Paleogene post-collisional lamprophyres in western Yunnan, western Yangtze Craton: Mantle source and tectonic implications

Yong-Jun Lu ^{a*}, T. Campbell McCuaig ^a, Zheng-Xiang Li ^b, Fred Jourdan ^c, Craig J.R. Hart ^d, Zeng-Qian Hou ^e, Suo-Han Tang ^e

^a Centre for Exploration Targeting, Australian Research Council Centre of Excellence for Core to Crust Fluid Systems (CCFS), School of Earth and Environment, University of Western Australia, Crawley, WA 6009, Australia

^b Australian Research Council Centre of Excellence for Core to Crust Fluid Systems, the Institute for Geoscience Research, Department of Applied Geology, Curtin University, GPO Box U1987, Perth, WA 6845, Australia

^c Western Australian Argon Isotope Facility, Department of Applied Geology and JdL-CMS, Curtin University, Perth, Australia

^d Mineral Deposit Research Unit, Department of Earth and Ocean Sciences, University of British Columbia, Vancouver, Canada V6T 1Z4

^e Institute of Geology, Chinese Academy of Geological Sciences, Beijing 100037, PR China

* Corresponding author: yongjun.lu@uwa.edu.au

A suite of lamprophyres, spatially associated with mafic lavas and potassic felsic intrusive rocks, was emplaced between 36.5 ± 0.2 and 33.7 ± 0.5 Ma (based on phlogopite $^{40}\text{Ar}/^{39}\text{Ar}$ dating) on the eastern side of the Ailao Shan-Red River shear zone in the western Yangtze Craton. These shoshonitic and ultrapotassic intrusive rocks post-date the ~60-55 Ma collisional event between the Indian and the Asian continents. They are characterized by: (1) enrichment in large-ion lithophile elements and light rare-earth elements with $(\text{La}/\text{Sm})_n = 3.15\text{--}7.15$; (2) strong positive Pb spikes; (3) depletion in high-field-strength elements (e.g. $\text{Nb}/\text{La} = 0.08\text{--}0.98$); (4) high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ($0.706 - 0.709$) with negative $\epsilon\text{Nd}(t)$ values of -10.5 to -0.9 ; (5) old Nd model ages of $1542\text{--}945$ Ma; and (6) radiogenic $(^{207}\text{Pb}/^{204}\text{Pb})_i$ of $15.57\text{--}15.70$ and $(^{208}\text{Pb}/^{204}\text{Pb})_i$ ($38.70\text{--}39.06$). These features suggest that the mantle source was metasomatized by Proterozoic subduction beneath the Yangtze Craton. The lamprophyres have similar trace element patterns, and Sr-Nd-Pb isotope compositions, as coeval mafic lava, indicating a common source of metasomatized veined continental lithospheric mantle (CLM). Lower degree partial melting of metasomatic veins likely generated the lamprophyres, whereas the coeval mafic lava was likely derived from melting of phlogopite harzburgite. The lamprophyres and mafic lava have similar Sr-Nd isotope systematics as CLM-derived Neoproterozoic mafic rocks and Late Permian Emeishan low-Ti basalt in the region, indicating that they share the same Proterozoic source. We envisage that mantle plumes thermally eroded the Proterozoic metasomatized CLM beneath the western part of the Yangtze Craton during $825\text{--}750$ Ma and $260\text{--}250$ Ma, although residual metasomatised domains remained before being tapped by delamination after the India-Asia continental collision during the Paleogene period.

Keywords: lamprophyres; western Yunnan; Himalayan orogen; lithospheric mantle; metasomatism; delamination

1. Introduction

Potassic igneous rocks are generally characterized by high concentrations of incompatible elements and, in the case of lamprophyres, high volatile contents (Foley et al., 1987; Rock, 1987). It is widely accepted that continental potassic mafic magmas are produced by small degrees of melting of metasomatized continental lithospheric mantle (CLM; Conticelli et al., 2009; Ersoy et al., 2010; Farmer, 2003; Foley, 1992; Pilet et al., 2008; Turner et al., 1996). Such melts, therefore, provide an important window into the composition of the CLM and its metasomatic history.

Eocene-Oligocene igneous rocks in the western Yunnan Province, southwestern China, form part of a semi-continuous potassic magmatic province in the eastern Indo-Asian collisional zone (Fig. 1; Chung et al., 1998, 2005; Zhang et al., 1987). The nature and origin of the mantle source, and the tectonic trigger for the partial melting of CLM beneath the western Yangtze Craton, remain controversial (Chung et al., 1998; Guo et al., 2005; Huang et al., 2010; Li et al., 2002; Wang et al., 2001; Xu et al., 2001a). For example, an enriched lithospheric mantle source beneath western Yangtze Craton has been interpreted to be exotic Tibetan lithosphere that extruded eastward under the Yangtze Craton during the Indo-Asian collision at ca. 55 Ma (Xu et al., 2001a). It has also been suggested that the western Yangtze lithospheric mantle was metasomatized locally by migrating melts from the seismic low-velocity zone after 250 Ma (Huang et al., 2010), or metasomatised by Proterozoic and Paleo-Tethyan oceanic subduction (Guo et al., 2005). The tectonic trigger for partial melting of the CLM beneath the western Yangtze Craton has been ascribed to convective thinning of the lithosphere (Chung et al., 1998), or eastward continental subduction along the Ailao Shan-Red River shear zone (Wang et al., 2001).

This contribution reports new geochronological, elemental, and Sr-Nd-Pb isotopic data for lamprophyres from several localities in western Yunnan with the aims of: (1) determining their magma sources, petrogenesis and the relationship with mafic lava and potassic felsic intrusions; (2) establishing the mantle metasomatic history and evolution of CLM beneath western Yangtze Craton;

and (3) resolving the reasons for the coeval emplacement of lamprophyres, mafic lava, and potassic felsic intrusions in a post-collisional tectonic setting.

2. Geological setting, sampling and petrography

2.1. Geological background

The collision of the Indian and Asian continents since the Early Cenozoic (ca. 60–55 Ma) resulted in the formation of the Himalayan orogenic belt, including the Tibetan Plateau (Chung et al., 2005). Lithospheric shortening was accommodated by subduction of continental lithosphere, and in part by southeastward extrusion tectonics facilitated by strike-slip motion along the NW- to WNW-trending Ailao Shan-Red River shear zone (Fig. 1; Chung et al., 1997; Leloup et al., 1995; Tapponnier et al., 1990; Yin and Harrison, 2000). The Tibetan Plateau can be subdivided into the Songpan-Ganzi, Qiangtang and Lhasa terranes, which were amalgamated before the Cretaceous along the Paleo-Tethys Jinshajiang and Meso-Tethys Nuijiang sutures (Fig. 1; Metcalfe, 2006; Ridd, 2013; Yin and Harrison, 2000). The Yangtze Craton lies to the southeast of the Tibetan Plateau (Fig. 1).

The western Yangtze Craton was the site of oceanic subduction during the late Precambrian (ca. 1050 to >860 Ma) followed by a likely long-lived mantle plume-related activity between 825 Ma and 750 Ma. The plume is expressed by mafic dyke swarms, continental flood basalts, and continental rifting over much of the craton accompanying the breakup of the supercontinent Rodinia (Ernst et al., 2008; Li et al., 2008; Wang et al., 2009). During the latest Neoproterozoic to Late Triassic (ca. 700–220 Ma), the Yangtze Craton constituted an isolated continental block in the Paleo-Pacific Ocean with its western side forming a passive continental margin (Li, 1998; Metcalfe, 2006; Pullen et al., 2008). During the Late Permian to Early Triassic (260–250 Ma), the Emeishan flood basalts erupted in the western part of the Yangtze Craton as the consequence of a mantle plume event (Chung and Jahn, 1995; Xu et al., 2001b; Zi et al 2010). Since the Late Triassic (220 – 0 Ma), the western Yangtze Craton has occupied an intracontinental position (Yang, 1998; Wang et al., 2000).

100 **2.2. Paleogene potassic igneous province**

101 Paleogene potassic magmatic suites are widely exposed in western Yunnan around the Ailao
102 Shan-Red River shear zone. These suites extend up to 50 km to the west into the Simao Block and
103 up to 270 km to the east into the Yangtze Craton (Fig. 1, Guo et al., 2005; Zeng et al., 2002). The
104 potassic rocks form small extrusive and intrusive bodies consisting of mafic to felsic units. Mafic
105 rocks are dominated by lamprophyre dykes (Guo et al., 2005; Li et al., 2002) with minor mafic lava
106 (Huang et al., 2010; Xu et al., 2001a), whereas felsic rocks are mainly porphyritic granites, some of
107 which host Cu (-Mo-Au) mineralization (Hou et al., 2006; Lu et al., 2013a). The lamprophyres
108 intruded as dykes and the mafic lava erupted from volcanic pipes (Huang et al., 2010). The
109 lamprophyres are distinguished from the mafic lava mainly by mineralogy. Phenocrysts in the
110 lamprophyres are dominated by phlogopite (Li et al., 2002). In contrast, the phenocrysts in the
111 mafic lava are dominated by olivine and clinopyroxene with rare phlogopite (Huang et al., 2010).
112 These potassic suites in western Yunnan belong to the Eocene-Oligocene potassic magmatic belt
113 emplaced between ca. 40 and 30 Ma that extends over 2000 km across the Qiangtang Terrane and
114 western Yangtze Craton, proximal to the Jinsha Suture (Fig. 1).

115 **2.3. Samples and petrography**

116 Twelve lamprophyre samples were collected from the Yao'an, Machangqing, Beiya,
117 Binchuan and Jianchuan areas, and are mostly associated with porphyry textured potassic felsic
118 intrusions that host Cu (-Mo-Au) deposits (Fig. 1). Previous work on the mafic lava and
119 lamprophyres focused on material distal from sites of mineral deposits (Guo et al., 2005; Huang et
120 al., 2010; Li et al., 2002; Xu et al., 2001a). The lamprophyre samples collected for this study are
121 porphyritic with phenocrysts dominated by phlogopite and clinopyroxene in a groundmass of
122 phlogopite-clinopyroxene-sanidine-Fe-Ti oxide (Table 1). The samples are fresh to moderately
123 altered with some samples having carbonate infillings (Table 1; Fig. 2). According to Rock (1987),
124 there are four types of lamprophyres: calc-alkaline lamprophyres, alkaline lamprophyres,
125 ultramafic lamprophyres, and lamproites. The Yunnan samples are classified as calc-alkaline

126 lamprophyres consisting of minettes. $^{40}\text{Ar}/^{39}\text{Ar}$ dating on fresh phlogopite was conducted for four
127 samples collected from the Yao'an porphyry Au deposit, the Beiya porphyry-skarn Au deposit, and
128 the Machangqing porphyry Cu-Mo deposit (Table 2; Fig. 3). Major and trace elements analyses
129 were determined for all samples (Table 3). Sr-Nd-Pb isotopic analyses were obtained from six
130 representative lamprophyre samples (Table 4).

131 3. Analytical methods

132 3.1. $^{40}\text{Ar}/^{39}\text{Ar}$ dating

133 Fresh phlogopite was separated from the sample of lamprophyre using a Frantz magnetic
134 separator, then carefully hand-picked under a binocular microscope. The selected minerals were
135 thoroughly rinsed with distilled water in an ultrasonic cleaner. Fish Canyon sanidine (FCs, $28.03 \pm$
136 0.08 Ma) was used as a flux monitor, for which good between-grain reproducibility has been
137 demonstrated (Jourdan and Renne, 2007). At the time of the measurement, we used an age of 28.03
138 ± 0.08 Ma for FCs, and we have recalculated the final age of all samples relative to the age of
139 $28.294 (\pm 0.13\%)$ based on the new values of the ^{40}K constants determined by Renne et al. (2011).
140 The samples were Cd-shielded to minimize undesirable nuclear interference reactions and irradiated
141 for 25 hours in the Hamilton McMaster University nuclear reactor (Canada). The mean J-values
142 computed from standard grains range from 0.0017193 ± 0.0000037 (0.20%) to $0.0017910 \pm$
143 0.0000039 (0.20%) determined as the average and standard deviation of J-values of the small wells
144 for each irradiation disc. The $^{40}\text{Ar}/^{39}\text{Ar}$ analyses were performed using a MAP 215-50 mass
145 spectrometer at the Western Australian Argon Isotope Facility at Curtin University. Samples were
146 step-heated using a 110 W Spectron Laser System, with a continuous Nd-YAG (IR; 1064 nm) laser
147 rastered over the sample during 1 minute to ensure a homogeneously distributed temperature. The
148 raw data were processed using the ArArCALC software (Koppers, 2002). The individual step ages
149 have been calculated using the decay constants recommended by Steiger and Jäger (1977) and
150 subsequently recalculated using the constant of Renne et al. (2011). Plateau ages are defined by
151 more than 70% of the total ^{39}Ar released. Mini-plateau ages need to include 50-70% of ^{39}Ar release

152 and are less reliable than plateau ages. Both need to have a P-value ≥ 0.05 (Jourdan and Renne,
153 2007). All source of uncertainties are included in the calculation.

154 **3.2. Major and trace elements**

155 The freshest rock slabs were crushed into small fragments before being further cleaned, then
156 ground into powder less than 74 μm in a tungsten carbide swing mill. Major and trace element
157 analyses were carried out at the National Research Center for Geoanalysis, Beijing. Major elements
158 were analyzed by wet chemistry and X-ray fluorescence spectrometry (XRF) methods with
159 analytical uncertainties < 5 relative % (Hou et al., 2012). Trace elements and rare earth elements
160 (REE) were analyzed using inductively coupled plasma-mass spectrometry (ICP-MS). Analytical
161 uncertainties are < 10 relative % for elements with abundances < 10 ppm, and < 5 relative % for
162 those > 10 ppm. The reference material used was GSR1 and the detailed analytical procedures were
163 described in He et al. (2002).

164 **3.3. Sr-Nd-Pb isotopic analyses**

165 The Sr isotope ratios were determined on a Finnigan MAT 262 thermal ionisation mass
166 spectrometer equipped with a multi-collector at the Isotope Geology Laboratory of the Institute of
167 Geology, Chinese Academy of Geological Sciences, Beijing. During the analytical period, several
168 measurements of NIST SRM 987 (formerly NBS 987) Sr reference standard gave $^{87}\text{Sr}/^{86}\text{Sr} =$
169 0.710232 ± 15 (2σ , $n=6$) (the recommended value is 0.710240). Mass fractionations were corrected
170 to $^{88}\text{Sr}/^{86}\text{Sr} = 8.37521$. Detailed sample preparation and analytical procedures for Sr isotopic
171 analysis follow those of Tang et al. (2007).

172 Nd isotope composition analyses were carried out on a Nu instrument MC-ICP-MS at the
173 Isotope Geology Laboratory of the Institute of Geology, Chinese Academy of Geological Sciences,
174 Beijing. The international JMC321 Nd reference solution yielded $^{143}\text{Nd}/^{144}\text{Nd} = 0.511124 \pm 10$ (2σ ,
175 $n=8$; the recommended value is 0.511125), and the international JNdi-1 standard gave $^{143}\text{Nd}/^{144}\text{Nd}$
176 $= 0.512117 \pm 11$ (2σ , $n=8$; the recommended value is 0.512115). Nd isotope ratios were internally
177 corrected for mass fractionation using a constant value of 0.7219 for $^{146}\text{Nd}/^{144}\text{Nd}$ by exponential

law. Detailed sample preparation and analytical procedures for Nd isotopic analysis follow those of He et al. (2007). Total procedural blanks were about 10^{-9} ~ 10^{-10} g Sr and 5×10^{-11} g Nd. The $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios were calculated using the whole-rock Rb, Sr, Sm and Nd concentrations obtained by ICPMS.

Pb was separated and purified using AG1-X8 resin (200–400 mesh) anionic ion-exchange columns with dilute HBr as eluant. Pb isotopic ratios of samples were measured by a Nu Instruments MC-ICPMS at the same lab as Nd isotopic analysis. Thallium was doped as an internal standard to correct the mass fractionation of Pb isotopes during analysis. During the period of analysis, repeated analyses of international standard NBS981 yielded ratios of $^{206}\text{Pb}/^{204}\text{Pb} = 16.939 \pm 0.013$ (the recommended value is 16.9356), $^{207}\text{Pb}/^{204}\text{Pb} = 15.497 \pm 0.011$ (the recommended value is 15.4891), and $^{208}\text{Pb}/^{204}\text{Pb} = 36.712 \pm 0.033$ (the recommended value is 36.7006; Todt et al., 1996; 2σ). Detailed sample preparation and analytical procedures for the Pb isotope measurements follow those of He et al. (2005).

Nd model ages (T_{DM}) were calculated assuming linear Nd isotopic growth of the depleted mantle reservoir from $\varepsilon_{\text{Nd}}(t) = 0$ at 4560 Ma to +10 at the present time (Goldstein et al., 1984).

$T_{\text{DMI}} = 1/\lambda \ln \{ 1 + [(^{143}\text{Nd}/^{144}\text{Nd})_{\text{s}} - 0.51315] / [(^{147}\text{Sm}/^{144}\text{Nd})_{\text{s}} - 0.2137] \}$ where s = sample.

4. Results

4.1. $^{40}\text{Ar}/^{39}\text{Ar}$ age data

Plateau ages were obtained from each of four samples selected for analysis with more than 80% cumulative ^{39}Ar released (Fig. 3; Table 2). All ages are calculated using the decay constants of Renne et al. (2011) and include all sources of uncertainties. Sample YA-3 from Yao'an yielded an age of 33.72 ± 0.53 Ma (2σ ; MSWD=1.01; $P = 0.43$), samples BY-6 and 64ZK12-284 from Beiya yielded plateau ages of 34.41 ± 0.22 (MSWD=1.17; $P = 0.31$) and 33.93 ± 0.24 Ma (MSWD=1.36; $P = 0.16$) respectively, and sample BXC-3 from Machangqing yielded an age of 36.52 ± 0.23 Ma (MSWD=1.30; $P = 0.26$) (Fig. 3). These ages are consistent with previous whole-rock or

203 mineral $^{40}\text{Ar}/^{39}\text{Ar}$ dating results of mafic rocks in western Yunnan, which range from 36.6 ± 0.2 to
204 32.9 ± 0.2 Ma (Chung et al., 1998; Guo et al., 2005; Huang et al., 2010). These lamprophyres form
205 part of the Eocene-early Oligocene (ca. 40-30 Ma) post-collisional potassic and ultrapotassic
206 magmatic pulse in the eastern part of the Tibetan plateau (Chung et al., 1998). The lamprophyres
207 are generally coeval with the potassic felsic intrusions in western Yunnan, which have zircon U-Pb
208 ages between 36.9 ± 0.3 and 32.5 ± 0.3 Ma (Lu et al., 2012).

209 4.2. Alteration effect

210 The major and trace element data of the lamprophyres are listed in Table 3. Given that most
211 of these samples were collected from, or close to porphyry Cu deposits characterized by widespread
212 hydrothermal alteration, it is essential to assess the effects of hydrothermal alteration on the
213 chemical composition of the lamprophyres. The geochemical data will be screened and any suspect
214 data should be discarded for petrogenetic discussion later.

215 A majority of the Yunnan lamprophyre samples have H_2O^+ contents of 1.12–3.42 wt%,
216 which are similar to the mean value of 2.7 ± 2.0 for H_2O^+ from world-wide calc-alkaline
217 lamprophyres (Rock, 1987; Table 3). Sample ZK500-142.5 has a significantly higher H_2O^+ content
218 of 5.62 wt%, consistent with its sericite alteration as observed in thin section (Figs. 2i and 4j; Table
219 3). The CO_2 contents of studied samples ranges from 0.18 to 7.37 wt%, which is more variable than
220 those of world-wide calc-alkaline lamprophyres with CO_2 of 1–5 wt% (Rock, 1987; Table 3). The
221 alteration effect on the major and trace element compositions of studied samples is assessed on
222 Figure 4 and 5, grouped by their different localities.

223 Among the three lamprophyre samples from Machangqing, one sample is carbonate and
224 sericite altered with CO_2 of 3.46 wt% and H_2O^+ of 2.8 wt% (Sample BXC-3; Fig. 2b; Table 3).
225 This altered sample has similar SiO_2 and MgO contents to those of fresh Machangqing samples, but
226 it has higher K_2O and lower TiO_2 , Al_2O_3 , $\text{Fe}_2\text{O}_3\text{T}$, CaO , Na_2O and P_2O_5 contents (Fig. 4). This
227 suggests that most of the major elements except SiO_2 and MgO may have been mobilized during
228 the alteration of Sample BXC-3. We note that the phlogopite crystals look mostly unaltered as

229 testified by the $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age obtained for selected crystals (Figs. 2b and 3d). In terms of
230 trace elements, the transition metals such as Ni and Cr of this sample remain similar to those of the
231 fresh samples (Fig. 5i; Table 3). Other large ion lithophile elements (LILE; Ba, Th, Sr) and rare
232 earth element (REE; La, Nd, Dy, Yb) are much lower in the altered sample compared with the fresh
233 ones (Fig. 5). Therefore, only SiO_2 , MgO and the transition metals such as Ni of Sample BXC-3
234 could be used for petrogenetic discussion.

235 The six lamprophyre samples from Beiya share similarities in most major element
236 concentrations despite differing degrees of alteration with variable CO_2 content of 2.19–7.37 wt%
237 (Fig. 4; Table 3). Altered sample ZK500-142.5 has abnormally higher TiO_2 content than all other
238 lamprophyre samples, and its TiO_2 content is discarded from further discussion (Fig. 4b). Samples
239 64ZK12-284 and 64ZK12-290 have the highest CO_2 contents of over 6 wt% as well as the highest
240 $\text{Fe}_2\text{O}_3\text{T}$ of over 14 wt%, suggesting the presence of secondary siderite which is supported by the
241 abundance of carbonate infillings in the two samples (Fig. 2k, l). The high $\text{Fe}_2\text{O}_3\text{T}$ contents of these
242 two samples are thus discarded for petrogenetic discussion. There is no obvious correlation between
243 CO_2 and SiO_2 , TiO_2 , Al_2O_3 , $\text{Fe}_2\text{O}_3\text{T}$ (except for samples 64ZK12-284 and 64ZK12-290), MgO ,
244 CaO , Na_2O , K_2O , and P_2O_5 for Beiya lamprophyre samples, indicating that carbonization in these
245 lamprophyres should be attributed to external CO_2 , but the other major element components were
246 from the lamprophyres. Samples 64ZK12-284 and 64ZK12-290 with the highest CO_2 have similar
247 heavy rare earth element (HREE) such as Yb and transition metals such as Ni to those of fresh
248 Beiya lamprophyres. However, the LILE and light rare earth element (LREE) of the two samples
249 are lower than fresh samples, suggesting possible remobilization of these elements during
250 carbonization (Fig. 5). The other four Beiya lamprophyre samples have similar trace element
251 concentrations (Fig. 5).

252 The two lamprophyre samples from Binchuan (BC-3) and Jianchuan (JC-13.3) have a low
253 loss on ignition (LOI) of 1.3 wt%, consistent with the petrographic observation that they are free of

254 alteration (Fig. 2e, f). Therefore, the major and trace elements of these two samples represent the
255 primary compositions.

256 Sericite-carbonate altered sample YA-3 from Yao'an has high CO₂ of 6.07 wt%, and
257 displays lower Al₂O₃, Na₂O, but higher K₂O, CaO and P₂O₅ content than other lamprophyre
258 samples, suggesting the possible mobility of these major elements. However, the SiO₂, TiO₂,
259 Fe₂O₃T, MgO contents of this sample are similar to the fresh samples. Sample YA-3 has
260 abnormally higher LILE and REE contents than other Yunnan lamprophyres, suggesting the
261 possible gain of these trace elements during carbonization. Its transition metals such as Ni, however,
262 are similar to the fresh samples from Binchuan and Jianchuan, indicating that transition metals
263 possibly remained unaffected during alteration.

264 4.3. Major and trace elements

265 The studied lamprophyres span a wide range of SiO₂ (43.5–54.8 wt%) and MgO (3.4–12.4
266 wt%) contents (Table 3). On a total alkalis-silica (TAS) diagram, all samples plot in the alkaline
267 series field, ranging from potassic trachybasalts to shoshonites, and latites, with two samples in the
268 phonotephrite field (Fig. 6a). These dykes have variably high K₂O (2.6–6.2 wt%; Table 3), but
269 relatively low Na₂O (1.5–4.4 wt%) contents, characteristic of potassic and ultrapotassic rocks
270 (Table 3; Fig. 6b). Lamprophyres of this study have alkalis content and K₂O/Na₂O ratios similar to
271 those of western Churchill minette in Canada (Cousen et al., 2001; Fig. 6).

272 Using MgO as an index of fractionation, there are weak negative covariations with SiO₂,
273 Al₂O₃, positive correlations with Cr and Ni, whereas other major element oxides scatter (Fig. 7).
274 The abundances of large ion lithophile elements (LILEs; such as Rb, Ba, Th) and high-field strength
275 elements (HFSEs; such as Nb, Zr) have no consistent trends when plotted against MgO (not shown).
276 Contemporaneous mafic lava with MgO < 10 wt% show similar overall trends with lamprophyres,
277 but those with MgO > 10 wt% show different trends due to olivine accumulation according to
278 Huang et al. (2010). Lamprophyres have higher Fe₂O₃T, TiO₂ and P₂O₅ contents than the mafic
279 lava (Fig. 7b, e, f).

280 Using Nd as an index of a moderately compatible element (Said et al., 2010), there are broad
281 covariant trends with the most incompatible elements Ba, Th, and La, but scatter with Nb, Zr, and
282 compatible elements such as Yb (Fig. 8). Both lamprophyre dykes and mafic lavas have higher
283 Ba/Nd, Th/Nd, La/Nd ratios, but lower Nb/Nd, Zr/Nd, and Yb/Nd ratios than for the primitive
284 mantle, suggesting derivation from a LILE-enriched and HFSE-depleted mantle source, similar to
285 the Paleoproterozoic western Churchill post-collisional minettes (Cousens et al., 2001; Fig. 8).

286 All lamprophyre samples exhibit coherent chondrite-normalized REE patterns with light
287 rare earth element (LREE) enrichment at La by ~100 to 900 times the chondrite value, with
288 fractionated REE where $(La/Sm)_n = \sim 3.15\text{--}7.15$, and $(Gd/Yb)_n = \sim 2.3\text{--}5.0$ (Fig. 9a, b). Mild
289 negative Eu anomalies are also present ($Eu/Eu^* = 0.76\text{--}0.89$, Table 3).

290 On primitive mantle-normalized multi-element diagrams, the large ion lithophile elements
291 (LILE), particularly Rb, Ba, and K, are significantly enriched relative to the high field strength
292 elements (HFSE) and LREE (Fig. 9c, d). There are significant Nb-Ta-Ti negative anomalies and
293 pronounced peaks at Pb (Fig. 9c, d).

294 4.4. Sr-Nd-Pb isotopes

295 The Paleogene lamprophyres of western Yunnan have relatively high initial $^{87}Sr/^{86}Sr$ ratios
296 of 0.706–0.709, but low initial $^{143}Nd/^{144}Nd$ ratios of 0.512053–0.512547, corresponding to $\epsilon Nd(t)$
297 values of –10.5 to –0.9 (Table 4). They plot as an array in the enriched quadrant of the Sr-Nd
298 isotope diagram (Fig. 10a). The Nd-isotope depleted mantle model ages range from 1542 to 945 Ma
299 (Table 4). Ratios of $(^{206}Pb/^{204}Pb)_i$ are variable and scatter with respect to initial $^{87}Sr/^{86}Sr$ ratios
300 (18.12–18.65; Fig. 10b, Table 4). Both $(^{207}Pb/^{204}Pb)_i$ (15.57–15.70) and $(^{208}Pb/^{204}Pb)_i$ (38.70–39.06)
301 are radiogenic, plotting as arrays well above the Northern Hemisphere Reference Line in the Pb
302 isotope diagrams (Fig. 10c, d; NHRL, Hart, 1984). The Sr-Nd-Pb isotopic compositions of the
303 lamprophyres in western Yunnan are within the range for the potassic felsic intrusions, in keeping
304 with a genetic relation between the potassic mafic and felsic magmas in the study region (Fig. 10).

305 5. Discussion

306 5.1. Magmatic evolution

307 5.1.1. Fractional crystallization

308 The lamprophyres sampled for this study have a large variation in Mg# [$\text{MgO}/(\text{MgO}+\text{FeO}_\text{T})$
309 in atomic ratio] ranging from 0.41 to 0.76 (Table 3), and there are broad correlations between MgO
310 and various major and trace elements that suggest fractional crystallization during magmatic
311 evolution (Fig. 7). The samples with MgO of ~10 wt % approximate primitive melt compositions as
312 they have Mg# of 0.70, which is generally assumed to be the value for melts in equilibrium with a
313 peridotite source, as determined experimentally (Table 3; Baker et al., 1995; Falloon et al., 1988).
314 The Ni and Cr contents of studied lamprophyres are positively correlated with MgO, and the SiO_2
315 and Al_2O_3 are negatively correlated with MgO, which is characteristic of olivine and clinopyroxene
316 fractional crystallization (Fig. 7a, c, g, h). A negative correlation between the Ni/Cr ratio and MgO
317 is evident for samples with MgO <10 wt%, indicating a clinopyroxene-dominated fractionation (Fig.
318 7i). Plagioclase was not an important fractionating phase given the negative correlation between
319 MgO and Al_2O_3 , and zero to minor negative Eu anomalies (Figs. 7c and 9a, b). The lack of obvious
320 correlations between P_2O_5 , TiO_2 and MgO indicates that accessory minerals such as apatite and Fe-
321 Ti oxides have not been significantly fractionated (Fig. 7e, f). The lamprophyre samples with
322 MgO > 10 wt% are most likely related to olivine accumulation as they have lower CaO and higher
323 Ni content and Ni/Cr ratios than the primitive melt, similar to the contemporaneous mafic lavas
324 having olivine accumulation (Fig. 7d, g, i; Huang et al., 2010). On broad terms, the fractionating
325 mineral assemblage of olivine-clinopyroxene can explain the variation in major element and
326 compatible trace element trends in the lamprophyres. However, the scatter for most elements vs
327 MgO suggests that these samples are probably not a comagmatic suite. Indeed, given that they were
328 collected from different areas and have variable Sr-Nd-Pb isotopic ratios (Figs. 1 and 10), they
329 should not be expected to be comagmatic.

330 5.1.2. Crustal contamination

331 The lamprophyres are mafic in terms of their SiO₂ and MgO contents, as well as having
332 enhanced concentrations of the mafic affiliated transitional metals Cr, Co, Ni, and V (Fig. 7; Table
333 3). However, the combination of the high contents of the incompatible elements Rb-Ba-Th-U-K-Pb-
334 LREE, Nb-Ta-Ti depletion, low Ce/Pb (1-8.75) and Nb/U (1.57-12.97) ratios, and the enriched Sr-
335 Nd isotopic signatures, suggest an enriched component was involved in the genesis of the Yunnan
336 lamprophyres. Accordingly, it is necessary to distinguish crustal contamination during magma
337 ascent from mantle source enrichment.

338 In order to assess the effect of crustal contamination, we have used the Energy-Constrained-
339 Assimilation and Fractional Crystallization (EC-AFC) model of Bohrson and Spera (2001) and
340 Spera and Bohrson (2001) to simulate interaction between the most isotopically primitive
341 lamprophyre (64ZK12-284) and possible crustal contaminants (Fig. 11). The crustal contaminants
342 include the Yangtze basement, proxied by the gneiss within the Ailao Shan-Red River shear zone
343 (sample YS-25; Zhang and Scharer, 1999), and average upper and lower crust (Bohrson and Spera
344 2001; input data and parameters are provided in Appendix). The simulations indicate that neither
345 upper crustal nor lower crustal contaminations could explain the trend of the trace element (such as
346 Sr and Nd) and isotopic data of the Yunnan lamprophyres and mafic lavas (Fig. 11). It is thus
347 unlikely that crustal contamination produced the enriched trace element and isotopic signatures of
348 these lamprophyres and mafic lavas.

349 The geochemical and isotopic signatures of the lamprophyre dykes of western Yunnan, and
350 coeval potassic mafic lava, were probably related to an enriched mantle source. This source would
351 have been variably enriched in LILE and LREE by earlier subduction-related metasomatism. As a
352 corollary, their enriched composition was not caused by crustal contamination during magma ascent.

353 **5.2. Mineralogy and composition of the mantle source region**

354 The presence of phlogopite harzburgite xenoliths in the lamprophyre in western Yunnan
355 may suggest these xenoliths are the source of the mafic magmas in western Yunnan (Lu and Qian,
356 1999). Below we use a series of trace element signature to prove this hypothesis.

357 Incompatible trace element ratios, such as Rb/Sr, Ba/Rb for the lamprophyre dykes in
358 western Yunnan remain broadly uniform with decreasing MgO concentrations (not shown). This
359 indicates that these trace element ratios were not affected by magma differentiation and therefore
360 can be used to fingerprint the source region (Furman and Graham, 1999).

361 The lamprophyres sampled from western Yunnan have high K₂O contents (mostly 3.4 to
362 6.7 %) over a range of MgO (3.5 to 11.6 %), and most samples have high K₂O/Na₂O ratios of
363 between 1.4 and 5 (Table 3). This suggests that the high K₂O values in the lamprophyres are a
364 primary feature. This is consistent with the presence of a potassium-rich mineral phase in the
365 mantle source region (which is probably related to phlogopite or amphibole). Positive correlations
366 of the La/K, La/Rb and La/Ba ratios with increasing La indicate that the bulk-rock partition
367 coefficients for K, Rb and Ba are higher than those for La (Fig. 12a, b, c; Feldstein and Lange,
368 1999). Similarly, the positive correlation between Th and Th/Rb indicates that Rb is more
369 compatible than Th (Fig. 12d). These features further suggest that phlogopite or amphibole are
370 present in the source region, which have a higher partition coefficient for K, Rb and Ba than for La
371 and Th (Foley et al., 1996; Green, 1994). The mafic lava lies at the low La end of the partial
372 melting trends defined by the lamprophyres, suggesting the mafic lava may have been derived from
373 larger degrees of partial melting than the lamprophyres (Fig. 12).

374 Both Rb and Ba are compatible elements and Sr is incompatible in phlogopite (Foley et al.,
375 1996; Green, 1994). The partition coefficient for Rb is higher than that for Ba in phlogopite,
376 whereas amphibole has the opposite partition coefficients for these elements (Foley et al., 1996;
377 Green, 1994). Consequently, melts of a phlogopite-bearing source are expected to have significantly

378 higher Rb/Sr and lower Ba/Rb ratios than those formed from amphibole-bearing sources (Furman
379 and Graham, 1999).

380 The western Yunnan lamprophyre dykes have high Rb/Sr (>0.1) and low Ba/Rb (<20) ratios,
381 strongly suggesting that they formed through melting of a phlogopite-bearing source (Fig. 13a).
382 Positive correlations in the plots of Sr vs Sr/Rb and Ba vs Ba/Rb indicate that the bulk-rock
383 partition coefficients for Rb is higher than those for Sr and Ba (Fig. 13 b, c), consistent with the
384 existence of phlogopite rather than amphibole in the source region. In addition, the
385 contemporaneous mafic lava also records evidence for a phlogopite-bearing source (Fig. 13a, b, c).
386 The presence of phlogopite in the mantle source for the lamprophyres from western Yunnan is
387 consistent with the presence of phlogopite-bearing peridotite xenoliths in these rocks (Lu and Qian,
388 1999).

389 The lamprophyres and mafic lava in western Yunnan fall in the depleted peridotite field,
390 which is defined by the study of experimental melts (Falloon et al., 1988) and is characterized by
391 low total Fe_2O_3 (<10 wt %) and TiO_2 (<2 wt %) (Fig. 14). These features indicate that these mafic
392 melts have been extracted from the depleted mantle source region in western Yunnan. The high
393 concentrations of LILE and LREE in the lamprophyres and mafic lava are most likely due to
394 subsequent subduction-related metasomatic enrichment of the depleted mantle source. This
395 suggestion is supported by the presence of phlogopite in the source region and the coincident Pb
396 spikes with Nb-Ta-Ti negative anomalies (Fig. 9). The likely source rocks are therefore
397 metasomatized phlogopite-bearing harzburgites, which agree well with the presence of phlogopite
398 harzburgite xenoliths in the lamprophyre in western Yunnan (Lu and Qian, 1999).

399 The lamprophyre dykes commonly have higher LILE and LREE contents, such as Ba and
400 La concentrations, than the mafic lava. This indicates that lamprophyres were derived from lower
401 degrees of partial melting than the mafic lava, possibly due to their higher volatile contents (Turner
402 et al., 1996; Figs. 8 and 12). The “vein-plus-wall-rock” melting model of Foley (1992) could
403 account for the compositional variations between the lamprophyres and contemporaneous mafic

lava in western Yunnan. The mafic lava involved a greater component of wall-rock depleted peridotite, supported by the abundance of olivine phenocrysts and higher Cr, Ni contents (Fig. 7g, h), as well as larger degree of partial melting (Fig. 12; Huang et al., 2010; Xu et al., 2001a). In contrast, the lamprophyres would have been mainly sourced from metasomatic phlogopite-bearing assemblages, reflected by their minor olivine and abundant clinopyroxene and phlogopite phenocrysts (Table 1).

5.3. Subduction-related enrichment of the mantle source

The lamprophyres and mafic lava in western Yunnan have high concentrations of LILEs, negative HFSE (e.g. Nb, Ta, Ti) anomalies, and enriched Sr-Nd isotopic characteristics, which appear to be lacking in the convecting asthenosphere (Xu et al., 2001a). Having ruled out crustal contamination as the cause through EC-AFC modeling (Fig. 11), these compositional and isotopic relationships suggest that the lamprophyres and mafic lava are derived from metasomatized lithospheric mantle.

As discussed above, there is phlogopite in the mantle source for the Yunnan lamprophyres, which could account for the LILE enrichment in the magmas because phlogopite is thought to be the major repository for these elements in the lithospheric mantle formed by fluid metasomatism (Ionov et al., 1997). Depletion in HFSEs relative to neighbouring elements, such as Nb-Ta and Ti anomalies, is considered indicative of subduction processes (Sun and Kerrich, 1995; Thirlwall et al., 1994 and references therein). Nb/U (1.6-13) and Ce/Pb (1.0-8.8) ratios for the Yunnan lamprophyres are significantly lower than those for mid-ocean ridge basalts and ocean island basalts, which have uniform Nb/U ratio of ~ 47 and Ce/Pb ratio of ~ 25 (Hofmann et al., 1986; Hofmann, 2003). This indicates a metasomatised mantle source of the lamprophyres, which is also supported by the strong positive Pb spike, a feature characteristic of metasomatised source (Table 3; Fig. 9c, d; Conticelli et al., 2009; Hofmann, 2003).

The above compositional characteristics of the western Yunnan lamprophyres do not support the hypothesis that metasomatism of the western Yangtze Craton CLM was caused by

small-volume melt derived from the seismic low velocity zone as proposed by [Huang et al. \(2010\)](#). If that were the case, such asthenosphere-derived melts would not have the HFSE depletion and positive Pb spike showed by the Yunnan lamprophyres and mafic lava ([Hofmann, 2003](#)). In addition, the lamprophyres do not have compositional features characteristic of alkaline basalts located at oceanic seamounts derived from shallow asthenosphere, such as the seismic low-velocity zone ([Geldmacher et al., 2006](#)). Alternatively, it was proposed that slab-derived melts and fluids metasomatized the mantle source region for the lamprophyres and mafic lava beneath western Yangtze Craton. However, the lamprophyres in western Yunnan are spatially remote from contemporaneous subduction zones, and the Proterozoic Nd-model age is difficult to reconcile with subduction during the Cenozoic ([Guo et al., 2005](#)). Accordingly, a plausible explanation is that the lithospheric mantle beneath western Yangtze Craton was metasomatized by an ‘ancient’ subduction event in keeping with the Proterozoic Nd model ages of 1542 to 945 Ma, possibly associated with the assembly of the Supercontinent Rodinia ([Table 4](#); [Li et al., 2008](#); [Zhao et al., 2004](#)).

5.4. Lithospheric mantle evolution beneath western Yangtze craton

Various models have been proposed to explain the origin of metasomatized lithospheric mantle beneath western Yangtze Craton. One model is that the metasomatism was younger than 250 Ma, assuming that the Late Permian Emeishan mantle plume event in western Yangtze Craton would have purged the lithospheric mantle of all its fusible constituents ([Huang et al., 2010](#); [Xu et al., 2001a](#)). Following this assumption, the enriched lithospheric mantle source beneath western Yangtze Craton has been interpreted to be an exotic Tibetan lithosphere extruding eastward under the craton during the Indo-Asian collision at ca. 55 Ma ([Xu et al., 2001a](#)), or the metasomatism was the result of migrating melts originating from the seismic low velocity zone after 250 Ma ([Huang et al., 2010](#)). It has also been proposed that Proterozoic and Paleo-Tethyan oceanic subduction caused the metasomatism beneath western Yangtze Craton ([Guo et al., 2005](#)).

We propose here that the metasomatism beneath western Yangtze Craton was related to Proterozoic subduction prior to 900 Ma. The metasomatically enriched CLM was thermally eroded

during the 825-750 Ma and 260-250 Ma plume events, but residual metasomatic domains were preserved and tapped by delamination of overthickened CLM formed during the Himalayan Orogeny in the Paleogene to form lamprophyres and mafic lava in western Yunnan. This interpretation is supported by several lines of evidence outlined below.

5.4.1. Double-sided Proterozoic subduction beneath the Yangtze Craton

The Xixiang arc in the north and the Shuangxiwu arc in the south of the Yangtze Craton suggest double-sided Proterozoic subduction before 900 Ma beneath the Yangtze Craton (Li et al., 2007, 2008; Wang et al., 2009; Fig. 16a). The 1050-900 Ma subduction on both sides of the Yangtze Craton had extensively metasomatized the lithospheric mantle of the Yangtze Craton (Wang et al., 2009).

5.4.2. Long-lived enrichments of the Yangtze lithospheric mantle

The Yunnan lamprophyres and mafic lava are characterized by high initial $^{87}\text{Sr}/^{86}\text{Sr}$ (>0.706) and low initial $^{143}\text{Nd}/^{144}\text{Nd}$ (<0.5125) ratios, and high $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ ratios (Fig. 10; Table 4). This suggests a long-lived (ca. 1500–900 Ma; Table 4) enrichments of Rb, light REE, and U in the subcontinental lithospheric mantle (Guo et al., 2005; Turner et al., 1996). It is possible that the old Nd model age is simply due to recent (e.g. Paleozoic or Mesozoic) subduction of old sediments into the mantle. However, the western Yangtze Craton was a passive continental margin during the latest Neoproterozoic to the Late Triassic (~ 700 – 220 Ma), characterized by deposition of carbonate and shale (Li, 1998; Li and Li, 2007; Metcalfe, 2006; Pullen et al., 2008). Since the Late Triassic (220 – 0 Ma), the western Yangtze Craton has experienced intracontinental deformation without oceanic plate subduction beneath its western margin where sandstone and conglomerate were deposited (Li, 1998; Yang, 1998; Wang et al., 2000). For that reason, Paleotethyan ocean subduction, such as that of the Lancangjiang Ocean, is unlikely to have metasomatized the western Yangtze Craton during the Devonian and the Early Carboniferous (Huang et al., 2010; Guo et al., 2005). Therefore, the enriched Sr-Nd-Pb isotopic compositions of

the Yunnan lamprophyres and mafic lava reflect the time-integrated effect of enrichment in the light REE, Pb, Rb, and U due to Proterozoic subduction.

5.4.3. Distinct Sr-Nd-Pb isotopes between Yunnan and Tibet mafic magmas

The Paleogene lamprophyres and contemporaneous mafic lavas in western Yunnan have different Sr-Nd-Pb isotopes from those in Tibet (Fig. 10). This is inconsistent with the hypothesis that the CLM beneath western Yangtze represents an exotic Tibetan lithosphere (Xu et al., 2001a). The western Yunnan lamprophyres have higher $\epsilon\text{Nd}(t)$ and lower initial $^{87}\text{Sr}/^{86}\text{Sr}$ values than potassic rocks in the Qiangtang Terrane of northern Tibet, albeit with some overlap (Fig. 10a). Their Sr-Nd-isotopes, however, are distinct from their mafic counterparts in the Lhasa Terrane of southern Tibet (Fig. 10a; Ding et al., 2003; Guo et al., 2006; Miller et al., 1999; Turner et al., 1996; Zhao et al., 2009). As a whole, the western Yunnan lamprophyres have partly overlapping Pb isotopic compositions of the potassic rocks in the Qiangtang Terrane of northern Tibet, but are distinctively lower in $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ values than potassic rocks of the Lhasa Terrane in southern Tibet (Fig. 10c, d Ding et al., 2003; Miller et al., 1999; Turner et al., 1996). Therefore, it is unlikely that an exotic Tibetan lithosphere was emplaced eastward under the Yangtze Craton during the Cenozoic (Xu et al., 2001a).

5.4.4. Similar Sr-Nd isotopic systematic of the mafic melts in western Yangtze Craton

The Paleogene lamprophyres and mafic lava in western Yunnan define the same linear trend on the $^{143}\text{Nd}/^{144}\text{Nd}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ plot in Figure 15a as that for the CLM-derived Neoproterozoic mafic rocks and Late Permian Emeishan low-Ti basalt in western Yangtze Craton. This suggests that these rocks were derived from the same CLM source (Fig. 15a; He et al., 2010; Song et al., 2008; Xiao et al., 2004; Zhou et al. 2006). Assuming the $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of Paleogene lamprophyres and mafic lava reflects that of the mantle sources, back calculation shows that the Neoproterozoic mafic rocks and Late Permian Emeishan low-Ti basalts are well within the evolution trends defined by the Paleogene lamprophyres and mafic lava (Fig. 15b). In summary the Sr-Nd isotopic systematic

506 supports the hypothesis that these mafic magmas in western Yangtze Craton were derived from the
507 same Proterozoic source.

508 **5.5. Tectonic trigger for the melting of CLM beneath the western Yangtze**

509 **Craton**

510 The tectonic setting for the generation of the Paleogene potassic felsic and mafic rocks in
511 western Yunnan remains controversial. It has been variously ascribed to eastward continental
512 subduction along strike-slip faults (Liang et al., 2007; Wang et al., 2001), continental rifting (Zhang
513 et al., 1987, 2000), or lithospheric thinning by convective removal of thickened lower lithosphere
514 (Chung et al., 1998, 2008). The potassic felsic and mafic rocks in western Yunnan are coeval and
515 were emplaced between 37 and 32 Ma, which predates the syn-kinematic leucogranite bodies (30-
516 23 Ma) hosted by the Ailao Shan-Red River shear zone (Lu et al., 2012). Therefore, any realistic
517 tectonic model proposed should explain the close proximity and coeval nature of potassic mafic and
518 felsic magmatism.

519 Wang et al. (2001) suggested that continental subduction during an early period of
520 transpression along the Ailao Shan-Red River shear zone may have induced fluid infiltration into
521 the mantle, which subsequently became the source for potassic magmatism in western Yunnan.
522 However, potassic rocks are not restricted to the Ailao Shan-Red River shear zone, and are
523 distributed through the interior of western Yangtze Craton (Guo et al., 2005; Lu et al., 2012). In
524 addition, potassic magmatism in western Yunnan was earlier than the earliest movement along the
525 Ailao Shan-Red River shear zone, which was initiated at ca. 32 Ma (Chung et al., 1997, 2008; Lu et
526 al., 2012; Searle et al., 2010). Also, the mantle source for the mafic rocks was metasomatized
527 during the Proterozoic rather than Cenozoic, which precludes continental subduction as the trigger
528 for partial melting of lithospheric mantle beneath western Yunnan.

529 It is known that the western Yangtze craton was in an intracontinental tectonic setting
530 during the Cenozoic and the India-Asia continental collision may have thickened the continental
531 lithosphere in western Yangtze (Lu et al., 2012; Yang, 1998; Wang et al., 2000). Given these

characteristics, one plausible tectonic trigger for the initiation of partial melting of the CLM with metasomatic domains is by delamination of an overthickened mantle lithosphere (Fig. 16d). Delamination would cause upwelling and decompressional partial melting of the asthenosphere, which invades the remaining lithospheric mantle containing residual metasomatised domains and causes melting of the lithosphere (Kay and Kay, 1993; Richards, 2009). The abundance of late Eocene to Oligocene rift basins in western Yunnan may be a consequence of “extensional collapse” after the convective lithospheric thinning (Chung et al., 1998). Asthenospheric mantle upwelling in western Yunnan is supported by geophysical studies, which show that there is uprising of a mantle diapir that is 300 km wide and derived from a depth of 450 km below western Yunnan (Lei et al., 2009; Liu et al., 2000). According to Xu et al. (2001a), potassic mafic lava in western Yunnan are spatially associated with early Tertiary picrite dykes generally thought to represent the products of mantle plumes (Xu et al., 2001a). This is consistent with upwelling of the asthenospheric mantle. It is therefore concluded that asthenospheric mantle uprising, probably caused by delamination focused along the pre-existing Jinsha Suture, most likely induced melting of the enriched lithospheric mantle in western Yunnan leading to the genesis of lamprophyres and mafic lava.

Lamprophyres in western Yunnan are similar in composition to Paleoproterozoic minettes (1830 Ma) from the western Churchill Province of NW Canada. The Churchill Province was the site of one of the largest ultrapotassic magmatic events in Earth history, and this ultrapotassic province post-dated the Paleoproterozoic (ca. 1830 Ma) Trans Hudson continent-continent orogeny (Figs. 6, 8, and 9; Cousens et al., 2001). Potassic magmatism in the province is interpreted to have been triggered by delamination following the Palaeoproterozoic orogeny, and the magmatism was derived from partial melting of residual continental lithospheric mantle metasomatised during ca. 2800 Ma (based on the Nd-model ages determined of Cousens et al., 2001). Interestingly, this metasomatized lithospheric mantle may have been thermally perturbed by younger mantle plumes at ca. 2700, 2400 and 2100 Ma, but remained partially intact before being tapped at 1800 Ma (Ernst and Bleeker, 2010; Cousens et al., 2001). This scenario is similar to that of the Yangtze Craton

558 where the metasomatized Proterozoic CLM was thermally eroded by mantle plumes at 825-750 Ma
559 and 260-250 Ma, but was not completely purged of metasomatised domains before being tapped
560 during the Paleogene (Fig. 16).

561 **6. Conclusions**

562 The following four main conclusions can be made from the observations discussed above:

- 563 (1) $^{40}\text{Ar}/^{39}\text{Ar}$ dating of lamprophyres from Beiya, Yao'an and Machangqing porphyry Cu (-Mo-
564 Au) deposits in western Yunnan yielded emplacement ages between 36.5 ± 0.2 and $33.7 \pm$
565 0.5 Ma, which are coeval with other lamprophyres and mafic lava in the region and postdate
566 the continental collision between India and Asia at ca. 60–55 Ma.
- 567 (2) Lamprophyres in western Yunnan are potassic to ultrapotassic in composition and are
568 characterized by enrichment in large-ion lithophile elements and light rare-earth elements,
569 strong positive Pb spikes, depletion in high-field-strength elements, high initial $^{87}\text{Sr}/^{86}\text{Sr}$
570 ($0.706 - 0.709$), negative $\epsilon\text{Nd}(t)$ values of -10 to -0.9 , highly radiogenic $(^{207}\text{Pb}/^{204}\text{Pb})_i$ of
571 $15.57 - 15.70$ and $(^{208}\text{Pb}/^{204}\text{Pb})_i$ of $38.70 - 39.06$, and old Nd model ages of 1542–945 Ma.
572 These features suggest that the mantle source was metasomatized by Proterozoic subduction
573 beneath the Yangtze Craton.
- 574 (3) The lamprophyres have similar trace element patterns and Sr-Nd-Pb isotopes with those of
575 coeval mafic lava, indicating a common source consisting of metasomatized veined
576 lithospheric mantle. The lamprophyres were generated by lower degree partial melting of
577 metasomatic veins in the continental lithospheric mantle, whereas the coeval mafic lava was
578 formed by larger degree melting of wallrock harzburgite.
- 579 (4) We tentatively propose that Proterozoic metasomatized lithospheric mantle beneath the
580 western Yangtze Craton was thermally eroded by mantle plumes at ca. 825-750 Ma and 260-
581 250 Ma, but some metasomatic domains remained before being tapped by delamination
582 during the Paleogene after the India-Asia continental collision.

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596 597 References

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Figure and Table captions

Fig. 1. Simplified geological map showing the distribution of Eocene-Oligocene potassic igneous rocks of western Yunnan (modified from [Zeng et al., 2002](#); [Huang et al., 2010](#)). The inset shows the tectonic framework of the Tibetan Plateau and surrounding areas, the geographic boundary of Yunnan Province (dashed lines) and the red fields denote the Eocene-Oligocene potassic magmatic rocks along Jinsha Suture (modified from [Yang, 1998](#); [Xu et al., 2001a](#); [Chung et al., 2005](#)). Please note there is no ophiolite outcrop of the Jinsha suture in the study area but the Ailao Shan-Red River shear zones are thought to mark the western boundary of the Yangtze craton (e.g. [Guo et al., 2005](#)). Samples collected in this study are shown within rectangles. ASRR, Ailao Shan-Red River shear zone.

855 Fig. 2. Photomicrographs of studied lamprophyres in western Yunnan. (a) Sample YA-3 showing
856 carbonate infilling surrounded by slightly altered phlogopite phenocrysts in the sericite altered
857 groundmass. (b) Sample BXC-3 showing carbonate replacing phenocrysts. Groundmass is sericite
858 and carbonate altered. Some relatively fresh phlogopite phenocrysts are preserved. (c) Sample
859 BXC-10 showing fresh phlogopite and clinopyroxene phenocrysts in the fresh groundmass
860 dominated by feldspar. (d) Sample BXC-16 with fresh phlogopite and clinopyroxene phenocrysts in
861 fresh groundmass dominated by feldspar. (e) Sample BC-3 with fresh large K-feldspar and
862 phlogopite phenocrysts in a fresh groundmass of clinopyroxene, phlogopite and feldspar. (f) Sample
863 JC-13.3 with fresh phlogopite and clinopyroxene phenocrysts in a fresh groundmass of feldspar. (g)
864 Sample BY-6 of fresh phlogopite and clinopyroxene phenocrysts in the fresh groundmass
865 dominated by feldspar. (h) Sample BY-L with some carbonate infilling in the sericite altered
866 groundmass. Phlogopite phenocrysts are relatively fresh. (i) Sample ZK500-142.5 of weakly altered
867 phlogopite and clinopyroxene phenocrysts in a sericite altered groundmass rich in K-feldspar. (j)
868 Sample BY-L (N-S) of fresh phlogopite and clinopyroxene phenocrysts in a fresh groundmass
869 dominated by K-feldspar. (k) Sample 64ZK12-284 of carbonate infilling surround by fresh
870 phenocrysts of phlogopite and clinopyroxene and the groundmass of feldspar. (l) Sample 64ZK12-
871 290 of carbonate developed as infilling or replacing phenocrysts. Weakly altered phlogopite and
872 clinopyroxene phenocrysts in the sericite altered groundmass. (a) is under cross-polarized light
873 whereas (b) to (i) are all under plane polarized light.

874

875 Fig. 3. $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age and related K/Ca ratio spectra of the phlogopite separates from
876 lamprophyres in western Yunnan vs. the cumulative percentage of ^{39}Ar released.

877

878 Fig. 4. Plot of CO_2 vs major elements for lamprophyres. The data with sample labels, e.g. YA-3, are
879 discarded for petrogenetic discussions due to possible mobilization during alteration.

880

881 Fig. 5. Plot of CO₂ vs trace elements for lamprophyres. The data with sample labels, e.g. YA-3, are
882 discarded for petrogenetic discussions due to possible mobilization during alteration.

883

884 Fig. 6. (a) TAS diagram ([Le Maitre, 2002](#)), and (b) K₂O versus Na₂O diagram for Paleogene
885 lamprophyres and potassic mafic lava in western Yunnan. Paleogene mafic lava, western Yunnan,
886 on the east side of the Ailao Shan-Red River shear zone are from [Xu et al. \(2001a\)](#) and [Huang et al.](#)
887 [\(2010\)](#). The minettes from the western Churchill Province in NW Canada, which is the site of one
888 of the largest ultrapotassic magmatic events in Earth history at ca. 1830 Ma, are plotted for
889 comparison ([Cousens et al., 2001](#)). Major element oxides are recalculated on a volatile-free basis.
890 U1, tephrite; U2, phonotephrite; U3, tephriphonolite; S1, potassic trachybasalt; S2, shoshonite; S3,
891 latite; T, trachyte; B, basalt; BA, basaltic andesite; A, andesite; D, dacite.

892

893 Fig. 7. MgO vs major- and selected trace-element diagrams for the Paleogene lamprophyres and
894 mafic lavas from western Yunnan, showing possible fractionating phases during magma evolution.
895 The dotted vertical lines represent primitive melt with MgO = 10%. The olivine (Ol) and
896 clinopyroxene (Cpx) accumulation and fractionation vectors on CaO vs MgO panel are after [Huang](#)
897 [et al. \(2010\)](#).

898

899 Fig. 8. Binary variations of selected trace elements versus Nd for the lamprophyres and mafic lavas
900 in western Yunnan. The Western Churchill Province minettes are plotted for comparison ([Cousens](#)
901 [et al., 2001](#)). Dotted lines are primitive mantle ratios from [Sun and McDonough \(1989\)](#). Data source
902 is same as in Fig. 6.

903

904 Fig. 9. Chondrite-normalized REE patterns (a, b) and primitive mantle-normalized diagrams (c, d)
905 for Paleogene lamprophyres in western Yunnan. Gray bands represent the field of literature data for
906 contemporaneous potassic mafic magmatism ([Xu et al., 2001a](#); [Li et al., 2002](#); [Guo et al., 2005](#);

907 [Huang et al., 2010](#)). They show similar REE and trace element patterns to the western Churchill
908 minettes (averaged from [Cousens et al., 2001](#)). Chondrite and primitive mantle normalizing values
909 from [Sun and McDonough \(1989\)](#).

910

911 Fig. 10. (a) $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs $\epsilon_{\text{Nd}}(t)$, (b) $(^{206}\text{Pb}/^{204}\text{Pb})_i$ vs $(^{87}\text{Sr}/^{86}\text{Sr})_i$, (c) $(^{206}\text{Pb}/^{204}\text{Pb})_i$ vs $(^{207}\text{Pb}/^{204}\text{Pb})_i$
912 and (d) $(^{206}\text{Pb}/^{204}\text{Pb})_i$ vs $(^{208}\text{Pb}/^{204}\text{Pb})_i$ plots for Paleogene lamprophyres and mafic lavas in western
913 Yunnan. The field for Yunnan potassic felsic intrusions is after [Lu et al. \(2013a, 2013b\)](#). Mafic
914 lavas in western Yunnan are from [Xu et al. \(2001a\)](#) and [Huang et al. \(2010\)](#). The data for post-
915 collisional potassic volcanic rocks in Tibet (South Tibet and North Tibet) are from [Arnaud et al.](#)
916 [\(1992\)](#), [Turner et al. \(1996\)](#), [Miller et al. \(1999\)](#), [Ding et al. \(2003\)](#), [Guo et al. \(2006\)](#), and [Zhao et](#)
917 [al. \(2009\)](#). Northern Hemisphere Reference Line (NHRL) is from [Hart \(1984\)](#); mantle source
918 reservoirs BSE, MORB, DMM, EM I and EM II are from [Zindler and Hart \(1986\)](#); Indian MORB
919 and Pacific MORB after [Guo et al. \(2005\)](#).

920

921 Fig. 11. (a) Sr vs $(^{87}\text{Sr}/^{86}\text{Sr})_i$, (b) Nd vs $(^{143}\text{Nd}/^{144}\text{Nd})_i$ plots for Paleogene lamprophyres and
922 potassic mafic lavas in western Yunnan. Also shown are crustal contamination vectors calculated
923 using EC-AFC thermodynamic model ([Bohrson and Spera, 2001](#); [Spera and Bohrson, 2001](#); input
924 parameters are listed in Appendix): the most isotopically primitive lamprophyre 64ZK12-284 is
925 used to model the composition of magma entering the crust; possible crustal contaminants are
926 represented by gneiss within the Ailao Shan-Red River shear zone (grey curve, sample YS-25,
927 [Zhang and Scharer, 1999](#)), and by average upper (red curve) and lower crust (blue curve) from
928 [Bohrson and Spera \(2001\)](#). The tick on curves indicate percentage of assimilated contaminant. Data
929 source same as in Fig. 6.

930

931 Fig. 12. (a) $1000 \cdot \text{La}/\text{K}$ vs La, (b) La/Rb vs La, (c) $100 \cdot \text{La}/\text{Ba}$ vs La and (d) Th/Rb vs Th, for
932 Paleogene lamprophyres and potassic mafic lava from western Yunnan. The regression lines are

constructed using the lamprophyre data from this study and the R^2 values are shown beside the lines. The vectors of increasing partial melting degrees are also shown with arrows.

935

Fig. 13. Plots of (a) Rb/Sr vs Ba/Rb, (b) Sr/Rb vs Sr, and (c) Ba/Rb vs Ba for Paleogene lamprophyres and potassic mafic lava in western Yunnan. In (a) PM represents the primitive mantle values of [Sun and McDonough \(1989\)](#). The field of continental lithospheric mantle (CLM; [Furman and Graham, 1999](#)) is reported for comparison along with qualitative arrows evidencing the expected compositional variation due to amphibole vs. phlogopite mantle metasomatism, respectively.

942

Fig. 14. TiO_2 vs total Fe_2O_3 for Paleogene lamprophyres and mafic lavas in western Yunnan in comparison with fields of the peridotitic melts reported by [Falloo et al. \(1988\)](#).

945

Fig. 15. Plots of (a) $^{143}Nd/^{144}Nd$ vs $^{87}Sr/^{86}Sr$ and (b) $\epsilon Nd(t)$ vs Age for continental lithospheric mantle (CLM)-derived mafic rocks in western Yangtze Craton. The 750 Ma, 1400 Ma and 1800 Ma evolution curves in (b) are constructed using $^{147}Sm/^{144}Nd$ ratios of Eocene mafic rocks assuming the ratios reflects mantle sources. (a) Shows Eocene mafic rocks have nearly the same Sr-Nd isotopes as the Emeishan CLM-derived low-Ti basalt and Neoproterozoic mafic intrusions, indicating that they are derived from the same CLM source. (b) Shows that the Emeishan low-Ti basalt and Neoproterozoic mafic rocks are well within the evolution trends defined by the Eocene mafic rocks, also suggesting that they were derivation from the same CLM source. All of the Sr-Nd isotopic data have been recalculated at 35 Ma for comparison. Data source: 900 Ma Yanbian basalt ([Li et al., 2006](#)); 825 Ma Lower Group Bikou basalt ([Wang et al., 2008](#)); 810 Ma Yanbian mafic intrusion ([Zhou et al. 2006](#)); 817 Ma Tiechuanshan basalt ([Ling et al., 2003](#)); 250 Ma Emeishan low-Ti basalt ([Xu et al., 2001b](#); [Xiao et al., 2004](#); [Song et al., 2008](#)); Eocene lamprophyre and mafic lava ([Xu et al., 2001a](#); [Li et al., 2002](#); [Guo et al., 2005](#); [Huang et al., 2010](#); this study).

960 Fig.16. Schematic illustration of the petrogenesis of post-collisional lamprophyre and mafic lava in
 961 western Yunnan and the evolution of mantle lithosphere beneath western Yangtze Craton (modified
 962 from [Wang et al., 2009](#); [Huang et al., 2010](#)). (a) Oceanic subduction beneath the Yangtze Craton
 963 before 900 Ma produced metasomatic veins within the continental lithospheric mantle; (b) In the
 964 period between 825 and 750 Ma, a possible mantle plume raised and migrated laterally to the
 965 margin of Yangtze Craton, where rifting took place and continental flood basalt and mafic dyke
 966 swarms formed. The metasomatic veins within continental lithospheric mantle were located far
 967 away from the rifted margin and survived the purging of mantle plume; (c) At Late Permian (260-
 968 250 Ma), the Emeishan mantle plume raised beneath western Yangtze Craton. This resulted in the
 969 production of high-Ti and low-Ti basal from the plume and continental lithospheric mantle,
 970 respectively ([Xiao et al., 2004](#)). However, the metasomatized domain within CLM was not
 971 completely purged; (d) In the Paleogene, the India-Asia continental collision thickened the
 972 lithospheric mantle in western Yangtze Craton and this overthickened CLM eventually delaminated
 973 at 37-32 Ma. The upwelling asthenosphere after delamination heated the CLM, causing partial
 974 melting of metasomatic veins and peridotite, producing lamprophyres and mafic lava, respectively.
 975 (a) and (b) are N-S cross sections while (c) and (d) are W-E section of Yangtze Craton. The
 976 sketches are not to scale. CFBs = continental flood basalt.

Fig.1

Fig. 2

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Fig.15

(a) 1100-900 Ma

(b) 825-750Ma

(c) 260-250Ma

(d) 37-32Ma

Fig.16

Table 1 Sample localities and brief description of Paleogene lamprophyres in western Yunnan.

Sample	Locality	Longitude (°E)	Latitude (°N)	Rock type	Mg#	Phenocrysts	Groundmass	Alteration description
YA-3	Yao'an	101.28913	25.38526	Phonotephrite	0.62	Phl (5) +Cpx (1)	Phl + Ap + Fe-Ti	Phenocrysts are weakly altered. Groundmass is intensely altered to sericite and carbonate. Some carbonate infilling also developed.
BXC-3	Machangqing	100.46304	25.54811	Shoshonite	0.76	Phl (5) +Ol (3)	Phl + San + Fe-Ti	Large phlogopite phenocrysts are fresh to weakly altered. Groundmass is intensely altered to sericite and carbonate. Some large phenocrysts are replaced by carbonate.
BXC-10	Machangqing	100.42578	25.52780	Shoshonite	0.70	Phl (7) + Cpx (3)	Phl + Cpx + San	Fresh phenocrysts and groundmass
BXC-16	Machangqing	100.42578	25.52780	Shoshonite	0.74	Phl (5) + Cpx (2)	Phl + Cpx + San	Fresh phenocrysts and groundmass
BY-6	Beiya	100.19207	26.16601	Latite	0.67	Phl (10) + Cpx (3) +Opx (1)	Phl + Cpx + Ap + Fe-Ti + Glass	Phenocrysts and groundmass are fresh. There are some carbonate infilling.
BY-L	Beiya	100.19207	26.16601	Latite	0.61	Phl (7) + Cpx (1)	Phl + Cpx + Fe-Ti + Glass	Phenocrysts are fresh. Groundmass is altered to sericite. There are some carbonate infilling.
ZK500-142.5	Beiya	Borehole ZK500, 142.5m		Potassic trachybasalt	0.70	Phl (7) + Cpx (3) + San (1)	Phl + Cpx + San + Fe-Ti	Some large phenocrysts are replaced by carbonate. The groundmass is sericite altered. There are some carbonate infillings.
BY-L(N-S)	Beiya	100.19154	26.16558	Phonotephrite	0.63	Phl (10) + Cpx (5) + San (5)	Phl + Cpx + San + Fe-Ti	Fresh phenocrysts and groundmass
64ZK12-284	Beiya	Borehole 64ZK12, 284m		Shoshonite	0.41	Phl (3) + Cpx (3) + Opx (1)	Phl + Cpx + San + Fe-Ti	Phenocrysts and groundmass are fresh, but there are abundant carbonate infillings.
64ZK12-290	Beiya	Borehole 64ZK12, 290m		Phonotephrite	0.45	Phl (1) + Ol (1)	Phl + Cpx + San + Fe-Ti	Some large phenocrysts are replaced by carbonate. The groundmass is sericite altered. There are abundant carbonate infillings.
BC-3	Binchuan	100.65682	25.77359	Shoshonite	0.56	Phl (2) + Amp (1) + Pl (1) + Cpx (1)	Phl + Cpx + San + Fe-Ti	Fresh phenocrysts and groundmass

JC-13.3	Jianchuan	99.90211	26.59196	Latite	0.45	Ol (3) + Phl (1) + Cpx (1)	Phl + San + Ap + Fe-Ti	Fresh phenocrysts and groundmass
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Amp, amphibole; Ap, apatite; Cpx, clinopyroxene; Opx, orthopyroxene; Ol, olivine; Phl, phlogopite; Pl, plagioclase; San, sanidine; Fe-Ti, Fe-Ti oxides.

Mg# =MgO/(MgO+FeOT) in atomic ratio. Proportions in percent of phenocrysts shown in parentheses (volume %)

Four samples in bold (e.g. **YA-3**) are dated by Ar-Ar method in this study

Table 2 $^{40}\text{Ar}/^{39}\text{Ar}$ dating of phlogopite in Paleogene lamprophyres from western Yunnan.

Laser intensity (%)	³⁶ Ar ±σ (V)		³⁷ Ar ±σ (V)		³⁸ Ar ±σ (V)		³⁹ Ar ±σ (V)		⁴⁰ Ar ±σ (V)		⁴⁰ Ar* (%)	⁴⁰ Ar*/ ³⁹ Ar _K ±σ		Age ±σ (Ma)		
YA-3	J=0.0017600 ± 0.0000130 (0.74%)						D=1.0012 ± 0.0036									
61.0	0.000213	0.000010	0.000475	0.000037	0.000057	0.000005	0.00121	0.00002	0.059367	0.000156	21.60	9.06	± 3.82	28.53	± 23.88	
62.5	0.000264	0.000013	0.001301	0.000045	0.000134	0.000014	0.00490	0.00007	0.101155	0.000183	41.46	7.75	± 1.05	24.45	± 6.57	
64.0	0.000467	0.000017	0.004062	0.000062	0.000242	0.000012	0.00812	0.00008	0.183786	0.000239	35.89	7.71	± 0.78	24.32	± 4.91	
65.5	0.000432	0.000014	0.005212	0.000072	0.000219	0.000009	0.00803	0.00006	0.191351	0.000332	44.45	10.08	± 0.69	31.73	± 4.28	
67.0	0.000882	0.000019	0.007615	0.000060	0.000582	0.000016	0.02781	0.00011	0.556748	0.000417	57.33	11.33	± 0.25	35.61	± 1.54	
67.7	0.001260	0.000021	0.000596	0.000025	0.000763	0.000013	0.03719	0.00018	0.770878	0.000595	54.22	11.12	± 0.23	34.98	± 1.41	
68.2	0.000105	0.000007	0.000305	0.000030	0.000053	0.000005	0.00226	0.00003	0.039025	0.000052	78.23	10.75	± 1.32	33.80	± 8.25	
68.7	0.000213	0.000010	0.000441	0.000022	0.000164	0.000012	0.00867	0.00004	0.139368	0.000234	71.07	10.78	± 0.42	33.90	± 2.60	
69.2	0.000933	0.000017	0.000446	0.000026	0.000624	0.000016	0.03371	0.00006	0.634007	0.000583	60.12	11.17	± 0.20	35.12	± 1.23	
69.7	0.000813	0.000019	0.000465	0.000021	0.000758	0.000018	0.04528	0.00009	0.705436	0.000380	69.28	10.68	± 0.15	33.59	± 0.96	
70.2	0.000272	0.000008	0.000429	0.000022	0.000420	0.000019	0.02781	0.00011	0.363136	0.000389	84.49	10.80	± 0.13	33.96	± 0.82	
70.7	0.000770	0.000024	0.000527	0.000031	0.001051	0.000021	0.06909	0.00014	0.951289	0.000775	78.64	10.74	± 0.12	33.79	± 0.78	
71.1	0.000399	0.000011	0.000517	0.000026	0.001105	0.000034	0.07949	0.00019	0.942324	0.000894	90.11	10.60	± 0.07	33.34	± 0.43	
71.5	0.000501	0.000013	0.000547	0.000025	0.001929	0.000028	0.13167	0.00029	1.539181	0.001308	91.99	10.70	± 0.06	33.67	± 0.36	
72.0	0.000702	0.000006	0.000551	0.000019	0.003203	0.000029	0.23294	0.00061	2.666029	0.003305	93.14	10.63	± 0.05	33.45	± 0.32	
72.5	0.000481	0.000015	0.000560	0.000021	0.002766	0.000033	0.20768	0.00029	2.324121	0.002103	94.90	10.59	± 0.05	33.31	± 0.31	
73.0	0.000191	0.000012	0.000435	0.000024	0.000597	0.000019	0.04520	0.00009	0.518246	0.000589	93.47	10.56	± 0.10	33.22	± 0.62	
79.5	0.000135	0.000008	0.000476	0.000016	0.000507	0.000017	0.03573	0.00004	0.398620	0.000272	95.58	10.46	± 0.10	32.90	± 0.60	
83.0	0.000177	0.000009	0.000357	0.000011	0.000767	0.000022	0.05883	0.00017	0.658124	0.000489	95.63	10.58	± 0.07	33.28	± 0.44	
BY-6	J=0.0017700 ± 0.0000050 (0.28%)						D=1.0012 ± 0.0036									
61.0	0.000167	0.000011	0.000463	0.000021	0.000050	0.000009	0.001261	0.000018	0.041135	0.000192	23.46	6.28	± 2.88	19.93	± 18.16	
62.5	0.000230	0.000006	0.000988	0.000028	0.000106	0.000005	0.003066	0.000018	0.076554	0.000173	36.72	8.29	± 0.77	26.27	± 4.86	
64.0	0.000279	0.000014	0.002605	0.000039	0.000120	0.000008	0.003432	0.000036	0.103166	0.000373	40.30	11.29	± 1.29	35.70	± 8.08	
65.0	0.000240	0.000011	0.002542	0.000061	0.000117	0.000006	0.004726	0.000040	0.102772	0.000185	50.89	10.23	± 0.78	32.37	± 4.92	
67.0	0.000426	0.000018	0.000944	0.000016	0.000438	0.000010	0.027026	0.000093	0.408434	0.000378	74.43	11.03	± 0.22	34.89	± 1.36	
68.0	0.000582	0.000014	0.000538	0.000015	0.001252	0.000019	0.088837	0.000111	1.128845	0.001001	86.77	10.95	± 0.07	34.64	± 0.44	
68.5	0.000245	0.000009	0.000454	0.000016	0.000804	0.000030	0.060414	0.000092	0.708257	0.000657	92.97	10.78	± 0.07	34.10	± 0.41	
69.0	0.000539	0.000018	0.000494	0.000021	0.003036	0.000036	0.233433	0.000765	2.683464	0.002412	94.93	10.89	± 0.06	34.43	± 0.37	

Laser intensity (%)	$^{36}\text{Ar} \pm \sigma$ (V)		$^{37}\text{Ar} \pm \sigma$ (V)		$^{38}\text{Ar} \pm \sigma$ (V)		$^{39}\text{Ar} \pm \sigma$ (V)		$^{40}\text{Ar} \pm \sigma$ (V)		$^{40}\text{Ar}^*$ (%)	$^{40}\text{Ar}^*/^{39}\text{Ar}_K \pm \sigma$		Age $\pm \sigma$ (Ma)	
69.5	0.000348	0.000014	0.000511	0.000032	0.002736	0.000029	0.209433	0.000428	2.354989	0.000698	96.65	10.84	± 0.05	34.28	± 0.32
70.0	0.000172	0.000011	0.000441	0.000009	0.001259	0.000032	0.094253	0.000250	1.047443	0.000968	97.50	10.76	± 0.07	34.04	± 0.41
70.5	0.000514	0.000017	0.000714	0.000026	0.004189	0.000022	0.320092	0.000718	3.581725	0.001831	96.47	10.78	± 0.05	34.09	± 0.31
71.0	0.000600	0.000015	0.000788	0.000039	0.005849	0.000061	0.460460	0.000684	5.126524	0.003587	97.05	10.79	± 0.04	34.14	± 0.28
71.5	0.000360	0.000014	0.000635	0.000025	0.004007	0.000032	0.310486	0.000294	3.426218	0.000976	97.63	10.76	± 0.04	34.02	± 0.27
72.0	0.000305	0.000016	0.000693	0.000025	0.003207	0.000021	0.248033	0.000398	2.738765	0.001375	97.61	10.75	± 0.05	34.02	± 0.30
72.5	0.000288	0.000020	0.000510	0.000018	0.002027	0.000044	0.157638	0.000233	1.745848	0.000959	96.48	10.65	± 0.06	33.68	± 0.37
80.0	0.000178	0.000017	0.000476	0.000019	0.001184	0.000030	0.091041	0.000247	1.016305	0.000827	97.06	10.77	± 0.08	34.06	± 0.49
64ZK12-284		$J=0.0017240 \pm 0.0000040$ (0.23%)						$D=1.0012 \pm 0.0036$							
61.0	0.020630	0.000129	0.000900	0.000025	0.004022	0.000060	0.006517	0.000078	6.121811	0.004424	1.17	10.99	± 14.63	33.86	± 89.32
62.0	0.004107	0.000047	0.000545	0.000028	0.000849	0.000028	0.005840	0.000062	1.278517	0.001172	6.83	14.92	± 3.86	45.82	± 23.38
63.0	0.002607	0.000025	0.000419	0.000026	0.000545	0.000011	0.007189	0.000067	0.840856	0.000804	10.81	12.55	± 1.94	38.62	± 11.80
64.0	0.001254	0.000031	0.000563	0.000017	0.000381	0.000014	0.008665	0.000070	0.460069	0.000548	23.81	12.42	± 1.32	38.22	± 8.03
64.5	0.001054	0.000020	0.000461	0.000017	0.000332	0.000010	0.011347	0.000034	0.413398	0.000233	30.48	10.84	± 0.64	33.41	± 3.89
65.5	0.001094	0.000032	0.000476	0.000025	0.000448	0.000012	0.017808	0.000091	0.517917	0.000440	42.50	12.12	± 0.59	37.31	± 3.62
66.5	0.001204	0.000038	0.000490	0.000030	0.000743	0.000014	0.035040	0.000192	0.735076	0.000597	54.71	11.35	± 0.36	34.95	± 2.21
67.5	0.001217	0.000024	0.000556	0.000014	0.000928	0.000017	0.050565	0.000192	0.904918	0.000559	62.67	11.11	± 0.19	34.23	± 1.16
68.0	0.001122	0.000022	0.000486	0.000019	0.001049	0.000029	0.059269	0.000118	0.961498	0.000838	67.70	10.89	± 0.15	33.54	± 0.92
68.5	0.001898	0.000019	0.000576	0.000022	0.002096	0.000042	0.120810	0.000180	1.870853	0.001575	71.17	10.97	± 0.09	33.81	± 0.57
69.0	0.004536	0.000045	0.000795	0.000025	0.005154	0.000075	0.310857	0.000324	4.691555	0.001988	71.96	10.84	± 0.09	33.42	± 0.52
69.5	0.003463	0.000036	0.001034	0.000028	0.007483	0.000068	0.483766	0.000906	6.249400	0.004370	84.02	10.84	± 0.06	33.41	± 0.36
70.0	0.000759	0.000018	0.000700	0.000015	0.003366	0.000041	0.232350	0.000433	2.722930	0.001762	92.57	10.82	± 0.05	33.33	± 0.33
71.0	0.000907	0.000024	0.000871	0.000018	0.004903	0.000064	0.344624	0.000337	4.039907	0.002276	93.95	10.99	± 0.05	33.87	± 0.29
72.0	0.000665	0.000023	0.000414	0.000027	0.000624	0.000020	0.039880	0.000199	0.621748	0.000509	72.23	11.09	± 0.21	34.18	± 1.28
BXC-3		$J=0.0017910 \pm 0.0000039$ (0.22%)						$D=1.0012 \pm 0.0036$							
61.0	0.000758	0.000021	0.000392	0.000014	0.000221	0.000014	0.003331	0.000028	0.227247	0.000254	9.03	5.98	± 2.15	19.21	± 13.78
62.0	0.000393	0.000005	0.000439	0.000018	0.000172	0.000008	0.006882	0.000043	0.117067	0.000163	15.73	2.49	± 0.44	8.03	± 2.82
63.5	0.000725	0.000020	0.000741	0.000019	0.000270	0.000019	0.010233	0.000041	0.264522	0.000319	26.19	6.57	± 0.67	21.12	± 4.27
65.0	0.001455	0.000012	0.000700	0.000020	0.000422	0.000010	0.010021	0.000032	0.512657	0.000437	19.96	10.07	± 0.72	32.26	± 4.57
66.0	0.000443	0.000025	0.000388	0.000021	0.000176	0.000008	0.005536	0.000027	0.185257	0.000311	39.99	12.84	± 1.41	41.01	± 8.88
67.0	0.001870	0.000029	0.000434	0.000034	0.000520	0.000015	0.012632	0.000073	0.685050	0.000779	22.32	11.99	± 0.93	38.32	± 5.86

Laser intensity (%)	$^{36}\text{Ar} \pm \sigma$ (V)		$^{37}\text{Ar} \pm \sigma$ (V)		$^{38}\text{Ar} \pm \sigma$ (V)		$^{39}\text{Ar} \pm \sigma$ (V)		$^{40}\text{Ar} \pm \sigma$ (V)		$^{40}\text{Ar}^*$ (%)	$^{40}\text{Ar}^*/^{39}\text{Ar}_K \pm \sigma$		Age $\pm \sigma$ (Ma)	
67.5	0.000571	0.000017	0.000448	0.000017	0.000195	0.000008	0.005872	0.000048	0.222204	0.000324	32.90	12.03	± 1.02	38.47	± 6.47
68.0	0.001854	0.000023	0.000538	0.000021	0.000601	0.000019	0.017695	0.000049	0.718218	0.000757	26.65	10.71	± 0.59	34.28	± 3.72
68.8	0.006518	0.000033	0.000664	0.000023	0.002550	0.000046	0.100776	0.000184	3.048962	0.002696	37.77	11.41	± 0.29	36.48	± 1.85
69.3	0.001646	0.000019	0.000525	0.000017	0.001830	0.000023	0.117881	0.000205	1.830453	0.001393	74.78	11.57	± 0.09	37.00	± 0.57
69.8	0.001128	0.000014	0.000670	0.000012	0.002679	0.000028	0.193103	0.000143	2.537186	0.001432	87.86	11.52	± 0.05	36.83	± 0.34
70.3	0.000767	0.000020	0.000705	0.000018	0.004205	0.000029	0.316207	0.000382	3.814825	0.002700	94.71	11.41	± 0.05	36.50	± 0.31
70.8	0.000657	0.000018	0.000703	0.000027	0.005445	0.000055	0.415374	0.000482	4.881737	0.001409	96.54	11.34	± 0.05	36.26	± 0.29
71.3	0.000580	0.000019	0.000631	0.000021	0.005575	0.000068	0.429817	0.000294	5.002837	0.003647	97.07	11.29	± 0.04	36.11	± 0.28
71.8	0.000538	0.000014	0.000594	0.000025	0.005440	0.000056	0.421324	0.000346	4.891436	0.002203	97.26	11.28	± 0.04	36.09	± 0.28
72.3	0.000404	0.000015	0.000555	0.000021	0.002791	0.000023	0.215873	0.000155	2.527728	0.002269	96.26	11.25	± 0.05	35.99	± 0.31
80.0	0.000394	0.000021	0.000348	0.000020	0.000193	0.000012	0.009204	0.000070	0.201163	0.000117	52.99	11.43	± 0.76	36.56	± 4.82

Table 3 Major oxide (wt %) and trace element (ppm) compositions of Paleogene lamprophyres in western Yunnan

Sample Locality Age(Ma)	YA-3 Yao'an 33.7	BXC-3 Machangqing 36.5	BXC-10 Machangqing	BXC-16 Machangqing	BY-6 Beiya 34.4	BY-L Beiya	ZK500-142.5 Beiya	BY-L(N-S) Beiya	64ZK12-284 Beiya 33.9	64ZK12-290 Beiya	BC-3 Binchuan	JC-13.3 Jianchuan
SiO ₂	47.61	50.88	49.45	50.10	54.19	51.94	45.21	48.22	46.55	43.51	51.76	54.80
TiO ₂	1.06	0.65	0.99	0.72	0.93	1.04	1.61	0.89	1.13	1.19	0.82	0.96
Al ₂ O ₃	11.09	10.86	13.99	11.62	13.56	13.86	12.10	13.25	13.29	13.52	14.83	13.90
FeO	4.44	3.75	5.26	5.95	3.07	2.39	1.11	3.13	9.68	9.14	5.55	1.85
Fe ₂ O ₃	2.52	2.53	2.47	2.18	2.27	3.50	6.54	4.22	2.70	3.92	4.29	6.35
MnO	0.16	0.11	0.17	0.14	0.10	0.09	0.11	0.12	0.35	0.33	0.06	0.16
MgO	6.00	10.80	9.84	12.42	5.72	4.85	9.16	6.44	4.61	5.71	6.72	3.39
CaO	8.57	5.61	8.39	6.90	5.09	5.10	8.06	7.71	4.55	3.94	4.65	6.38
Na ₂ O	0.56	1.23	3.12	2.51	2.36	2.91	1.48	2.43	2.93	2.98	4.41	3.44
K ₂ O	8.22	6.13	3.28	2.63	5.57	6.18	4.51	5.92	4.04	4.14	3.50	6.03
P ₂ O ₅	1.42	0.49	0.86	0.63	0.73	0.79	1.23	0.56	0.71	0.76	0.84	0.69
H ₂ O ⁺	1.30	2.80	1.20	2.22	2.42	2.26	5.62	1.96	2.44	3.42	1.58	1.12
CO ₂	6.07	3.46	0.23	0.41	2.87	4.74	2.19	5.23	6.58	7.37	0.18	0.32
Total	99.00	99.31	99.25	98.42	98.87	99.65	98.93	100.08	99.56	99.93	99.19	99.39
LOI	6.10	5.78	0.90	2.15	4.39	6.43	8.05	6.53	8.28	10.04	1.30	1.32
Mg#	0.62	0.76	0.70	0.74	0.67	0.61	0.70	0.63	0.41	0.45	0.56	0.45
Sc	19.9	16.8	23.1	21.2	13.7	13.1	17.6	22.0	18.3	19.1	16.7	17.4
V	133.0	115.0	164.0	148.0	93.2	104.0	126.0	182.0	150.0	160.0	128.0	167.0
Cr	122.0	778.0	507.0	524.0	207.0	255.0	364.5	411.0	271.0	286.0	375.0	17.6
Co	22.1	39.0	39.0	35.7	24.6	25.5	39.5	34.3	30.1	30.2	17.7	20.2
Ni	59.5	374.0	210.0	403.0	163.0	164.0	132.5	188.0	117.0	127.0	78.1	14.6
Cu	232.0	30.1	10.1	22.2	56.5	36.7	32.3	64.0	38.4	42.2	1274.0	44.7
Zn	56.5	151.0	92.1	85.7	87.4	126.3	86.3	95.0	131.0	89.0	113.0	105.0
Ga	23.5	15.1	22.0	22.5	21.5	20.7	16.4	16.9	17.6	17.6	26.3	19.2
Ge	2.4	1.5	1.9	2.7	1.6	1.1	1.3	1.4	1.1	1.1	1.5	1.4
As	19.1	2.3	1.5	1.4	6.5	4.9	1.6	5.3	22.2	10.6	1.4	4.7
Rb	341	271	305	263	281	246	159	278	160	157	189	222
Sr	1914	1006	2104	1195	953	850	782	1229	541	483	1674	1273
Y	50.3	14.9	27.3	27.0	19.4	22.2	23.2	23.5	18.1	17.4	29.3	38.2
Zr	723	127	219	179	353	369	491	154	201	208	275	256
Nb	28.9	6.3	14.8	15.2	19.9	18.1	32.4	7.8	22.4	22.5	15.1	11.0

Sample Locality Age(Ma)	YA-3 Yao'an 33.7	BXC-3 Machangqing 36.5	BXC-10 Machangqing	BXC-16 Machangqing	BY-6 Beiya 34.4	BY-L Beiya	ZK500-142.5 Beiya	BY-L(N-S) Beiya	64ZK12-284 Beiya 33.9	64ZK12-290 Beiya	BC-3 Binchuan	JC-13.3 Jianchuan
Sn	2.74	1.52	2.12	7.13	3.49	3.30	3.04	1.89	2.61	2.40	8.40	2.66
Sb	1.48	0.79	0.21	0.82	1.02	3.00	0.17	0.95	7.59	9.62	0.17	0.31
Cs	13.2	36.3	35.9	20.0	36.4	4.2	3.4	19.2	21.2	13.0	8.7	6.4
Ba	5454	1312	2792	1822	1452	1636	1769	1785	1144	1057	2942	1896
La	168	27	152	196	45	50	75	39	24	23	73	51
Ce	341	55	282	363	93	104	162	84	49	48	168	108
Pr	40.2	6.6	30.8	37.5	10.9	11.8	17.4	9.8	5.8	5.8	19.2	12.2
Nd	154	26	110	127	42	47	68	40	24	23	79	49
Sm	27.5	5.2	16.6	17.7	8.1	8.8	11.7	7.6	4.8	4.7	14.8	10.2
Eu	6.94	1.28	4.04	3.59	2.02	2.09	2.88	1.87	1.31	1.33	3.50	2.50
Gd	22.50	4.85	11.70	11.75	6.90	8.00	9.28	7.06	4.93	4.86	12.20	9.94
Tb	2.43	0.58	1.24	1.22	0.83	0.94	1.03	0.87	0.66	0.63	1.31	1.28
Dy	11.60	3.05	5.94	5.89	4.32	4.69	5.10	4.78	3.76	3.69	6.40	7.21
Ho	1.84	0.54	1.00	0.98	0.74	0.83	0.88	0.88	0.72	0.69	1.08	1.40
Er	4.48	1.51	2.60	2.63	1.87	2.12	2.29	2.45	1.87	1.83	2.77	3.93
Tm	0.53	0.20	0.34	0.32	0.24	0.26	0.28	0.32	0.25	0.24	0.34	0.55
Yb	3.08	1.33	1.98	1.95	1.48	1.64	1.74	2.07	1.59	1.46	2.02	3.57
Lu	0.44	0.20	0.29	0.27	0.21	0.23	0.25	0.31	0.23	0.21	0.28	0.53
Hf	16.2	3.1	4.8	3.9	8.7	9.1	11.8	4.1	5.5	5.6	6.0	6.0
Ta	1.30	0.47	0.73	0.74	1.68	1.14	1.92	0.51	1.58	1.53	0.69	0.76
W	5.68	1.70	12.80	1.28	3.30	2.75	0.30	1.36	1.79	2.27	0.23	4.98
Pb	59.0	17.7	5.6	2.8	92.6	37.0	22.9	32.2	21.0	18.9	19.2	34.0
Th	27.3	8.6	22.0	23.3	14.8	15.7	13.0	9.4	7.9	8.0	29.1	14.7
U	6.28	2.96	9.08	4.86	6.50	9.37	2.50	3.35	3.35	3.13	9.61	4.84
Eu/Eu*	0.85	0.78	0.89	0.76	0.83	0.76	0.85	0.78	0.83	0.85	0.80	0.76
Rb/Sr	0.2	0.3	0.1	0.2	0.3	0.3	0.2	0.2	0.3	0.3	0.1	0.2
Ba/Rb	16.0	4.8	9.2	6.9	5.2	6.7	11.2	6.4	7.2	6.7	15.5	8.5
Nb/U	4.6	2.1	1.6	3.1	3.1	1.9	13.0	2.3	6.7	7.2	1.6	2.3
Ce/Pb	5.8	3.1	50.7	129.4	1.0	2.8	7.1	2.6	2.3	2.5	8.8	3.2

Mg# = MgO/(MgO+FeOT) in atomic ratio; LOI=loss of ignition; Age (Ma) cited are the phlogopite ^{40}Ar - ^{39}Ar dates from this study.

Table 4 Sr-Nd-Pb isotope data for Paleogene lamprophyres in western Yunnan

Sample	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm 2\sigma$	$(^{87}\text{Sr}/^{86}\text{Sr})_i$	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\pm 2\sigma$	$(^{143}\text{Nd}/^{144}\text{Nd})_i$	$\epsilon_{\text{Nd}(t)}$	T_{DM} (Ma)
YA-3	0.5156	0.709715	14	0.709459	0.1079	0.512078	8	0.512053	-10.53	1542
BY-6	0.8532	0.708375	15	0.707951	0.1178	0.512362	10	0.512335	-5.03	1252
64ZK12-284	0.8532	0.706573	9	0.706149	0.1209	0.512575	5	0.512547	-0.89	945
BXC-3	0.7794	0.708221	18	0.707834	0.1223	0.512438	5	0.512410	-3.57	1187
BXC-10	0.4194	0.706575	16	0.706367	0.0912	0.512344	5	0.512323	-5.26	1003
BC-3	0.3269	0.707010	15	0.706847	0.1127	0.512219	5	0.512193	-7.80	1403

Table 4 continued

Sample	$^{206}\text{Pb}/^{204}\text{Pb}$	$\pm 2\sigma$	$^{207}\text{Pb}/^{204}\text{Pb}$	$\pm 2\sigma$	$^{208}\text{Pb}/^{204}\text{Pb}$	$\pm 2\sigma$	$(^{206}\text{Pb}/^{204}\text{Pb})_i$	$\pm 2\sigma$	$(^{207}\text{Pb}/^{204}\text{Pb})_i$	$\pm 2\sigma$	$(^{208}\text{Pb}/^{204}\text{Pb})_i$	$\pm 2\sigma$
YA-3	18.1588	0.0007	15.5683	0.0007	38.7549	0.0019	18.1220	0.0022	15.5666	0.0007	38.7023	0.0035
BY-6	18.6450	0.0006	15.6808	0.0007	39.0223	0.0018	18.6204	0.0015	15.6797	0.0007	39.0039	0.0021
64ZK12-284	18.6664	0.0006	15.7018	0.0005	39.1051	0.0011	18.6105	0.0032	15.6992	0.0005	39.0617	0.0027
BXC-3	18.7080	0.0006	15.6416	0.0005	38.8922	0.0012	18.6496	0.0034	15.6389	0.0005	38.8364	0.0034
BXC-10	19.1227	0.0007	15.6588	0.0006	39.3620	0.0016	18.5453	0.0327	15.6318	0.0016	38.9018	0.0261
BC-3	18.4016	0.0007	15.6090	0.0006	38.9025	0.0014	18.2275	0.0099	15.6009	0.0008	38.7291	0.0099

$^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ are calculated using whole-rock Rb, Sr, Sm and Nd values in Table 3.

Chondritic Uniform Reservoir (CHUR) at the present day [$(^{87}\text{Rb}/^{86}\text{Sr})_{\text{CHUR}} = 0.0847$ (McCulloch and 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 and Wasserburg, 1980); $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} = 0.512638$ (Goldstein et al., 1984)] were used for the calculations; Nd depleted mantle model ages (T_{DM}) were calculated using $(^{147}\text{Sm}/^{144}\text{Nd})_{\text{DM}} = 0.2137$ and $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{DM}} = 0.51315$ (Peucat et al., 1989) at the present day. $\lambda_{\text{Rb}} = 1.42 \times 10^{-11} \text{ year}^{-1}$ (Steiger and Jager, 1977); $\lambda_{\text{Sm}} = 6.54 \times 10^{-12} \text{ year}^{-1}$ (Lugmair and Marti, 1978). Initial Sr-Nd-Pb isotopic ratios are corrected at $t = 35 \text{ Ma}$.