	0.1621	0.511765±9	0.511741	-16.9
1	GZ-S27	Shiquanhe	2.6983	0.732057±16	0.731195	379.3	0.1405	0.511683±11	0.511662	-18.5
1	SQH-08	Shiquanhe	4.7444	0.735113±14	0.733597	413.4	0.1454	0.511716±8	0.511695	-17.8
1	SQH-15	Shiquanhe	7.3703	0.736793±15	0.734438	425.4	0.1514	0.511769±6	0.511747	-16.8
3	16YS	Bongba	5.7284	0.725416±13	0.723431	269.1	0.1389	0.511880±7	0.511858	-14.6
3	20YS	Bongba	9.5209	0.734271±12	0.730972	376.2	0.1426	0.511703±7	0.511680	-18.1
3	23YS	Bongba	7.0429	0.731082±10	0.728641	343.1	0.1346	0.511739±9	0.511718	-17.3
3	25YS	Bongba	3.9564	0.728623±11	0.727252	323.4	0.1230	0.511834±8	0.511814	-15.5
3	JS07	Bongba	4.5272	0.737944±18	0.736375	452.9	0.1392	0.511718±6	0.511696	-17.8
4	99-B06	Xungba	3.2795	0.726018±15	0.724965	290.9	0.1393	0.511945±7	0.511924	-13.4
4	99-B09	Xungba	2.7297	0.721629±14	0.720753	231.1	0.1420	0.511867±7	0.511846	-14.9
4	99-B23	Xungba	1.9282	0.718385±17	0.717766	188.7	0.1441	0.511881±8	0.511860	-14.6
4	99-C12	Xungba	2.6537	0.719638±16	0.718786	203.2	0.1407	0.511809±9	0.511788	-16.0
11	99-C15	Sailipu	3.1398	0.725834±13	0.725077	292.4	0.1440	0.511833±7	0.511817	-15.6
11	99-C28	Sailipu	2.7726	0.726121±13	0.725453	297.7	0.1601	0.511794±9	0.511776	-16.4
11	XB-06	Sailipu	3.7600	0.729294±12	0.728388	339.4	0.1587	0.511886±8	0.511868	-14.6
11	XB-24	Sailipu	2.7114	0.724286±15	0.723633	271.9	0.1613	0.511857±5	0.511839	-15.2
12	BG015	Zabuye	1.0989	0.718163±14	0.717913	190.7	0.1352	0.511816±6	0.511802	-15.9
12	BG026	Zabuye	1.2704	0.715573±14	0.715284	153.4	0.1284	0.511958±10	0.511945	-13.1
12	BG037	Zabuye	0.9952	0.714098±16	0.713872	133.3	0.1368	0.511857±11	0.511843	-15.1
12	BG201	Zabuye	0.7695	0.712554±13	0.712379	112.1	0.1297	0.511848±9	0.511834	-15.3
12	BG203	Zabuye	1.0489	0.712828±14	0.712589	115.1	0.1268	0.511932±6	0.511919	-13.6
13	MG-008	Maiga	1.1260	0.721105±17	0.720838	232.2	0.1313	0.511834±6	0.511820	-15.5
13	MG-012	Maiga	1.4993	0.720519±16	0.720163	222.6	0.1166	0.511821±5	0.511808	-15.8
13	MG-015	Maiga	1.6402	0.724673±18	0.724284	281.1	0.1160	0.511830±8	0.511817	-15.6
13	MG-016	Maiga	1.9372	0.725438±12	0.724979	291.0	0.1288	0.511782±8	0.511768	-16.6
16	GZF3	Garwa	0.7429	0.715228±11	0.715013	149.6	0.0824	0.511957±7	0.511946	-13.0
16	GZF9	Garwa	0.6130	0.714435±14	0.714258	138.9	0.1107	0.511974±9	0.511959	-12.7
17	G-029	Yaqian	0.5222	0.718317±15	0.718215	194.9	0.0868	0.511954±6	0.511946	-13.2
17	GH03	Yaqian	0.7649	0.717498±16	0.717349	182.6	0.1011	0.511919±7	0.511910	-13.9
19	GH05	Mibale	1.2176	0.718412±13	0.718148	194.0	0.1041	0.511971±9	0.511961	-12.8
19	GH09	Mibale	2.3061	0.716591±12	0.716090	164.8	0.0974	0.511994±7	0.511984	-12.4
21	ZFC004	Chazi	2.1798	0.722832±10	0.722476	255.4	0.1192	0.511914±7	0.511905	-14.0
21	ZFC008	Chazi	1.6339	0.725266±17	0.724999	291.2	0.1053	0.511867±9	0.511859	-14.9
21	ZFC009	Chazi	1.4504	0.729849±13	0.729612	356.7	0.1178	0.511808±8	0.511799	-16.1
22	GB036	Pabbai	1.4569	0.713384±14	0.713078	122.0	0.1126	0.511987±8	0.511976	-12.5

Field no. refers to number of volcanic field in Fig. 1. Chondritic uniform reservoir (CHUR) at the present day [$(^{87}\text{Rb}/^{86}\text{Sr})_{\text{CHUR}} = 0.0847$ (McCulloch & Black, 1984); $(^{87}\text{Sr}/^{86}\text{Sr})_{\text{CHUR}} = 0.7045$ (DePaolo, 1988); $(^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}} = 0.1967$ (Jacobsen & Wasserburg, 1980); $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} = 0.512638$ (Goldstein *et al.*, 1984)] were used for the calculations. $\lambda_{\text{Rb}} = 1.42 \times 10^{-11} \text{ year}^{-1}$ (Steiger & Jager, 1977); $\lambda_{\text{Sm}} = 6.54 \times 10^{-12} \text{ year}^{-1}$ (Lugmair & Marti, 1978). $\epsilon\text{Sr}(i)$ and $\epsilon\text{Nd}(i)$ were calculated using the average ages of the volcanic fields (Table 1).

Table 5: Pb isotope compositions of the K-rich mafic magmatic rocks in south Tibet

Field no.	Sample no.	Field name	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$	$^{238}\text{U}/^{204}\text{Pb}$	$^{235}\text{U}/^{204}\text{Pb}$	$^{232}\text{Th}/^{204}\text{Pb}$	$(^{206}\text{Pb}/^{204}\text{Pb})_i$	$(^{207}\text{Pb}/^{204}\text{Pb})_i$	$(^{208}\text{Pb}/^{204}\text{Pb})_i$
1	GZ-S05	Shiquanhe	18.713	15.785	39.818	16.002	0.118	145.225	18.657	15.782	39.656
1	GZ-S06	Shiquanhe	18.812	15.806	39.682	23.024	0.169	245.625	18.731	15.802	39.408
1	GZ-S12	Shiquanhe	18.806	15.841	39.794	18.282	0.134	101.354	18.742	15.838	39.681
1	GZ-S15	Shiquanhe	18.864	15.786	39.758	28.875	0.212	376.037	18.763	15.781	39.339
1	GZ-S27	Shiquanhe	18.877	15.806	39.815	41.389	0.304	323.449	18.732	15.799	39.455
1	SQH-08	Shiquanhe	19.005	15.796	39.801	24.723	0.182	281.129	18.919	15.792	39.488
1	SQH-15	Shiquanhe	18.685	15.837	39.613	17.014	0.125	143.751	18.626	15.834	39.453
3	16YS	Bongba	18.738	15.781	39.964	21.989	0.162	191.809	18.655	15.777	39.732
3	20YS	Bongba	18.815	15.862	39.801	14.548	0.107	148.835	18.760	15.859	39.621
3	23YS	Bongba	18.839	15.824	39.742	14.360	0.105	116.839	18.785	15.821	39.601
3	25YS	Bongba	18.652	15.836	39.829	25.735	0.189	188.222	18.554	15.831	39.602
3	JS07	Bongba	18.835	15.877	39.918	15.433	0.113	129.875	18.776	15.874	39.761
4	99-B06	Xungba	18.609	15.696	39.453	17.707	0.130	119.265	18.547	15.693	39.320
4	99-B09	Xungba	18.449	15.718	39.751	14.106	0.104	126.703	18.399	15.716	39.609
4	99-B23	Xungba	18.584	15.765	39.643	18.678	0.137	151.849	18.518	15.762	39.473
4	99-C12	Xungba	18.473	15.748	39.477	13.266	0.097	125.662	18.426	15.746	39.336
11	99-C15	Sailipu	18.721	15.751	39.772	31.470	0.231	190.487	18.638	15.747	39.612
11	99-C28	Sailipu	18.718	15.743	39.853	27.384	0.201	191.457	18.646	15.740	39.692
11	XB-06	Sailipu	18.705	15.732	39.819	32.878	0.242	240.678	18.618	15.728	39.617
11	XB-24	Sailipu	18.697	15.744	39.808	31.951	0.235	231.436	18.613	15.740	39.614
12	BG015	Zabuye	18.502	15.679	39.316	9.669	0.071	70.423	18.478	15.678	39.260
12	BG026	Zabuye	18.453	15.683	39.395	4.990	0.037	73.259	18.441	15.682	39.337
12	BG037	Zabuye	18.474	15.725	39.518	9.539	0.070	66.240	18.450	15.724	39.465
12	BG201	Zabuye	18.583	15.681	39.436	12.341	0.091	61.381	18.552	15.680	39.387
12	BG203	Zabuye	18.457	15.729	39.403	7.709	0.057	83.126	18.438	15.728	39.337
13	MG-008	Maiga	18.401	15.653	39.426	13.442	0.099	60.385	18.366	15.651	39.376
13	MG-012	Maiga	18.419	15.706	39.561	20.513	0.151	152.739	18.366	15.704	39.435
13	MG-015	Maiga	18.568	15.727	39.493	18.342	0.135	157.566	18.520	15.725	39.363
13	MG-016	Maiga	18.528	15.686	39.325	19.058	0.140	139.246	18.479	15.684	39.210
16	GZF3	Garwa	18.374	15.672	39.113	10.866	0.080	95.376	18.340	15.670	39.017
16	GZF9	Garwa	18.326	15.681	39.105	11.101	0.082	87.172	18.291	15.679	39.017
17	G-029	Yaqian	18.373	15.671	39.386	5.638	0.041	47.299	18.361	15.670	39.354
17	GH03	Yaqian	18.352	15.746	39.402	5.406	0.040	39.171	18.340	15.745	39.375
19	GH05	Mibale	18.533	15.745	39.514	12.018	0.088	92.859	18.504	15.744	39.444
19	GH09	Mibale	18.427	15.752	39.484	12.109	0.089	47.536	18.398	15.751	39.448
21	ZFC004	Chazi	18.659	15.762	39.273	16.354	0.120	61.705	18.630	15.761	39.238
21	ZFC008	Chazi	18.706	15.783	39.391	12.202	0.090	63.358	18.684	15.782	39.355
21	ZFC009	Chazi	18.543	15.826	39.285	14.956	0.110	90.573	18.516	15.825	39.233
22	GB036	Pabbai	18.739	15.726	39.194	14.924	0.110	82.407	18.705	15.724	39.134

Field no. refers to number of volcanic field in Fig. 1. $\lambda_{\text{U}238} = 0.155125 \times 10^{-9} \text{ year}^{-1}$, $\lambda_{\text{U}235} = 0.98485 \times 10^{-9} \text{ year}^{-1}$ and $\lambda_{\text{Th}232} = 0.049475 \times 10^{-9} \text{ year}^{-1}$ (Steiger & Jager, 1977). $(^{207}\text{Pb}/^{204}\text{Pb})_{\text{NHRL}} = 0.1084 \times (^{206}\text{Pb}/^{204}\text{Pb})_i + 13.491$ (Hart, 1984); $(^{208}\text{Pb}/^{204}\text{Pb})_{\text{NHRL}} = 1.209 \times (^{206}\text{Pb}/^{204}\text{Pb})_i + 15.627$ (Hart, 1984). Initial Pb isotope ratios were calculated using the average ages of the volcanic fields (Table 1).

Table 6 Mineral-melt partition coefficients for modeling calculations of partial melting of the Indian eclogite

Indian eclogite	D_{Cpx}	D_{Gt}	D_0	P
La	0.0281	0.00270	0.01921	0.02556
Dy	0.4588	2.65940	1.22901	0.67886
Yb	0.5983	10.0544	3.90794	1.54391

Abbreviations: Cpx, clinopyroxene; Gt, garnet.

Data source: [Liu *et al.* \(2014\)](#) and references therein.

$D_0 = 0.65D_{\text{Cpx}} + 0.35D_{\text{Gt}}$, represents the bulk partition coefficient during partial melting. The proportions of the mineral phases (Cpx=0.65, Gt=0.35) in the Indian eclogite are from [Rehman *et al.* \(2008\)](#).

$P = 0.9D_{\text{Cpx}} + 0.1D_{\text{Gt}}$, represents the bulk partition coefficient for the minerals entering the partial melt. The proportions of the mineral phases entering the melt (Cpx=0.90, Gt=0.10) during partial melting of the Indian eclogite are based on [Guo & Wilson \(2012\)](#).

Table 7 Result of the trace element modeling of partial melting of the Indian eclogite

Degree of melting	La (ppm)	Dy (ppm)	Yb (ppm)	La/Yb	Dy/Yb
0.01%	1165.20	9.45	1.30	897.06	7.28
0.05%	1146.69	9.45	1.30	882.81	7.28
0.1%	1130.32	9.45	1.30	870.29	7.28
0.5%	966.99	9.44	1.30	744.43	7.27
1%	853.25	9.42	1.30	657.43	7.26
2%	672.08	9.39	1.30	518.49	7.25
3%	556.75	9.36	1.29	430.38	7.24
5%	398.68	9.30	1.29	309.13	7.21
7%	307.65	9.23	1.28	239.54	7.19
8%	275.83	9.20	1.28	215.36	7.18
9%	249.08	9.17	1.28	195.00	7.18
10%	226.47	9.13	1.27	177.80	7.17
15%	153.11	8.98	1.26	121.21	7.11
20%	115.27	8.82	1.25	92.32	7.06
25%	92.31	8.65	1.23	74.93	7.02
30%	76.95	8.49	1.21	63.39	6.99
35%	65.96	8.32	1.19	55.23	6.96
40%	57.72	8.14	1.17	49.17	6.94
45%	51.30	7.97	1.15	44.54	6.92
50%	46.17	7.79	1.13	40.90	6.90

Compositions of the Indian eclogite for the modeling calculation during partial melting are as follows: La 22.50 ppm, Dy 11.62 ppm, Yb 5.08 ppm, based on Indian eclogite sample Ph380 from [Rehman et al. \(2008\)](#). Similar modeling results can be obtained using other samples of eclogite from [Rehman et al. \(2008\)](#).

The partial melting curve in Fig. 23 for Indian eclogite melting was calculated using a non-modal batch melting equation ([Wilson, 1989](#)).

Table 8 Modelling results for two-component mixing between an Indian eclogite-derived melt and a depleted MORB-source mantle (DMM) peridotite protolith

Proportion of eclogite melt	M1		M2		M3	
	La/Yb	Dy/Yb	La/Yb	Dy/Yb	La/Yb	Dy/Yb
5%	14.53	2.25	28.03	2.28	82.09	2.31
10%	25.81	2.95	50.06	3.00	147.08	3.04
15%	35.08	3.52	68.09	3.59	200.08	3.64
20%	42.84	4.00	83.12	4.08	244.12	4.14
25%	49.43	4.41	95.85	4.49	281.30	4.56
30%	55.10	4.76	106.76	4.85	313.11	4.92
35%	60.02	5.06	116.23	5.16	340.63	5.23
40%	64.34	5.33	124.51	5.43	364.68	5.51
45%	68.16	5.57	131.81	5.67	385.87	5.75
50%	71.56	5.78	138.31	5.88	404.69	5.96

M1, M2 and M3 represents modeling curves for two-component mixing (Fig. 23) between the DMM peridotite protolith and a 20%, 10% and 2% melt of the Indian eclogite, respectively.

Compositions of the depleted MORB-source mantle (DMM) peridotite protolith (La 0.192 ppm, Dy 0.505 ppm, Yb 0.365 ppm) are from [Workman & Hart \(2005\)](#).