The cause and source of melting for the most recent volcanism in Tibet: a combined geochemical and geophysical perspective

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Abstract

We investigate the youngest volcanic activity on the Tibetan Plateau by combining observations from petrologic, geochemical and seismic tomography studies. Recent (from 2.80 Ma to present) post-collisional potassium-rich lavas from the Ashikule

Volcanic Basin (AVB) in northwestern Tibet are characterised by remarkably enriched light rare earth elements (LREE) relative to heavy rare earth elements (HREE), and enriched large ion lithophile element (LILE) relative to high field strength elements (HFSE). Strontium and neodymium isotopic compositions are surprisingly restricted, and show little evidence for mixing or crustal contamination, despite the thick crust upon which they are erupted. Geochemical characteristics indicate a homogeneous source, highly enriched in trace elements, which is most consistent with derivation from long-lived subcontinental lithospheric mantle (SCLM). P-wave anisotropy tomography documents a gap between the north-subducting Indian slab and south-subducting Tarim slab directly beneath the AVB. We propose that volcanism in northwestern Tibet is associated with the progressive closure of this gap, during which shear heating of the SCLM can generate localised melting, with deep-seated faults providing a mechanism for erupted lavas to escape large-scale crustal contamination and fractionation in magma reservoirs. Thus, shear heating may provide an explanation for the restricted range of radiogenic isotope compositions from a SCLM source that should be, by its nature, heterogeneous on a large scale.

1. Introduction

The Tibetan Plateau is the most protuberant structure on Earth, formed in

45   response to the collision of the Indian and Asian plates from 50-55 Ma ago (e.g.,

46   Royden et al., 2008; Tapponnier et al., 2001). Since the initial collision, the Indian

47   Plate has continued to subduct northward with recent studies depicting it

48   underthrusting the entire Tibetan Plateau to its northwestern margin and colliding with

49   the south-directed subducting Tarim Block (e.g., Tunini et al., 2016, Wei et al., 2015,

50   2016; Zhao et al., 2010). Collision with the Tarim block likely contributed to the rapid

51   uplift of the Tibetan Plateau at ~13 Ma (Turner et al., 1993), and resulting global

52   climate change (e.g., Kutzbach et al., 1993; Zhisheng et al., 2001).

53   A striking feature of the plateau, and northern Tibet in particular, is its widespread

54   Cenozoic post-collisional magmatism, which is distributed between 78°E and 92°E

55   and between 34°N and 36°N, forming a volcanic belt ~1300 km in length with a 100

56   km western and 400 km eastern width (e.g., Arnuad et al., 1992; Chung et al., 2005;

57   Cooper et al., 2002; Ding et al., 2003; Guo et al., 2006, 2014; Royden et al., 2008;

58   Turner et al., 1993, 1996; Wang et al., 2005; Xia et al., 2011). This volcanic belt

59   consists of more than fifty volcanic clusters that can be spatially divided into four

60   domains: (1) the Qiangtang volcanic area; (2) the Hoh Xil and middle Kunlun

61   volcanic area; (3) the west Kunlun volcanic area; (4) the Karakoram volcanic area

62   (Mo et al., 2006). The rich literature of previous studies on the post-collision lavas in

63   north Tibet shows that these Cenozoic K-rich lavas generally show remarkable

64   enrichment of LILE and LREE relative to HFSE and HREE. The petrogenesis of

65   K-rich continental lavas in Tibet, and in general worldwide, is a topic of long-standing

66   debate, with hypotheses ranging from the involvement of plumes (Guo et al., 2006),

67   crustal melting (Hacker et al., 2014; Wang et al., 2005, 2016), sub-continental

68   lithospheric mantle (Conticelli et al., 2009; Guo et al., 2015; Huang et al., 2015;

69   Turner et al., 1993, 1996; Williams et al., 2004), and contributions from subducted

70   slabs and recycled mélanges (Guo et al., 2014).

71   Temporally, magmatism in Tibet spans a period of ~45 Ma: 45-26 Ma in central

72   Tibet, 26-8 Ma in southern Tibet and 20 Ma to historic activity in northern Tibet (e.g.,

73   Guo et al., 2006, 2014; Turner et al., 1993, 1996; Williams et al., 2004; Xia et al.,

74   2011). The most recent eruption on Tibetan Plateau occurred in 1951 AD from the

75   Ashi Volcano in the Ashikule Volcanic Basin (AVB; Liu and Maimaiti, 1989; Xu et al.,

76   2012, 2014).

77   Several geophysical studies suggested that the Indian plate and Tarim Block

78   collide below the northwestern Tibetan Plateau (e.g., Tunini et al., 2016; Zhao et al.,

79   2010). However, recent tomographic studies focussed on the AVB show that the

80   volcanic basin is in fact located above a gap in the mantle lithosphere between the

81   Indian plate and Tarim Block (Wei et al., 2015, 2016; Fig. 2). The upwelling of

82   asthenospheric mantle through this gap may be responsible for the generation of the

83   Ashikule volcanoes (Fig.2; Wei et al., 2015, 2016). Because of the inaccessibility of

84   the AVB (it has an average altitude of ~4800 m and poor transport infrastructure due

85   to the sparse human population), the petrogenesis of its lavas has not been studied

86   systematically. The combined studies of Williams et al. (2004) and Guo et al. (2006,

87   2014) present some chemical analyses of nine samples from the region, but focused

88   more broadly on magmatism in the whole of northern Tibet, and thus necessarily

89   treated the limited samples from the AVB as an integrated unit. Cooper et al (2002)

90   studied the isotopic and chemical characteristics of lavas in the youngest episode,

91   which are only a part of the AVB.

92   The occurrence of present day magmatic activity and improved seismic imaging

93   allows us to combine geochemical and geophysical information. Understanding this

94   region has direct impact on evaluating hypotheses for the generation of continental

95   K-rich lavas in general and on the style and evolution of magmatism across northern

96   Tibet in particular. We present whole rock major, trace and Sr-Nd isotopic

97   compositions of 25 lavas encompassing all six volcanic episodes in the AVB

98   suggested by Liu and Maimaiti (1989) and Xu et al. (2014). In combination with

99   previous chronological data (Liu and Maimaiti, 1989; Xu et al., 2014) and

100   geophysical results (e.g., Tunini et al., 2016; Wei et al., 2015, 2016; Wittlinger et al.,

101   2004; Zhao et al., 2010), we examine the petrogenesis of recent AVB volcanism with

102   relation to its geodynamic setting.

103

104   ## 2. Geological background

105   The AVB is located in the western Kunlun orogenic belt (Fig. 1), which marks the

106   northwestern margin of Tibetan Plateau (Wang et al., 2003; Wang, 2004) where the

107   Tarim Block has subducted southward, perhaps to a depth of 300 km (e.g., Tunini et

108   al., 2016; Wei et al., 2016; Wittlinger et al., 2004; Zhao et al., 2010). Structurally

109   controlled by Altyn Tagh, Karakax and Longmu-Gozha Co fault systems (Fig. 1a), the

110   AVB can be characterised as a pull-apart basin (Li et al., 2006), suffering dominantly

111 EW extensional stress and shearing (Furuya and Yasuda, 2011; Xu et al., 2012). The

112 AVB is seismically active and three strong earthquakes have occurred close to this

113 basin in 2008 (Mw 7.1), 2012 (Mw 6.2) and 2014 (Mw 6.9) (Fig. 1a; Bie and Ryder,

114 2014). The 2008 Yutian earthquake (Mw 7.1) ruptured a normal fault in the AVB,

115 resulting in a surface rupture zone of up to 50 km (Bie and Ryder, 2014; Elliott et al.,

116 2010; Furuya and Yasuda, 2011; Xu et al., 2013).

117 The AVB covers an area of ~700 km^2 and encompasses 14 main volcanoes, with

118 activity spanning from 2.80 Ma to 1951 AD (Fig. 1b). The main volcanological

119 features in the region are volcanic cones, lava terraces and lava valleys (Xu et al.,

120 2012). The largest and highest volcano in the basin is Dahei volcano, which has an

121 altitude of ~5090 m above sea level and a relative altitude of ~350 m. Most volcanoes

122 in the basin are characterised by explosive eruptions (e.g., Ashi, Dahei and Wuluke

123 volcanoes), accompanied by some effusive eruptions in the west and east of the basin

124 (e.g., Xishan volcano and Dongshan volcanoes). Chronological studies (Liu and

125 Maimaiti, 1989; Xu et al., 2014) show that most volcanoes were active in the

126 Quaternary. Six episodes of volcanic activity have been defined as follows: 1. Late

127 Ashi Episode (in 1951; Liu and Maimaiti, 1989), 2. Ashi Episode (0.12-0.17 Ma), 3.

128 Wuluke Episode (0.20-0.29 Ma), 4. Dahei Episode (0.46-0.60 Ma), 5. Mati Episode

129 (1.02-1.65 Ma), 6. Xi Episode (2.34-2.8 Ma).

130

131 ## 3. Sample selection and petrography

132 Two field campaigns were conducted by Institute of Geology, China Earthquake

133   Administration in May of 2011 and 2015 where over 140 lava samples were collected.

134   From these, twenty-five representative samples were chosen to cover the main

135   volcanoes in the AVB according to the geological map (Fig. 1b) revised from Xu et al.

136   (2014) and Liu and Maimaiti (1989). A thin section of each sample was made for

137   petrographic analysis from the same material as was processed for whole-rock

138   chemistry.

139   All lavas studied are porphyritic, with phenocrysts of plagioclase, clinopyroxene,

140   orthopyroxene, phlogopite, rare olivine and Fe-Ti oxides in a ground-mass consisting

141   of volcanic glass, plagioclase laths, clinopyroxene, orthopyroxene and rare olivine

142   and Fe-Ti oxides (visually estimated volume percentages are presented in Table 1).

143   The dominant phenocryst phase is plagioclase, which displays a notable lack of

144   commonly developed zoning. All samples are vesicular, ranging from 10% to 80% by

145   volume. A low degree of vesicular refilling by secondary minerals (calcite and clay

146   minerals) is petrographically present in some samples (ASKL-15, ASKL-16,

147   ASKL-17, ASKL-18, 515-1, 516-2). These obvious alteration products were removed

148   by hand-picking the crushate before reducing the sample to powder. The phenocrysts

149   and groundmass of analysed samples are unaltered. Figure 3 depicts representative

150   textures encountered in the AVB lavas.

151

152   ## 4. Methods

153   ### 4.1. Major and trace element analysis

154   Visibly altered portions of the rocks were removed and fresh material was

155   reduced to centimetre size by jaw crusher. Material was then hand-picked and

156   hand-crushed in agate mortar to fine powders for whole-rock major, trace element and

157   Sr-Nd analysis. Whole-rock major element analysis was conducted by X-ray

158   fluorescence (XRF) at Activation Laboratories Ltd. (Ontario, Canada;

159   www.actlabs.com).

160   High-precision trace element analysis was undertaken at the Open University, UK,

161   with an Agilent 8800 ICP-QQQ ('triple-quad' inductively coupled plasma mass

162   spectrometer). Sample digestion was performed at the Mass Spectrometry and

163   Geochemistry Labs at Imperial College London (MAGIC). Approximately 100 mg of

164   rock powder was dissolved in sealed Teflon vials with a 3:1 mixture of distilled

165   concentrated HF: HNO_3. The solutions were ultra-sonicated for 25 minutes and heated

166   at 160 °C hot plate for at least 24 hours. They were then evaporated to near dryness

167   and re-dissolved in 2 ml distilled 6M HCl at 120 °C for at least 24 hours. The

168   solutions were then evaporated to complete dryness at 120 °C and re-dissolved in ~1

169   ml concentrated HNO_3 and evaporated at 180 °C. The last re-dissolution and

170   evaporation step was repeated at least three times until the samples turned brownish

171   or brown, indicating the destruction of fluorides from initial HF dissolution. Finally, 2%

172   distilled HNO_3 was added to samples for a 1000-fold dilution.

173   Diluted samples were aspirated into the ICP-MS using a quartz microflow

174   nebuliser, with an uptake rate of 0.5ml per minute, and count rates in the order of 1 –

175   5 x 10^7 cps/ppm. Analyses were performed in three different collision/reaction cell

176   modes (no gas, He, O_2 mass shift). Oxide levels (measured as CeO/Ce) were kept low,

177   at 1% in no gas, and 0.5% in He collision mode, and doubly charged species

178   (Ce++/Ce+) at 1.6% in no gas, and 1.2% in He collision mode. Analyses were

179   standardised against five reference materials (digested at both the Open University

180   and Imperial College London) that were measured at the beginning of each analytical

181   run. The reference materials were selected on the basis of their similarity to the

182   samples analysed, and include BIR-1, W-2, GSP-2, BHVO-2 and AGV-1. An internal

183   standard solution (consisting of Be, Rh, In, Tm, Re, and Bi) was added to samples and

184   run on-line throughout all analyses to correct for any instrumental drift. Drift was

185   further monitored with a measurement block consisting of USGS reference material

186   BIR-1 (separate digest to that used in the standardisation), a 2% HNO_3 blank, and a

187   repeated unknown sample (ASKL-7) performed every five unknown measurements.

188

189   *4.2. Sr-Nd isotopes*

190   *4.2.1. Chemical isolation of Sr and Nd*

191   Approximately 50 mg of sample powder was leached following the method of

192   Weis and Frey (1991, 1996) and digested using the same method as trace element

193   digestion described above. Strontium and Nd were separated by a three-stage column

194   chemistry. The first column employs 1.4 ml volume Bio-Rad AG50X8 200-400 mesh

195   cation exchange resin, to separate Sr and rare earth elements (REE) from the sample

196   matrix. Samples were loaded with 1M HCl, and the Sr fractions were collected in 4

197   ml 3M HCl and 0.5 ml 6M HCl. REE cuts were subsequently collected in 6.5 ml 6M

198   HCl. The second column purifies the Sr fraction using 100 ul shrink Teflon columns

199 and 100-150 mesh Eichrom Sr resin. Strontium fractions were loaded with 300 ul 3M

200 HNO$_3$ and purified Sr solutions were collected with 4 ml 0.05M HNO$_3$. The third

201 stage column chemistry separates Nd from the REEs using a Teflon column loaded

202 with 1 mL of Eichrom Ln-Spec resin (100-150 um). REE cuts were loaded with 200

203 ul 0.2M HCl and Nd was collected in 6 ml 0.2M HCl.

204

205 *4.2.2. Mass Spectrometry*

206 Strontium and neodymium isotopic analyses were performed in the MAGIC

207 laboratories at Imperial College London. Strontium isotopes employed a Thermo

208 Finnigan Triton thermal ionization mass spectrometer (TIMS) whilst Nd isotope

209 analysis used a Nu Instruments HR Multi-collector Inductively Coupled Plasma Mass

210 Spectrometer (MC-ICPMS).

211 Strontium fractions were dissolved in 6M distilled HCl and loaded on single

212 zone-refined rhenium filaments, with a TaCl$_5$ activator to enhance Sr ionization.

213 Faraday cups collected masses ^{83}Kr, ^{84}Sr, ^{85}Rb, ^{86}Sr, and ^{88}Sr on channels L4, L3, L2,

214 L1, C, and H1, respectively. Data were collected in static mode with virtual amplifier

215 rotation, using a dummy mass of ^{83}Kr in L4. Measurement consisted of 12 blocks of

216 15 cycles with an electronic baseline between each block. Mass fractionation was

217 corrected online using ^{86}Sr/^{88}Sr = 0.1194 and the exponential law. Potential

218 contribution of ^{87}Rb on ^{87}Sr was corrected online using ^{87}Rb/^{85}Rb = 0.3856. Standard

219 NIST SRM 987 was measured before and during measurement sessions and unknown

220 samples were normalised to the accepted value of ^{87}Sr/^{86}Sr = 0.710248 (Thirlwall,

221   1991).

222   Neodymium fractions were re-dissolved in 0.1M HNO_3 for analysis by

223   MC-ICPMS. Samples were introduced as 50 ppb solutions via a Nu DSN-100

224   Desolvation Nebuliser System. Masses ^{138}Ba, ^{140}Ce, ^{142}Nd, ^{143}Nd, ^{144}Nd, ^{145}Nd, ^{146}Nd,

225   ^{147}Sm, ^{148}Nd were collected on Faraday cups L4, L3, L2, L1, C, H1, H2, H3, and H4,

226   respectively. Measurements were collected in three blocks of 20 cycles, with an

227   electronic baseline in between blocks. Mass fractionation was corrected using

228    $^{146}Nd/^{145}Nd$ = 0.7219 and the exponential law. Neodymium standard JNdi was

229   measured at the beginning and during the measurement session, and unknown

230   samples were normalised according to the offset of JNdi in a single measurement

231   session from its accepted value ($^{143}Nd/^{144}Nd$ = 0.512115; Tanaka et al., 2000).

232

233   ## 5. Results

234   ### 5.1. Whole rock major and trace elements

235   ### 5.1.1. Major elements

236   United States Geological Survey (USGS) basaltic reference materials BIR-1a and

237   BHVO-2 were processed as unknowns for whole rock major element analyses and are

238   in good agreement with accepted values (see supplemental information). Whole rock

239   major element concentrations for AVB lavas are presented in Table 2. The samples

240   yield SiO_2 contents between 45.91 and 59.48 wt%. For classification, the major

241   element compositions were recalculated to 100% on an anhydrous basis, and plotted

242   on the total alkali versus silica (TAS) diagram (Fig. 4a). Most samples plot within the

243 trachyandesite field, with several relatively old samples (ASKL-14, ASKL-17,

244 ASKL-18, 516-11, 516-12) within the basaltic trachyandesite, phonotephrite and

245 basanite fields. Samples greater than 2 Ma, in particular, lie at distinctly higher total

246 alkali contents for their degree of evolution than the rest of the samples. In general, all

247 samples are characterised by high K_2O (3.48-4.47 wt%) and K_2O/Na_2O ratios (1.08 –

248 1.53) (Fig. 4b), and belong to the shoshonitic series. MgO contents range from 2.24 to

249 7.40 wt% with Mg# of 46 to 63, displaying negative correlations with SiO_2 contents

250 (Fig. 5a and Fig. 5b). One young sample (ASKL-3) from the recent eruption at Ashi

251 Volcano, and 4 relatively old samples (ASKL-17, ASKL-18, 516-11, 516-12) exhibit

252 higher MgO contents and Mg# numbers compared with other AVB samples. Given

253 that the lavas are not co-genetic, distinct liquid lines of descent cannot be constructed.

254 However, $Fe_2O_{3(T)}$ (5.96-10.04 wt%) and CaO (5.15-9.39 wt%) contents display

255 negative correlations with SiO_2, consistent in general with the fractionation of mafic

256 phases. TiO_2 (1.36-2.22 wt%) and Al_2O_3 (12.69-14.97 wt%) have less distinct

257 correlations with SiO_2, but support the extensive plagioclase fractionation and the

258 presence of oxides observed in thin sections (Fig. 5; Table 1).

259

260 *5.1.2. Trace elements*

261   Trace element compositions are given in Table 2. Based on repeated analyses of

262 sample ASKL-7 and USGS standard BIR-1a, the reproducibility of most elements was

263 better than 2% and the measured values of standards deviate < 5% from

264 recommended values for most elements (see supplementary information). Total

265   procedural blanks from digestion at MAGIC are also given in the supplemental

266   information and are negligible compared to the concentration of elements measured.

267   A primitive mantle normalised incompatible element plot is presented in Fig. 6a,

268   whilst a chondrite-normalised REE pattern is depicted in Figure 6b. Figure 6

269   illustrates the markedly elevated trace element concentrations of AVB lavas,

270   consistent with K-rich continental magmatism in general. The AVB samples are all

271   very similar in trace element composition, despite covering a range of > 2 Ma. They

272   are characterised by significant positive Ba, Th, Pb and Nd anomalies and negative

273   Nb, Ta, Zr and Ti anomalies. For comparison, the most isotopically enriched mantle

274   endmember composition of "EM II" calculated for Samoan ocean island basalts

275   (Jackson et al., 2007), global subducting sediment (GLOSS; Plank and Langmuir,

276   1998) and average upper continental crust (UCC, Rudnick and Gao, 2003) are also

277   plotted. It is noteworthy that the enrichment of LILEs and LREEs in AVB lavas is

278   more pronounced than all of the comparative suites.

279   The AVB lavas have $(La/Yb)_n$ ratios of 40 to 68, which correlate negatively with

280   MgO contents (Fig. 7a) and positively with La_n (La composition normalised to

281   primitive mantle, 131 to 295; Fig. 7b). All samples in the AVB are characterised by

282   negative Eu anomalies ($Eu/Eu^*<1$), which correlate positively with MgO contents,

283   consistent with plagioclase fractionation (Fig. 7c). Titanium anomalies (Ti/Ti^*) show

284   negative correlation with SiO_2, consistent with oxide fractionation (Fig. 7d). Ba/Th

285   ratios correlate negatively with Th/Nd ratios, whist Th/La ratios show less distinct

286   negative correlation with Sm/La ratios (Fig. 7e and Fig. 7f).

287

5.2. Sr-Nd isotope geochemistry

289   Table 3 presents Sr and Nd isotopic data for the AVB samples. USGS reference

290   materials BCR-2, BHVO-2 and BIR-1a were processed with all unknown samples and

291   are in excellent agreement with literature values (see supplemental information).

292   Despite the fresh appearance of samples in the field and the effort to exclude potential

293   alteration phase during creation of rock powders, we performed leaching tests on the

294   rock powders to exclude the effect of potential secondary alteration phases. Initial

295   tests of leached versus unleached powders showed no discernible difference in Nd

296   isotopes, but significant differences in Sr isotopes of up to 1640 ppm (see

297   supplemental information), thus all rock powders were leached according to the

298   method of Weis and Frey (1991, 1996).

299   All lavas have enriched $^{87}Sr/^{86}Sr$ (0.707490 – 0.710523) and $^{143}Nd/^{144}Nd$

300   (0.512265 – 0.512472) relative to bulk silicate earth (BSE) (Fig. 8). $^{87}Sr/^{86}Sr$ and

301   $^{143}Nd/^{144}Nd$ plot away from the bulk silicate Earth towards the enriched mantle II

302   endmember of Zindler and Hart (1986). The AVB data is remarkably restricted, with

303   two outliers belonging to the > 2 Ma lavas, which display slightly less radiogenic Sr

304   and more radiogenic Nd. Our data are in good accordance with the isotopic results of

305   younger lavas measured by TIMS (Cooper et al., 2002). Additionally, our most

306   primitive sample (ASKL-18) has similar Sr-Nd isotopic ratios to comparable literature

307   samples AH602 and AH609 of Guo et al. (2014).

308

6. Discussion

We examine the petrogenesis of the AVB lavas and attempt to link chemical and isotopic information with the unique tectonic setting as depicted by recent seismic tomography (Wei et al., 2016). First, we assess modifications to the primary lava signature. We then examine and evaluate common models for the generations of continental K-rich magmas to determine the most likely petrogenesis of the AVB lavas. Finally, we integrate recent seismic observations with geochemistry for a richer understanding of the interplay of tectonics and magma generation in the region.

6.1. Fractional crystallisation and crustal contamination

Due to the collision between the Indian and Asian plates, the thickness of the crust in northwestern Tibet has reached up to 90 km (Tunini et al., 2016; Wittlinger et al., 2004). Thus, there is ample opportunity for fractional crystallisation in crustal magma chambers and crustal assimilation to modify the chemical signature of magmas after they leave their source region(s).

Petrographic investigations of the Ashi Volcano carried out by Yu et al. (2014) proposed mixing between two crustal magma pockets based on distinct zoned phenocryst populations of pyroxene and plagioclase cores and rims. Orthopyroxene-liquid thermometry was used to estimate magmatic equilibrium temperature and pressure conditions, suggesting that an evolved magma pocket (13-18 km in depth) may connect with a deeper trachyandesite magma pocket (18-30 km in depth). Yu et al. (2014) speculated that because the two magma pockets are

331   vertically connected, the evolved magma pocket may be generated by fractional

332   crystallisation of the trachyandesite magma. Beyond the petrographic analysis of Yu

333   et al. (2014), there are only sparse reports of evolved erupted lavas in the region by

334   Deng et al. (1998).

335   The impact of fractional crystallisation can be generally demonstrated in our

336   dataset. For example, the correlations between Eu anomaly and MgO, and between Ti

337   anomaly and SiO_2 (Fig. 7c, 7d), provide evidence for plagioclase and oxide

338   fractionation, respectively. These phenocryst phases are also observed

339   petrographically (Table 1). Among the younger samples, the youngest episode lavas

340   (ASKL-3, ASKL-4 and 518-5) have the highest MgO contents, suggesting lowest

341   degrees of fractional crystallisation. Therefore, the most recent eruption may be

342   related to new magma inputs from the source.

343   With the exception of the three oldest samples (ASKL-17, ASKL-18, 516-11),

344   Sr-Nd isotopic ratios are remarkably uniform (Fig. 8). The restricted Sr-Nd signature

345   of the younger samples suggests a homogenous source, which is also supported by

346   their similar trace element patterns (Fig. 6). In plots of SiO_2 versus $^{87}Sr/^{86}Sr$ and

347    $^{143}Nd/^{144}Nd$ (Fig. 9a and Fig. 9b), younger samples do not show significant variation

348   of Sr and Nd isotopes over a modest range of SiO_2 contents (53.78% to 59.48%).

349   Therefore, despite the thick continental crust, significant crustal contamination in the

350   younger AVB lavas are unlikely, as no correlation with SiO_2 contents are observed. It

351   is also difficult to reconcile how crustal assimilation would produce near-identical

352   Sr-Nd isotope compositions over million year timescales. Additionally, the LILEs and

353 LREEs of the AVB lavas (i.e., Rb, Sr, Sm and Nd concentrations) are significantly

354 higher than those typically found in the continental crust (Fig. 6). Therefore, the

355 effects of crustal assimilation are not necessarily apparent in these lavas from simple

356 mass balance considerations. Thus, although some degree of crustal contamination

357 seems inevitable, it is unlikely to modify the magmatic chemistry significantly.

358

359 *6.2. Petrogenesis of the AVB lavas from geochemical considerations*

360   Modification by fractional crystallisation does not affect long-lived radiogenic

361 Sr-Nd isotope compositions and should therefore provide information about magmatic

362 source(s). Guo et al. (2014) proposed that post-collisional K-rich magmatism in

363 northern Tibet reflects subduction-related metamorphic mélanges. They suggested

364 mélange zones on top of both the Indian and Tarim subducting slabs mix slab

365 materials and mantle rocks. These mixed mélanges are fluid-rich and thus buoyant,

366 and subsequently upwell and partially melt to high degrees. This model does not

367 include a means of focussing volcanism and explicitly involves materials with a range

368 of isotope compositions. For example, the Indian and Tarim slabs are isotopically

369 distinct (Tarim block, $^{87}Sr/^{86}Sr \approx 0.708$ and $^{143}Nd/^{144}Nd \approx 0.5119$; Indian plate,

370 $^{87}Sr/^{86}Sr > 0.713$ and $^{143}Nd/^{144}Nd < 0.5122$; Guo et al., 2014) and a wide range of

371 radiogenic isotope variability is possible when considering the inaccessible

372 sedimentary package that these slabs may host.

373   According to the spatial classification in Guo et al. (2014), the AVB belongs to

374 the middle sub-group volcanic field, which displays variable mixing proportions of

375   the Indian and Tarim materials. The Sr-Nd isotopic signatures of the AVB lavas are

376   extremely restricted (Fig. 8, 9), and do not conform to the notion of three

377   endmember-mixtures (DMM, Tarim and Indian mélanges). Marschall and

378   Schumacher (2012) proposed a 'mélange average' to highlight trace element

379   similarities with arc volcanism. Even though the AVB lavas share similar

380   characteristics of distinct enrichment in Pb and depletion in Nb, Ta and Ti, mélange

381   abundances of LILEs and LREEs are much lower than those of the AVB lavas and the

382   slope of the REE is markedly flatter in mélange averages compared to the AVB lavas.

383   Although we cannot definitively rule out a small contribution from mélanges, they are

384   highly unlikely to dominate the chemical signature of AVB lavas.

385   Tomographic images (Fig. 2) depict a low-velocity gap between the Indian and

386   Tarim lithospheres, which may be a channel for upwelling asthenosphere (Wei et al.,

387   2015, 2016). Derivation solely from a shallow asthenospheric, MORB-like source is

388   inconsistent with isotopic (Fig. 8) and trace element (Fig. 6b) signatures of the AVB.

389   The elevated trace element concentrations of AVB lavas, and K-rich continental lavas

390   in general, require a source enriched beyond fertile primitive mantle. Given the thick,

391   enriched continental crust in the region, it may provide such a source. Volcanism

392   related to crustal melting has been reported in the Hohxil area of northern Tibet by

393   Wang et al. (2005), but the K-rich "adakitic" volcanic rocks in this area exhibit high

394   SiO_2 contents (61-67%). The major element features of the AVB lavas, such as their

395   low SiO_2 contents (Fig. 4, 5) make it unlikely that they are derived by direct partial

396   melting of continental crust. High source concentrations of trace elements can also be

397   achieved by multistage enrichments by small degree melt metasomatism of

398   underlying lithosphere. Indeed, partial melting of the metasomatised sub-continental

399   lithospheric mantle (SCLM) has often been invoked for potassic magmatism in Tibet

400   (e.g., Cooper et al., 2002; Turner et al., 1993, 1996; Williams et al., 2004) and the

401   western Mediterranean (e.g., Conticelli et al., 2009).

402   At face value, melting of the metasomatised SCLM appears to offer the best

403   explanation for the chemical characteristics of the AVB lavas. The depth of melt

404   generation can be explored using the REE and HFSE characteristics. Williams et al.

405   (2004) compared the difference between peridotite melting models of both spinel and

406   garnet facies. They suggested that partial melting of a spinel-facies

407   phlogopite-bearing peridotite source was responsible for the volcanism in north Tibet.

408   However, as the authors explained, their model cannot reproduce reasonable

409   concentrations of yttrium, a HREE. Additionally, spinel-facies sources are

410   inconsistent with the absence of Zr/Hf fractionation in our AVB samples (46.7-51.9;

411   Fig. 10). Fractionation is expected, because D_{Zr}/D_{Hf} in garnet facies is around one,

412   whilst D_{Zr}/D_{Hf} of clinopyroxene in spinel stability field is around 0.5 (Weyer et al.,

413   2003), thus different degrees of partial melting in the spinel stability field would cause

414   variable Zr/Hf ratios.

415   Trace element geochemistry can also provide clues about the mineralogy of the

416   mantle source. For example, Nb-Ta, are chemical twins, having the same valency (+5)

417   and similar ionic radii, thus, they are expected to behave congruently in both fluids

418   and melts (e.g., Dostal and Chatterjee, 2000; Kalfoun et al., 2002; Weyer et al., 2003).

419   In Fig. 10, the AVB lavas display variable Nb/Ta ratios, significantly higher than the

420   primitive mantle (PM), UCC, GLOSS and EM2 endmembers. Depleted Nb-Ta

421   concentrations and high Nb/Ta ratios of the AVB lavas may be related to Ti-rich oxide

422   minerals in the residual phases such as rutile, amphibole and phlogopite. Xenolith

423   data from Kalfoun et al. (2002) indicate that rutile may be an important host for HFSE,

424   which is consistent with the HFSE depletions (such as Nb, Ta, Zr and Hf) and low

425   Ti_2O content in the AVB lavas. Other alterative residual minerals are amphibole and

426   phlogopite, which are the only Nb- and Ta-rich silicate minerals in the upper mantle

427   (Kalfoun et al., 2002).

428   The existence of phlogopite in the upper mantle has long been demonstrated by

429   petrological studies (e.g., Kushiro et al., 1968; Sudo and Tatsumi, 1990; Wyllie and

430   Sekine, 1982). In Fig. 4, K_2O contents are high, but similar (3.48-4.47 wt%) over a

431   comparatively wide range of SiO_2 (45.91-59.48%). In combination with the relatively

432   low MgO contents of the AVB lavas, these characteristics indicate buffering of K_2O

433   by potassium-rich phases in the mantle source (i.e., phlogopite and/or amphibole).

434   Because Rb and Ba are compatible in phlogopite, whilst Rb, Sr and Ba are moderately

435   compatible in amphibole, melts in equilibrium with phlogopite have high Rb/Sr and

436   lower Ba/Rb ratios than that formed in amphibole-bearing source. As suggested by

437   Furman and Graham (1999) and Yang et al. (2004), samples with low Ba/Rb (<20)

438   and high Rb/Sr (>0.1) are generally derived from phlogopite-bearing source. In Fig.

439   11, the plot of Rb/Sr versus Ba/Rb ratios implies that the AVB lavas are derived from

440   a phlogopite-rich source.

441   The ~90 km-thick crust (Tunini et al., 2016; Wittlinger et al., 2004) suggests that

442   the SCLM is deep with high temperature and pressure. Phlogopite is more stable than

443   amphibole at these P-T conditions (Condamine and Medard, 2014; Sudo and Tatsumi,

444   1990). Therefore, we suggest that the AVB lavas are likely derived by partial melting

445   of the Tibetan phlogopite-garnet-bearing SCLM. This result is in accordance with

446   melting experiments conducted by Condamine et al. (2016), which suggest that

447   worldwide post-collisional K-rich lavas are generated by melting of metasomatised

448   phlogopite-bearing garnet-peridotite. The three oldest samples (ASKL-17, ASKL-18,

449   516-11) exhibit the most mafic major element characteristics in our dataset. Thye are

450   distinct, with Sr-Nd isotope compositions slightly less enriched than the younger lavas

451   of the dataset (Fig. 8). The oldest two samples are also geographically located to the

452   northwest of the main AVB (Fig. 1).   They likely have a slightly different

453   petrogenetic history, however, we envision a similar process of melting

454   phlogopite-garent-bearing SCLM.

455   It is an interesting question as to how such an isotopically homogeneous signature

456   can be generated from the SCLM, which should be highly heterogeneous given its

457   long-term enrichment by multiple metasomatic events. Initial metasomatic events of

458   the Tibetan lithosphere may have begun as early as the Proterozoic when the

459   Proto-Tethys lithosphere subducted beneath the basement of Tibet, and continued

460   until the Late Triassic with the closure of Paleo-Tethys Ocean (Jiang et al., 2013;

461   Mattern and Schneider, 2000; Pullen et al., 2008; Sone and Metcalfe, 2008). The

462   impacts of Tethyan oceanic lithosphere on Tibetan lithosphere have been

463   demonstrated by several studies of Cenozoic volcanism (e.g., Chung et al., 2005;

464   Ding et al., 2003).

465   Generally, Nb-Ta-Ti negative anomalies in igneous rocks are indicators of

466   subduction (Briqueu et al., 1984; Foley et al., 1999). However, contemporary

467   subduction of the Indian plate is unlikely to result in homogeneous metasomatism of

468   the entire Tibetan lithosphere (Guo et al., 2006; Turner et al., 1996; Williams et al.,

469   2004). The metasomatised upper mantle in the plateau is likely to be associated with

470   slab-derived components such as ancient subducting sediments, which may originate

471   from Proto-Tethys and Paleo-Tethys oceanic lithospheres.

472   The petrogenesis of AVB discussed above is in accordance with most shoshonites

473   of the world, such as western Alps (Conticelli et al., 2009), Fiji (Leslie et al., 2009),

474   Izu-Bonin-Mariana intra-oceanic arc (Ishizuka et al., 2010; Sun and Stern, 2001) and

475   Aeolian (Peccerillo et al., 2013). They are generally generated by partial melting of an

476   enriched mantle and display similar geochemical characteristics, including enrichment

477   in LILE and depletion in Nb-Ta-Ti. However, tectonic mechanisms for partial melting

478   of enriched mantles are diverse. The majority of shoshonitic magmatism is generated

479   in three broad tectonic settings (Gill et al., 2004): (1) post-collisional orogens, e.g.

480   Tibet (Turner et al., 1996) and the Alps (Conticelli et al., 2009); (2) rifting in

481   continental post-subduction areas and continental magmatic arcs, e.g. Cascades

482   (Conrey et al., 1997); (3) intra-ocean island arc settings, e.g. Fiji (Leslie et al., 2009)

483   and Izu-Bonin-Mariana arc (Ishikuka et al., 2010; Sun and Stern, 2001). Shoshonitic

484   rocks in the former two tectonic settings are mainly characterised by Ce/Yb ratios

485 higher than 46.5 and oceanic-arc related shoshonitic rocks display Ce/Yb ratios lower

486 than 46.5 (Gill et al., 2004). The Ce/Yb ratios of AVB lavas range from 114 to 191,

487 which is consistent with the post-collisional settings. Due to the complicated tectonic

488 setting of the AVB and the multiple possible contributors to their chemical budget

489 (e.g., subduction package of both Indian and Tarim slabs), a combined geophysical

490 and geochemical perspective may yield more insight to the triggering mechanism for

491 partial melting of the enriched SCLM.

492

493 *6.3. Integrated Geophysical and Geochemical views*

494 The historic volcanic activity in the AVB allows us to combine present day

495 seismic images of the region with geochemistry. There are four competing

496 geodynamical models for the petrogenesis of the volcanism in northwestern Tibet: (1)

497 convective removal of the lower part of lithosphere (e.g., Chung et al., 2005; Turner et

498 al., 1993, 1996; Williams et al., 2004; Xia et al., 2011); (2) mixing of Indian and

499 Tarim mélanges (Guo et al., 2014); (3) addition of subduction components without

500 mélanges (e.g., Arnaud et al., 1992; Ding et al., 2003; Tapponnier et al., 2001); and (4)

501 upwelling asthenosphere (Wei et al., 2015, 2016). Here we have the unique

502 opportunity to investigate which of these scenarios is most consistent with both

503 geochemical and geophysical constraints.

504

505 *6.3.1. Convective removal of the lower part of lithosphere*

506 Several previous studies highlight the possibility of partial melting of the SCLM

507   by convective removal of the lower part of lithosphere (e.g., Chung et al., 2005;

508   Cooper et al., 2002; Turner et al., 1993, 1996; Williams et al., 2004; Xia et al., 2011).

509   According to these geochemical studies, after the primary collision at 50-55 Ma,

510   under the squeezing pressure of the north-subducting Indian plate and the resisting

511   Tarim Block, the northwest Tibet lithosphere started to shorten and thicken to twice its

512   original thickness, accompanied by the fast uplift of the Tibetan Plateau. Subsequently,

513   owing to gravitational instability and asthenospheric heating, the lower part of the

514   Tibetan lithosphere was convectively removed. Thinning of the lithosphere led to the

515   upwelling of asthenosphere, which caused partial melting of the Tibetan SCLM. An

516   important piece of evidence for this model is that no young volcanism has been found

517   in southern Tibet after the southern Tibetan lithosphere was underthrusted by the cold

518   Indian Plate at ~10 Ma, which shut down the asthenospheric source (Chung et al.,

519   2005; Xia et al., 2011).

520   This model can explain the petrogenesis of the K-rich volcanic rocks and their

521   trace element enrichment. However, it requires the existence of K-rich layer in the

522   Tibetan lithosphere. Even if it is true, it is still difficult to explain the coincidence of

523   wide exposure of K-rich volcanism on the whole Tibetan Plateau, which indicates that

524   volcanism only happens when the lithosphere is thinned to this K-rich layer. In

525   addition, if the magmatism of AVB is associated with removal of the lower part of

526   lithosphere, the sinking of the delaminated cold lithosphere would decrease the

527   temperature of the asthenosphere beneath the AVB. However, tomographic images

528   (Fig. 2) in this region do not show a fast seismic anomaly expected for delaminated

529   lithosphere, although such a feature could be below the resolution of the seismic

530   models (~50 km, Wei et al., 2016).

531

532   *6.3.2. Mélange melting*

533   As discussed, the mélange model proposed by Guo et al. (2014) is inconsistent

534   with the radiogenic isotopic homogeneity we document in AVB lavas. The

535   mélange-melting model is also inconsistent with the following geophysical

536   observations: (1) Recent seismic results do not support the steep subduction of Indian

537   plate beneath the AVB (Fig. 2; Tunini et al., 2016; Wei et al., 2015, 2016; Zhao et al.,

538   2010). Steep subduction of the Indian plate is only observed on the west

539   (Pamir-Hindu Kush region; Negredo et al., 2007) and east of Tarim basin (south of

540   Qaidam Basin; Li et al., 2008; Zhao et al., 2010). (2) Because the Indian plate has

541   subducted more than 600 km beneath the Tibet, it is unclear how the Indian plate is

542   able to dehydrate and specifically focus volcanism in the AVB.

543

544   *6.3.3. Subduction components without mélange assemblages*

545   From south to north, the Tibetan Plateau consists of four roughly

546   east-west-trending terranes: Lhasa, Qiangtang, Songpan-Ganzi and Kunlun (Dewey,

547   1988). Tectonic studies suggest that volcanism on the Tibetan Plateau may be

548   associated with the subduction of Indian and Asian plates (e.g., Arnaud et al., 1992;

549   Ding et al., 2003; Tapponnier et al., 2001). Specifically, in northwestern Tibet, the

550   AVB lavas may be generated by the subduction of Indian and Tarim slabs. This

551 subduction model is supported by field observations, such as the East-west-trending

552 K-rich volcanic belts.

553   However, Cenozoic volcanism on the Tibetan Plateau does not display a pattern

554 of stepwise subduction of continental lithosphere (e.g., Guo et al., 2006; Williams et

555 al., 2004). Moreover, this large-scale subduction model is inconsistent with recent

556 seismic studies. For example, large-scale subduction of Asian lithospheres would

557 result in an uneven base to the Tibetan lithosphere, which is not observed (Fig. 2; Wei

558 et al., 2015, 2016; Zhao et al., 2010). Though the existence of the subducting Tarim

559 Block beneath northwestern Tibet is well-documented (e.g., Wei et al., 2015, 2016;

560 Wittlinger et al., 2004, Zhao et al., 2010), the impacts of Tarim Block on northwestern

561 magmatism may be minor, as plate subduction would cause an increase in age of the

562 northwestern magmatism from north to south, which are not observed (Guo et al.,

563 2006, 2014).

564

565 *6.3.4. Asthenosphere Upwelling and Shear Heating*

566   Wei et al. (2015; 2016) performed high resolution (1°) P-wave velocity

567 tomography beneath the AVB. They revealed a mantle lithospheric between the Indian

568 and Tarim slabs directly beneath the AVB, and suggested that the volcanism is

569 associated with the upwelling of the asthenosphere. These models show that in

570 northwestern Tibet, the Indian and Tarim LABs are visible in seismic tomography

571 with fast seismic velocities extending to depths of approximately 250 km (Figure 2),

572 which is in accordance with the results of Zhao et al. (2010). The upper boundaries

573   between the subducting slabs and Tibetan lithosphere are not observed, suggesting

574   that the lithosphere of the subducting slabs and the Tibetan lithosphere may have

575   integrated into one thick lithosphere.

576   However, some lateral variations exist in lithospheric structure. The continuous

577   fast seismic velocities in the west vertical cross section (80°E) of P-wave tomography

578   indicates that the Indian and Tarim slabs have collided with each other. However, the

579   middle vertical cross section (81.58°E, on the AVB) show discontinuous fast seismic

580   velocities suggesting that the north edge of Indian plate has not yet collided with the

581   Tarim Block, thus there exists a gap ~120 km wide between the two slabs directly

582   beneath the AVB; the east vertical cross section (83°E) shows a larger gap between

583   those two slabs. This discontinuous lithospheric structure is supported by a recent

584   joint analysis of seismic velocity, gravity and topography in the Tarim Basin and

585   surrounding region (Deng et al., 2017) showing that the AVB is underlain by weak,

586   low-density material compared to the thick, dense and strong Tarim block to the north.

587   Owing to the limited number of seismic stations in the sparsely populated region, it is

588   hard to constrain the details of this, but Deng et al. (2017) suggest that the mantle in

589   this region is anomalously buoyant. They argue that the region is too narrow to be

590   caused by a thermal anomaly and suggest a compositional anomaly (e.g.,

591   metamorphism linked to subduction fluids) or a localised heat source (e.g.,

592   shear-heating) are likely mechanisms to weaken the lithosphere.

593   Global positioning system measurements indicate that the convergence rate

594   between India and Tarim Basin in the direction of relative motion between India and

595   Eurasia (~N20°E) (Sella et al., 2002) is ~28 mm/year (Zhang et al., 2004). As a result,

596   the upper mantle in the gap between the Indian and Tarim lithospheres is being

597   squeezed by the relative motion of the two subducting lithospheres. Since we have

598   excluded the models of convective removal of the SCLM, subduction and mélange

599   melting, we speculate that partial melting of the SCLM is likely related to shear

600   heating and/or upwelling of asthenosphere in the upper mantle. Since a significant

601   chemical signature of asthenosphere has not been detected in this study, we suggest

602   that the heat contribution of shear heating may be dominant.

603   Convergence of lithospheres may result in viscous strain-rate melting in the upper

604   mantle (Kincaid and Silver, 1996), which can provide the heat for partial melting of

605   the SCLM without removing of the lower part of the SCLM. Furthermore, shear

606   heating can cause very localised heating in the lithosphere close to large faults

607   (Leloup et al. 1999) providing a possible explanation for the near invariant radiogenic

608   isotope signature derived from the SCLM, which should be heterogeneous on a large

609   scale. The duration of this viscous heating is thought to be short, spanning a period of

610   ~12-15 Ma (Kincaid and Silver, 1996), which is consistent with the age of

611   northwestern volcanism (8.27 Ma to present; Cooper et al., 2002; Guo et al., 2016,

612   2014; Turner et al., 1993; 1996; Williams et al., 2004).

613   Shear heating has been invoked as a meachanism for crustal magmatism, with

614   recent models suggesting that shear heating coupled with the insulating nature of hot

615   rocks provides an efficient mechanism for melting in orogenic belts (e.g., Whittington

616   et al., 2009). For example, leucogranites on an active thrust in the Himalayas have

617   been proposed to result from partial melting triggered by shear-heating (Harrison et al.,

618   1997, 1998) as well as in other orogens (e.g., Trans-Hudson, Appalachain, Nabalek

619   and Liu, 2004). In the AVB, the Altyn Tagh fault extends down to a depth of 140 km

620   with strike-slip shear in the SCLM (Wittlinger et al., 1998), thus it is possible that

621   shear heating in the SCLM precedes shoshonitic melt generation. Once melts are

622   formed, the deep-seated lithospheric faults may allow melt to migrate to the Earth

623   surface, resulting in minimal crustal contamination of the primary magma.

624   If our model that the northwestern volcanism is related to the mantle gap between

625   the two subducting slabs is correct, the closure of the gap in the west prior to the east

626   should result in older lavas in the west. As suggested by seismic images (Fig. 2), there

627   must be a critical point between 80°E and 81.56°E, where the Indian slab has just

628   collided with the Tarim slab. Tomographic results based on the model of Wei et al.,

629   (2016) suggest that this critical point is located at ~81°E. To the west of 81°E, the gap

630   between the two slabs closed prior to the east, thus the volcanism related to the gap

631   are chronologically older than the volcanism in the east. This chronological difference

632   is in accordance with the ages of northwestern volcanism summarised by Guo et al.

633   (2014). To the west of 81°E, the average ages of Tianshuihai, Quanshuigou,

634   Qitaidaban, Dahongliutan and Kangxiwa lavas are 5.20, 5.23, 8.27, 3.78, 3.24 Ma,

635   respectively (Fig. 12). In the east of 81°E, the average age of Ashikule, Keliya, Pulu

636   and Heishibei are 1.07, 0.56, 1.20 and 1.28 Ma, respectively, which are significantly

637   younger than those of the western volcanic fields (Fig. 12). Additionally, as strong

638   tectonic activities are usually manifested geomorphologically, we speculate that the

639 formation of arc-shape Kegang fault may also be related to the closure of the

640 lithospheric gap (Fig. 1, Fig. 12).

641

642 ## 7. Summary and Outlook

643 New geochemical analyses of the most recent volcanism in Tibet are combined

644 with recent tomographic studies beneath the AVB at the northwestern margin of the

645 Tibetan Plateau. We suggest that AVB lavas are most likely generated by partial

646 melting of the Tibetan phlogopite-bearing SCLM within the garnet stability field.

647 While the resolution of the seismic models alone cannot distinguish the mechanism of

648 heat generation, the homogeneous characteristics of the radiogenic isotope signatures

649 point to a very localised source. This argues against a broad asthenospheric upwelling

650 or direct link with subduction-derived fluids or chemically heterogeneous mélange

651 contributions. Rather, we suggest heat is provided via localised shear-heating along

652 lithospheric faults accommodating the closure of the Indian and Tarim slabs. This

653 mechanism, as well as explaining the seismic and radiogenic isotope signatures,

654 provides a viable means of transport to account for the lack of crustal contamination

655 in the samples. This model suggests that evolving volcanism in Tibet, and possibly in

656 other transpressive regimes, are more directly linked to the evolution of large-scale

657 lithospheric faults as opposed to the subduction of the Indian and Tarim slabs. More

658 precise dating of tectonic and volcanic activity, more concentrated geochemical study

659 and improved geophysical imaging are required to fully test this hypothesis.

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1016 Figure Captions

1017 Fig. 1. (a) Tectonic map of the northwest margin of the Tibetan Plateau, modified

1018 after Furuya and Yasuda (2011), Xu et al. (2013) and Bie and Ryder (2014). Black,

1019 green and red focal mechanisms represent the main shocks of the 2008, 2012 and

1020 2014 earthquakes, respectively, according to the Global Centroid-Moment-Tensor

1021 (CMT) catalog. (b) Geological map of the Ashikule volcanic basin (AVB), modified

1022 after Liu and Maimaiti (1989) and Xu et al. (2012). Red arrows represent the direction

1023 of the lava flow. Black open circles represent the location of lava samples, and the ID

1024 of samples is also marked with a black arrow. Locations of main volcanoes in the

1025 AVB are marked with red triangles. AS: Ashi Volcano; DH: Dahei Volcano; XS:

1026 Xishan Volcano; MT: Mati Volcano; DS: Dong volcano; HL: Heilong Volcano; MN:

1027 Maoniu Volcano; YY: Yueya Volcano; MG: Migong Volcano; WL: Wuluke Volcano;

1028 YZ: Yizi Volcano; GT: Gaotai Volcano; BH: Binhu Volcano; YS: Yin Volcano.

1029 Fig. 2. Vertical cross sections of P-wave anisotropy tomography (isotropic velocities

1030 only are shown) along 3 profiles shown on the inset map (see Wei et al., 2016 for

1031 details of the method). (a) Vertical cross section of 80°E suggests that the Indian slab

1032 has collided with the Tarim slab. (b) Vertical cross section of 81.53°E (on the AVB)

1033 suggests that there is a small gap between the two slab beneath the AVB. (c) Vertical

1034 cross section of 83°E displays a wider gap between the two slabs.

1035 Fig. 3. Photomicrographs of four representative volcanic rocks in the AVB under

1036 cross-polarised light. Cpx, clinopyroxene; Opx, orthopyroxene; Phl, phlogopite; Pl,

1037 plagioclase. (a) Sample ASKL-5. Phenocrysts of orthopyroxene and plagioclase

1038 showing disequilibrium textures including extensive embayments. (b) Sample 513-11.

1039 Phenocrysts of clinopyroxene, plagioclase and phlogopite in a highly vesicular matrix.

1040 (c) Sample ASKL-12. A large plagioclase glomerocryst in a crystalline matrix. (d)

1041 Sample 518-9. Phenocrysts of clinopyroxene, plagioclase and orthopyroxene in a

1042 porphyritic texture.

1043 Fig. 4. Classification for lava samples in the AVB: (a) SiO_2 versus $Na_2O + K_2O$ (wt%),

1044 (b) Na_2O versus K_2O (wt%). The boundaries of diagram (a) and (b) are from Le Bas

1045 et al. (1986) and Miller et al. (1999), respectively.

1046 Fig. 5. Major element variation diagrams for the studied samples in AVB. Symbols

1047 are as in Fig. 2. The most recent lava sample (ASKL-3) is marked with a cross inside

1048 the circle.

1049 Fig. 6. Primitive mantle-normalized trace element (a) and chondrite-normalized rare

1050 earth element patterns (b) for AVB lavas. Primitive mantle and chondrite

1051 normalization values are from Sun and McDonough (1989). EMII (Samoa) (Jackson

1052 et al., 2007), globally subducting sediment (GLOSS; Plank and Langmuir, 1998) and

1053 upper continental crust (UCC, Rudnick and Gao, 2003) compositions are presented

1054 for comparison.

1055 Fig. 7. (a) Primitive mantle-normalized La/Yb ratios ($(La/Yb)_n$) versus MgO (wt%);

1056 (b) $(La/Yb)n$ versus La (ppm). (c) Eu anomaly (Eu/Eu*) versus MgO (wt%); Eu/Eu*

1057 $= Eu_n \times 2/ (Sm_n + Gd_n)$; Eu_n, Sm_n and Gd_n are the primitive mantle-normalized Eu,

1058 Sm and Gd concentrations of studied sample, respectively. (d) Ti anomaly (Ti/Ti^*)

1059 versus SiO_2 (wt%); $Ti/Ti^* = Ti_n \times 2/ (Eu_n + Tb_n)$; Ti_n and Tb_n are the primitive mantle

1060 –normalized Ti and Tb, respectively. (e) Ba/Th ratios versus Th/Nd ratios. (f) Th/La

1061 ratios versus Sm/La ratios. Symbols are as in Fig. 2.

1062 **Fig. 8.** Present day $^{87}Sr/^{86}Sr$ versus $^{143}Nd/^{144}Nd$ for lavas from the AVB. Literature

1063 data specifically from the AVB are presented for comparison (Cooper et al., 2002;

1064 Guo et al., 2006, 2014; Williams et al., 2004). Enriched mantles (EM I, EM II), bulk

1065 silicate earth (BSE), Indian and Pacific MORB fields are modified from Zindler and
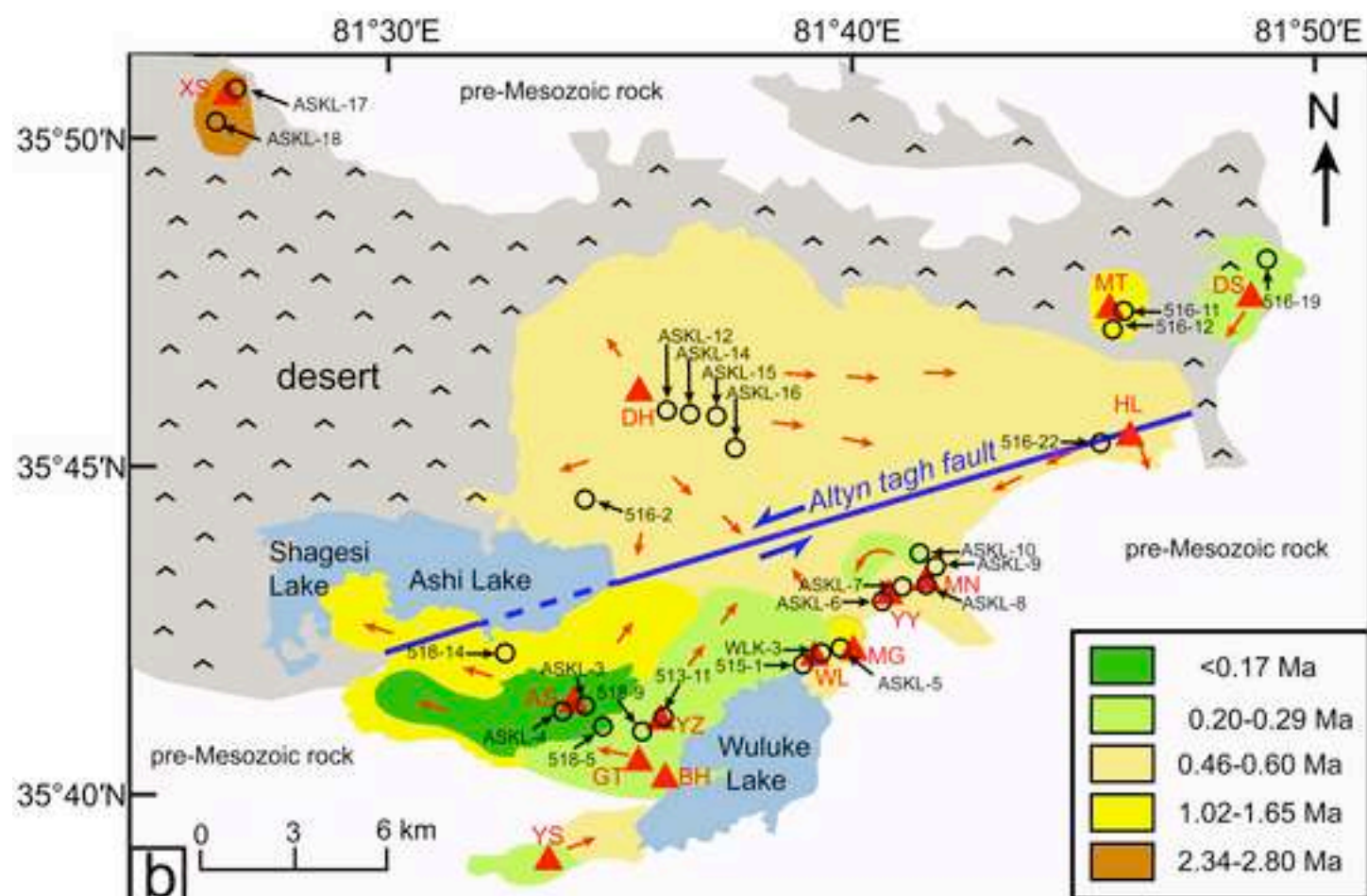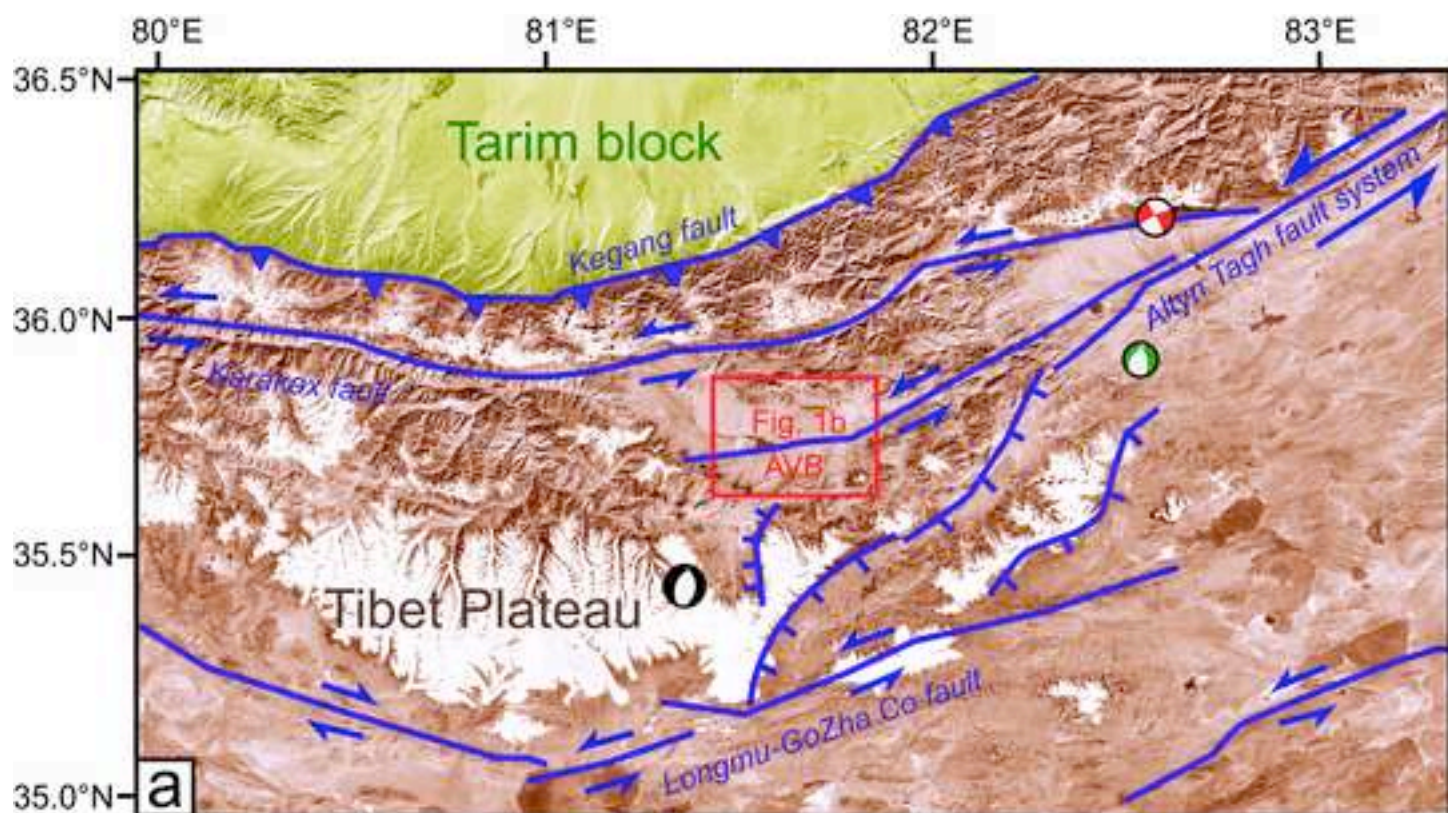
1066 Hard (1986) and White (2015).

1067 **Fig. 9.** SiO_2 versus $^{87}Sr/^{86}Sr$ (a) and $^{143}Nd/^{144}Nd$ ratios (b) for the AVB lavas. Symbols
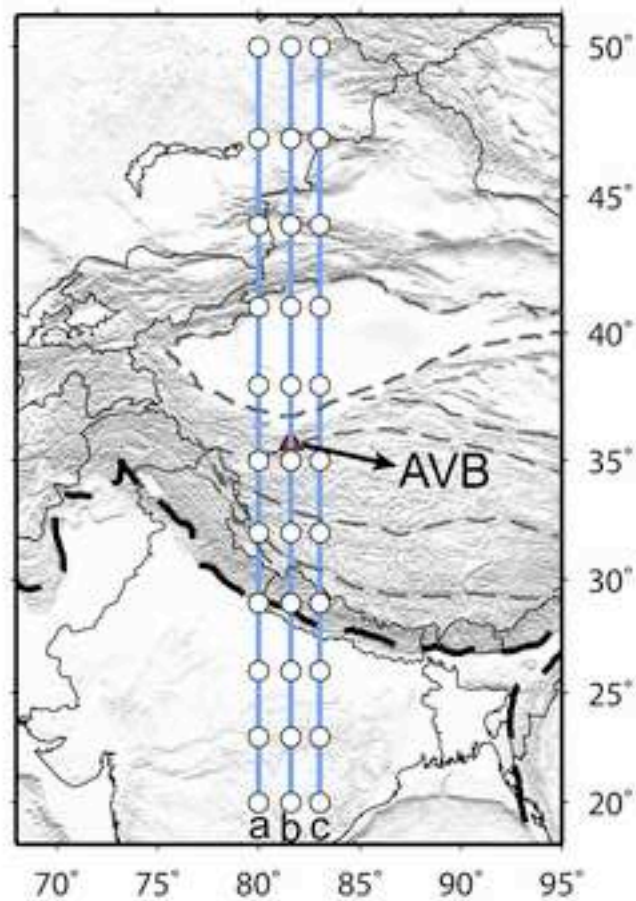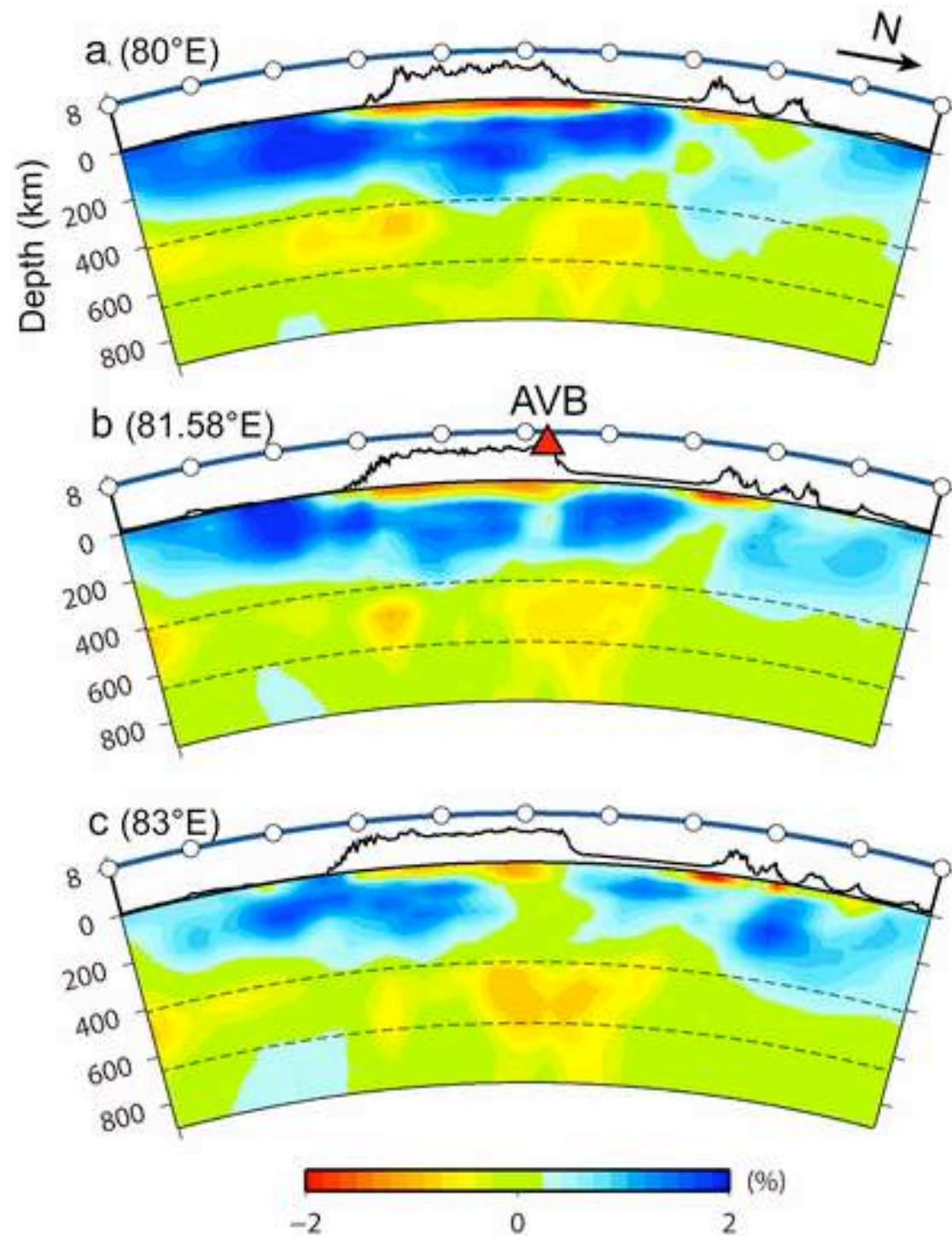
1068 as in Figure 2.

1069 **Fig. 10.** Zr/Hf versus Nb/Ta ratios of the AVB lavas. Plots of upper continental crust

1070 (UCC; Rudnick and Gao, 2003), primitive mantle (PM; Sun and McDonough, 1989),

1071 global subducting sediments (GLOSS; Plank and Langmuir, 1998) and EM II Samoan

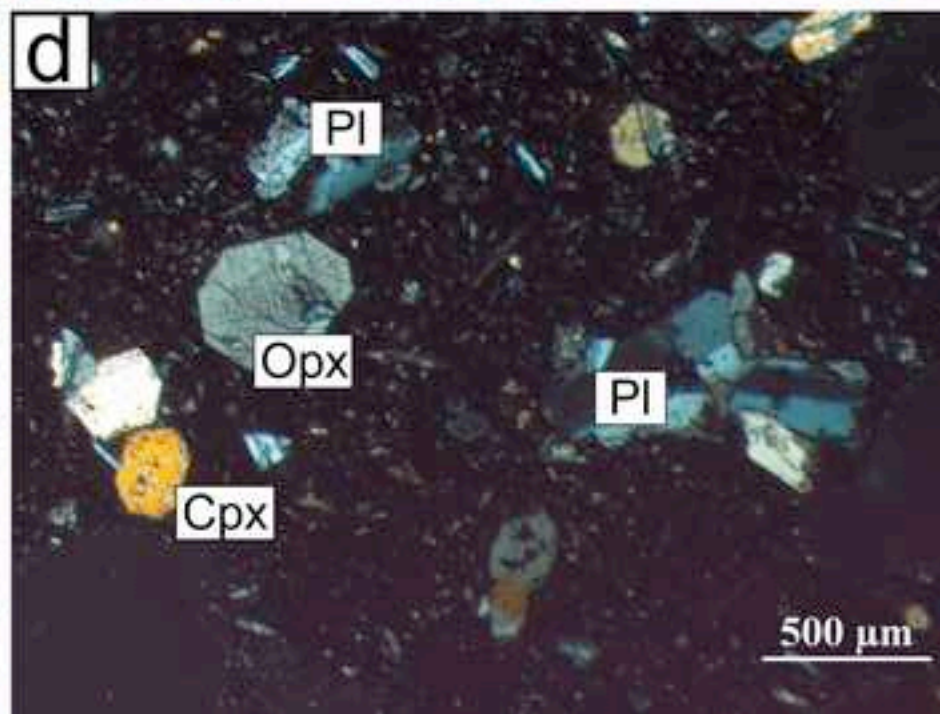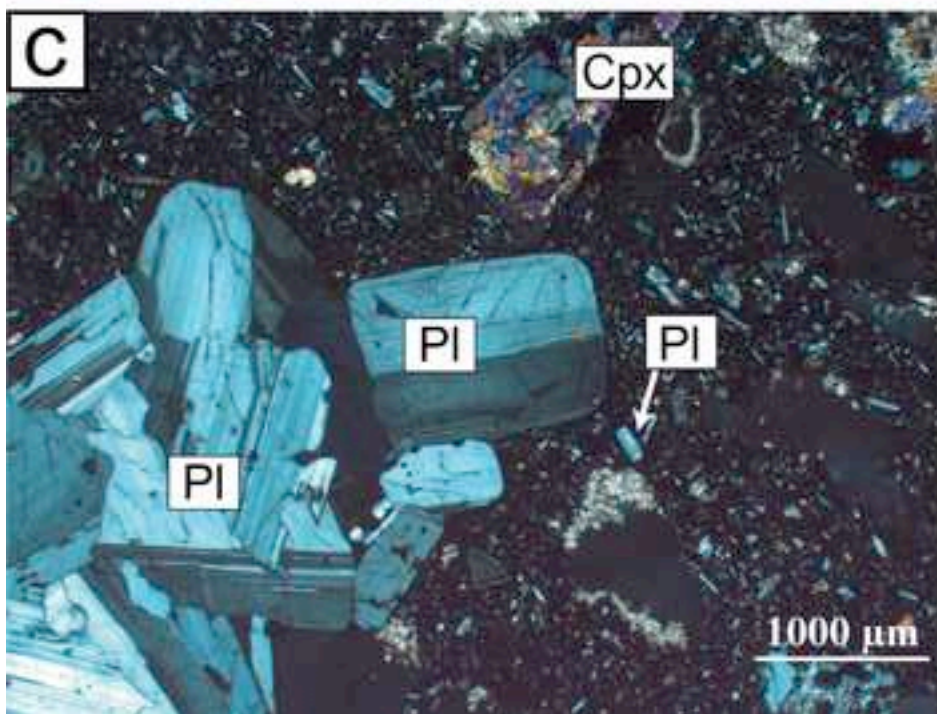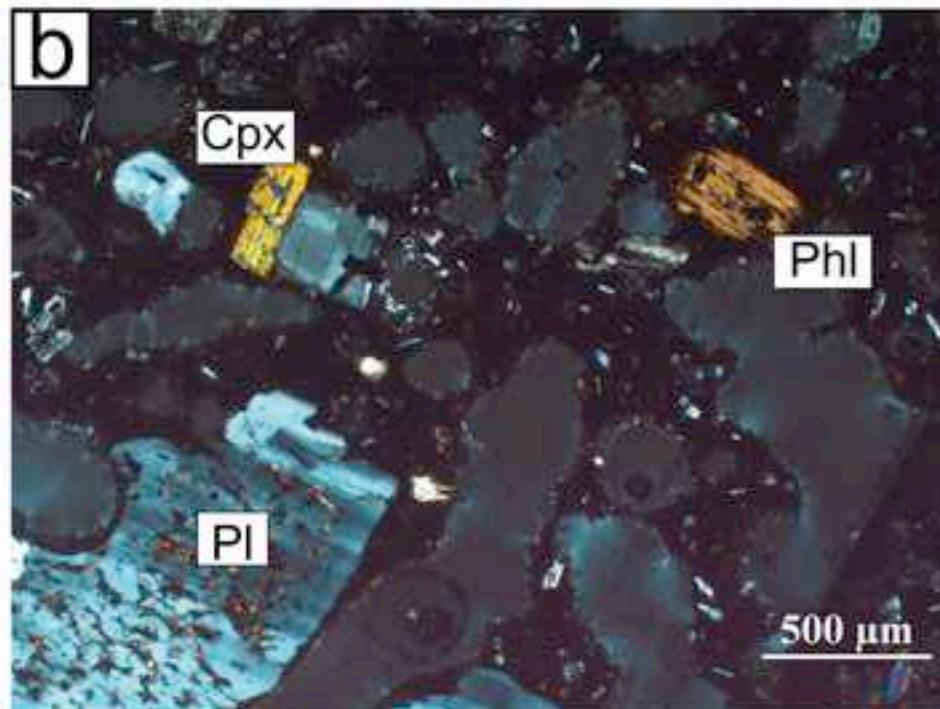1072 source (Jackson et al., 2007) are displayed for comparison.

1073 **Fig. 11.** Rb/Sr versus Ba/Rb for the AVB lavas (after Furman and Graham, 1999). The

1074 field of common lithospheric mantle (CLM) is taken from Furman and Graham

1075 (1999).

1076 **Fig. 12.** Schematic model for the post-collisional magmatism in northwestern Tibet

1077 and lithospheric sections of the present settings. The locations and average ages of the

1078 volcanic fields are modified from Guo et al. (2014). In the section of 79°E, Indian and

1079   Tarim lithospheres have collided together beneath the west Kunlun Mountains. In the

1080   section of 81°E, the two slabs just collide together. In the section of 83°E, there is a

1081   gap between the Indian and Tarim lithospheres. The progressive closure of this gap

1082   may lead to shear heating, melting of the Tibetan SCLM and focussing of magmatism

1083   with transport along deep-seated lithospheric faults.

Left panel (block diagram):

79°E   81°E   83°E

Kegang fault

Tarim Basin

N

1.20 Ma

3.24 Ma

36°N

3.78 Ma   8.27 Ma   Tibetan Plateau   AVB   1.07 Ma

5.23 Ma   0.56 Ma   1.28 Ma

35°N

5.20 Ma

Tarim block

Indian plate

Right panel (cross-sections):

N

79°E

Tibetan lithosphere

Indian plate   Tarim block

81°E

Tibetan lithosphere

Indian plate   Tarim block

shear heating zone

83°E

Tibetan lithosphere

Indian plate   Tarim block

shear heating zone

Table 1

Percentage abundance of vesicles, groundmass and phenocrysts in Ashikule volcanic rocks.

Episode (age)	Sample	Vesicle	Groundmass	Phenocrysts	Groundmass
1 and 2 (<0.17 Ma)	ASKL-3	60%	28%	Pl(6)+Cpx(3)+Opx(2)+Ol(1)	G
1 and 2 (<0.17 Ma)	ASKL-4	55%	32%	Pl(1)+Cpx(4)+Opx(6)+Ol(1)+Bi(<1)+Fe-Ti(<1)	Pl+Ol+G
1 and 2 (<0.17 Ma)	518-5	35%	51%	Opx(9)+Cpx(4)+Pl(1)	Pl+Cpx+Opx+G
3 (0.20-0.29 Ma)	ASKL-10	30%	60%	Pl(8)+Opx(1)+Fe-Ti(1)	Pl+Opx+Fe-Ti+G
3 (0.20-0.29 Ma)	513-11	45%	40%	Pl(10))+Cpx(2)+Opx(2)+Phl(1)	Pl+Cpx+Opx+G
3 (0.20-0.29 Ma)	516-19	20%	74%	Pl(4)+Opx(1)+Cpx(1)	Pl+Ol+Cpx+Opx+G
3 (0.20-0.29 Ma)	518-9	45%	39%	Pl(10)+Cpx(3)+Opx(2)+ Fe-Ti(1)	Pl+Cpx+Opx+Fe-Ti+G
4 (0.46-0.60 Ma)	ASKL-6	80%	15%	Pl(<1)+Cpx(3) +Opx(<1)+Ol(<1)+Fe-Ti(<1)	Fe-Ti+G
4 (0.46-0.60 Ma)	ASKL-7	20%	65%	Pl(12)+ Opx(2)+Cpx(<1)	G
4 (0.46-0.60 Ma)	ASKL-8	25%	42%	Pl(25)+ Opx(4)+Cpx(2)+ Fe-Ti(2)	Pl+Opx+Fe-Ti+G
4 (0.46-0.60 Ma)	ASKL-9	40%	38%	Pl(16) + Opx(5)+Cpx(1)+ Fe-Ti(<1)	Pl+Opx+Ol+Fe-Ti+G
4 (0.46-0.60 Ma)	ASKL-12	10%	58%	Pl(22)+ Opx(4)+Cpx(3)+Phl(1)+Fe-Ti(2)	Pl+OPX+CPX+Ol+G
4 (0.46-0.60 Ma)	ASKL-14	25%	66%	Pl(5)+Opx(4)	Pl+Opx+G
4 (0.46-0.60 Ma)	ASKL-15	30%	52%	Pl(7)+Cpx(6)+Opx(5)	Pl+Ol+Cpx+Opx+Fe-Ti+G
4 (0.46-0.60 Ma)	ASKL-16	15%	58%	Pl(15)+Opx(8)+Cpx(2)+Phl(1)+Fe-Ti(1)	Pl+Opx+Cpx+G
4 (0.46-0.60 Ma)	515-1	15%	68%	Pl(12)+Opx(2)+Phl(2)+Cpx(1)	Pl+Phl+Cpx+Opx+G
4 (0.46-0.60 Ma)	WLK-3	30%	61%	Pl(6)+Opx(2)+Ol(<1)+Fe-Ti(<1)	Pl+Cpx+Opx+Fe-Ti+G
4 (0.46-0.60 Ma)	516-2	25%	56%	Cpx(11)+Opx(6)+Pl(2)	Pl+Cpx+Opx+G
4 (0.46-0.60 Ma)	516-22	40%	57%	Pl(2)+ Cpx(2)+Opx(1)	Pl+Cpx+G
5 (1.02-1.65 Ma)	ASKL-5	50%	34%	Pl(11)+ Opx(2)+Fe-Ti(2)+Cpx(1)	Pl+Cpx+G
5 (1.02-1.65 Ma)	516-11	40%	48%	Opx(6)+Cpx(3)+Pl(2)+Fe-Ti(1)	Pl+Cpx+Opx+G
5 (1.02-1.65 Ma)	516-12	50%	39%	Opx(6)+ Pl(4)+Cpx(1)	Pl+Cpx+G
5 (1.02-1.65 Ma)	518-14	30%	64%	Pl(2)+Opx(2)+Cpx(1)+Fe-Ti(1)	Pl+Cpx+Opx+Fe-Ti+G
6 (2.34-2.80 Ma)	ASKL-17	30%	57%	Cpx(9)+Opx(4)	Pl+Cpx+Opx+G
6 (2.34-2.80 Ma)	ASKL-18	30%	55%	Cpx(12)+Opx(2)+Pl(1)	Pl+Cpx+Opx+Fe-Ti+G

Pl = plagioclase; Cpx = clinopyroxene; Opx = orthopyroxene; Phl = phlogopite; Fe-Ti = Fe-Ti oxides; G = glass. Numbers in the brackets after the phenocrysts indicate the visually estimated volumetric percentages of the different phenocryst types in the thin section.

Table 2

Whole-rock trace and major element results for volcanic rocks from Ashikule Volcanoes.

Episode (age)	1 and 2 (<0.17 Ma)			3 (0.20-0.29 Ma)				4 (0.46-0.60 Ma)					
Sample	ASKL-3	ASKL-4	518-5	ASKL-10	513-11	516-19	518-9	ASKL-6	ASKL-7	ASKL-8	ASKL-9	ASKL-12	ASKL-14
(wt %)													
SiO_2	54.11	54.52	54.67	55.01	58.95	55.9	57.12	54.33	53.78	57.07	54.82	58.76	54.07
TiO_2	1.83	1.98	2.01	2.21	1.62	2.18	1.89	2.05	2.15	2.12	2.17	1.36	1.87
Al_2O_3	14.57	14.43	14.49	14.59	14.97	14.77	14.65	14.78	14.44	14.89	14.67	14.19	14.16
$Fe_2O_3(T)$	7.75	7.66	7.75	7.78	6.67	8.45	7.28	7.46	7.57	7.52	7.65	6.12	8.11
MnO	0.11	0.11	0.11	0.11	0.09	0.12	0.1	0.10	0.11	0.10	0.11	0.09	0.12
MgO	5.31	4.12	3.99	3.03	3.35	4.10	3.47	2.96	3.44	2.86	3.14	3.16	4.83
CaO	5.46	6.56	6.62	6.39	5.24	6.50	5.63	5.88	6.26	5.71	6.16	5.97	7.20
Na_2O	3.18	3.08	3.12	3.25	3.4	3.11	3.14	3.05	3.19	3.27	3.24	2.88	2.92
K_2O	3.97	4.07	4.19	4.00	4.30	3.62	4.43	3.97	3.98	4.20	3.95	4.41	3.68
P_2O_5	0.99	1.10	1.14	1.23	0.75	1.15	1.01	1.10	1.19	1.15	1.19	0.69	1.01
LOI	2.91	1.48	0.96	1.45	1.27	0.60	1.01	3.70	2.86	1.03	2.68	2.99	1.20
Total	100.19	99.11	99.05	99.05	100.61	100.5	99.93	99.38	98.97	99.92	99.78	100.6	99.2
Mg#	61	55	54	47	53	52	52	47	51	46	48	54	57
(ppm)													
Li	41.2	33.4	29.0	33.6	61.6	26.2	37.9	31.7	30.8	27.1	23.7	34.5	26.5
Sc	14.5	6.31	9.74	10.9	11.0	8.97	11.2	11.0	8.06	8.92	10.7	10.7	9.42
V	105	108	107	98.8	92.2	113	97.9	98.9	74.0	78.7	98.5	78.4	115
Cr	78.5	66.1	66.2	24.6	49.2	71.5	52.9	26.3	14.8	19.0	22.9	75.3	104
Co	20.0	19.7	19.3	16.6	17.5	20.6	17.5	16.4	12.0	12.5	16.9	13.5	22.9
Ni	49.3	43.9	45.8	13.7	41.2	43.7	35.0	13.7	10.4	9.94	14.6	24.0	58.5
Cu	19.8	20.8	20.0	18.8	26.4	25.3	20.9	10.2	17.7	13.7	16.9	14.3	26.5
Zn	131	126	127	143	118	144	133	139	105	114	146	112	125
Rb	120	125	126	143	195	111	159	136	112	134	143	171	112
Sr	1386	1433	1432	1247	885	1012	1203	1130	874	956	1400	786	1193
Y	34.4	31.8	32.6	35.2	32.1	35.5	32.9	33.2	26.6	29.6	33.8	30.8	33.2
Zr	520	555	549	703	521	613	561	562	530	602	610	411	521
Nb	47.0	51.3	50.0	63.4	53.5	57.8	51.9	10.5	47.1	52.9	14.3	46.2	49.0
Cs	3.22	3.27	2.92	2.55	9.42	1.90	4.24	3.80	2.25	2.76	2.69	5.07	2.35
Ba	2144	2348	2322	2169	1620	1846	2205	2074	1740	1800	2136	1570	1814
La	159	169	174	191	166	158	174	175	142	165	188	150	150
Ce	305	329	336	367	311	310	335	334	273	315	362	285	289
Pr	32.5	35.9	36.4	39.0	32.8	33.0	36.0	36.5	29.4	33.2	39.2	29.8	30.7
Nd	127	138	140	147	122	127	137	132	110	126	143	111	119
Sm	19.3	21.1	21.5	22.3	18.0	19.9	21.2	20.7	16.8	18.4	22.0	16.5	18.4
Eu	4.05	4.33	4.38	4.41	3.18	4.11	4.18	4.08	3.33	3.57	4.35	2.91	3.80
Gd	11.3	12.2	12.3	12.9	10.4	12.0	12.1	12.4	9.7	10.6	13.3	9.7	11.0
Tb	1.47	1.51	1.54	1.61	1.36	1.57	1.54	1.50	1.24	1.33	1.60	1.27	1.46
Dy	6.58	6.60	6.69	7.10	6.07	7.21	6.66	6.67	5.38	5.87	6.91	5.81	6.64
Ho	1.10	1.07	1.08	1.14	1.00	1.22	1.08	1.12	0.87	0.95	1.14	0.96	1.13
Er	2.62	2.51	2.50	2.65	2.37	2.93	2.51	2.57	2.04	2.19	2.58	2.31	2.71
Yb	2.14	1.87	1.86	1.96	1.85	2.32	1.87	1.97	1.52	1.65	1.96	1.81	2.18
Lu	0.31	0.26	0.26	0.28	0.26	0.34	0.26	0.28	0.22	0.23	0.27	0.26	0.32
Hf	10.5	11.5	11.5	14.3	10.7	12.6	12.1	10.8	10.7	11.8	12.0	8.64	10.8
Ta	1.11	1.32	1.14	1.59	1.38	1.57	1.23	0.30	1.04	1.15	0.33	1.08	1.56
Tl	8.26	0.66	0.73	0.81	0.15	0.64	0.95	0.29	0.15	1.96	0.62	1.29	0.83
Pb	66.3	39.6	41.2	41.1	36.2	33.7	42.7	39.0	31.4	28.3	42.1	42.2	33.3
Th	27.6	23.2	29.0	28.3	35.9	19.9	35.0	27.2	22.2	26.3	28.9	33.3	19.9
U	5.67	5.87	5.47	5.59	8.39	4.51	5.76	4.88	4.67	5.11	5.65	6.41	5.14

Table 2 *(continued)*

Episode (age)	4 (0.46-0.60 Ma)						5 (1.02-1.65 Ma)				6 (2.34-2.80 Ma)	
Sample	ASKL-15	ASKL-16	515-1	WLK-3	516-2	516-22	ASKL-5	516-11	516-12	518-14	ASKL-17	ASKL-18
(wt %)												
SiO_2	55.76	55.69	59.48	58.75	55.45	55.43	55.07	50.37	47.3	55.73	48.08	45.91
TiO_2	1.76	1.73	1.61	2.14	1.74	2.13	2.13	1.73	1.53	2.17	2.22	2.14
Al_2O_3	14.11	14.00	14.14	14.93	13.75	14.6	14.74	13.39	12.69	14.36	14.74	14.27
$Fe_2O_3(T)$	7.40	7.51	5.96	7.73	7.67	8.30	7.51	7.70	6.92	7.65	10.04	9.55
MnO	0.11	0.11	0.08	0.11	0.11	0.12	0.10	0.10	0.09	0.10	0.14	0.13
MgO	3.94	4.08	2.24	2.93	4.36	4.03	3.14	5.74	4.88	3.07	7.40	7.09
CaO	6.78	6.85	5.30	5.15	6.93	6.49	6.30	7.83	9.39	5.69	7.67	8.58
Na_2O	2.95	2.89	3.23	3.31	2.83	3.04	3.30	2.91	2.95	3.21	3.87	3.65
K_2O	3.94	3.74	4.47	4.34	3.76	3.65	4.03	3.72	3.48	4.13	4.30	3.95
P_2O_5	0.91	0.88	0.83	1.16	0.89	1.12	1.16	0.89	0.77	1.19	1.12	1.07
LOI	1.54	1.49	1.95	0.26	1.38	0.57	1.5	3.21	5.93	1.64	0.41	3.52
Total	99.20	98.97	99.29	100.81	98.87	99.48	98.98	97.59	95.93	98.94	99.99	99.86
Mg#	55	55	46	46	56	52	48	63	61	47	62	63
(ppm)												
Li	29.7	33.6	65.3	45.6	33.1	29.3	30.1	21.1	26.0	35.5	18.6	15.6
Sc	8.80	11.1	8.37	10.4	13.6	13.0	10.0	12.8	11.7	9.83	5.04	13.4
V	101	104	69.2	92.5	109	110	94.5	104	82.7	95.0	146	141
Cr	73.7	88.9	23.8	32.5	95.6	71.2	19.9	142	127	37.2	141	132
Co	20.6	19.2	11.8	16.3	20.2	20.1	15.3	26.5	23.7	16.2	36.9	35.6
Ni	57.7	38.4	11.5	22.8	51.5	46.4	10.4	122	118	21.2	122	107
Cu	23.8	19.3	15.5	22.1	23.6	25.9	15.4	24.5	41.8	18.1	32.7	31.0
Zn	121	125	128	146	125	167	137	112	95.9	146	111	109
Rb	138	139	214	151	134	112	142	104	106	140	95.0	92.8
Sr	1853	945	821	954	911	1025	1499	1451	1334	1062	1411	1657
Y	32.7	34.0	30.1	33.9	33.5	35.7	32.0	27.0	26.7	33.1	23.4	25.7
Zr	549	494	518	648	510	586	588	402	419	669	355	302
Nb	50.5	49.8	58.7	63.9	48.9	54.8	21.1	46.3	42.8	64.3	58.8	45.0
Cs	3.40	3.49	10.6	4.28	3.38	1.92	3.09	2.24	2.41	3.45	3.57	3.40
Ba	1813	1656	1519	1816	1619	1787	2115	1753	1590	1956	1590	1540
La	153	156	159	174	150	155	183	123	113	169	88.3	85.1
Ce	294	301	309	339	290	300	352	236	217	334	174	166
Pr	31.6	31.9	32.6	35.9	30.7	32.2	37.8	25.7	23.4	35.7	19.5	19.1
Nd	119	122	121	138	118	123	139	97.4	89.7	137	79.5	73.5
Sm	18.2	18.5	18.7	21.2	18.2	19.3	21.2	15.4	14.0	21.4	13.2	13.0
Eu	3.53	3.48	3.18	4.07	3.49	3.96	4.12	3.37	3.08	4.20	3.40	3.25
Gd	10.8	11.0	10.8	12.3	10.9	11.7	12.6	9.76	8.54	12.5	8.63	8.85
Tb	1.41	1.44	1.39	1.58	1.46	1.52	1.52	1.21	1.11	1.59	1.14	1.11
Dy	6.32	6.65	6.13	6.97	6.59	7.01	6.69	5.53	5.09	6.92	5.38	5.14
Ho	1.06	1.12	0.98	1.12	1.11	1.19	1.11	0.95	0.85	1.11	0.89	0.87
Er	2.57	2.66	2.30	2.56	2.66	2.87	2.50	2.16	2.03	2.55	2.05	1.99
Yb	1.99	2.14	1.73	1.92	2.13	2.27	1.91	1.71	1.59	1.88	1.50	1.45
Lu	0.29	0.31	0.24	0.27	0.31	0.33	0.27	0.24	0.23	0.26	0.21	0.20
Hf	11.0	10.5	11.1	13.3	10.5	12.2	11.6	7.21	8.26	13.7	7.31	6.24
Ta	1.13	1.13	1.55	1.49	1.12	1.21	0.69	1.89	0.95	1.54	1.59	1.25
Tl	0.76	0.76	1.25	0.62	0.78	0.60	0.73	0.42	0.64	0.96	0.29	0.28
Pb	35.9	35.8	42.7	37.0	34.2	32.3	40.7	20.9	23.5	41.2	20.1	20.3
Th	23.7	27.1	41.0	29.3	27.9	22.5	28.6	15.5	16.8	25.9	10.2	13.4
U	6.09	5.29	10.73	5.98	5.51	4.51	5.69	3.04	3.36	5.45	4.29	5.33

Table 3

Present day Sr and Nd isotopic compositions of volcanic rocks from Ashikule Volcanoes.

Episode (age)	Sample	^{87}Sr/^{86}Sr	Error (2σ)	^{143}Nd/^{144}Nd	Error (2σ)	Ave. ^{143}Nd/^{144}Nd	Ave. εNd
1 and 2 (<0.17 Ma)	ASKL-3	0.710300	0.000005	0.512272	0.000011	0.512267	-7.1
				0.512262	0.000010		
1 and 2 (<0.17 Ma)	ASKL-4	0.710295	0.000005	0.512283	0.000010	0.512283	-6.8
				0.512283	0.000012		
1 and 2 (<0.17 Ma)	518-5	0.710337	0.000005	0.512290	0.000008	0.512292	-6.6
				0.512293	0.000011		
3 (0.20-0.29 Ma)	ASKL-10	0.710463	0.000005	0.512276	0.000012	0.512265	-7.1
				0.512256	0.000010		
				0.512263	0.000016		
3 (0.20-0.29 Ma)	513-11	0.710102	0.000005	0.512275	0.000010	0.512273	-7.0
				0.512271	0.000009		
3 (0.20-0.29 Ma)	518-9	0.710353	0.000005	0.512281	0.000010	0.512279	-6.8
				0.512277	0.000011		
4 (0.46-0.60 Ma)	ASKL-7	0.710516	0.000005	0.512259	0.000012	0.512254	-7.3
				0.512250	0.000012		
4 (0.46-0.60 Ma)	ASKL-8	0.710427	0.000005	0.512265	0.000011	0.512265	-7.1
				0.512264	0.000010		
4 (0.46-0.60 Ma)	ASKL-14	0.709738	0.000005	0.512277	0.000013	0.512277	-6.9
				0.512278	0.000010		
4 (0.46-0.60 Ma)	ASKL-15	0.709696	0.000005	0.512275	0.000009	0.512277	-6.9
				0.512280	0.000011		
4 (0.46-0.60 Ma)	ASKL-16	0.709871	0.000006	0.512295	0.000009	0.512295	-6.5
4 (0.46-0.60 Ma)	515-1	0.710016	0.000005	0.512280	0.000009	0.512280	-6.8
4 (0.46-0.60 Ma)	WLK-3	0.709887	0.000005	0.512288	0.000011	0.512280	-6.8
				0.512273	0000011		
4 (0.46-0.60 Ma)	516-2	0.709870	0.000005	0.512291	0.000011	0.512291	-6.6
5 (1.02-1.65 Ma)	ASKL-5	0.710523	0.000005	0.512265	0.000010	0.512269	-7.0
				0.512272	0.000014		
5 (1.02-1.65 Ma)	516-11	0.708887	0.000005	0.512337	0.000012	0.512336	-5.7
				0.512330	0.000010		
				0.512342	0.000014		
5 (1.02-1.65 Ma)	518-14	0.709876	0.000005	0.512291	0.000010	0.512284	-6.8
				0.512271	0.000009		
				0.512289	0.000014		
6 (2.34-2.80 Ma)	ASKL-17	0.708650	0.000005	0.512449	0.000010	0.512461	-3.3
				0.512473	0.000013		
6 (2.34-2.80 Ma)	ASKL-18	0.707490	0.000005	0.512487	0.000010	0.512472	-3.1
				0.512461	0.000011		
				0.512469	0.000011		

^{143}Nd/^{144}Nd of chondritic uniform reservoir (CHUR) (0.512630; Bouvier et al., 2008) is used to calculate εNd.