

1 **Derivation of Jurassic HIMU-like intraplate basalts from mantle transition**
2 **zone in South China: New geochemical constraints from olivine-**
3 **hosted melt inclusion**

4 **Yangming Wu^{1,2}, Feng Guo^{1*}, Xuan-Ce Wang³, Simon A. Wilde^{3,4}, Weiming Fan⁵**

5 ¹ State Key Laboratory of Isotope Geochemistry, Guangzhou Institute of Geochemistry, Chinese
6 Academy of Sciences, Guangzhou 510640, China

7 ² College of Earth Sciences, University of Chinese Academy of Sciences, Beijing 100049, China

8 ³ Research Centre for Earth System Science, Yunnan University, Kunming 650500, China

9 ⁴ The Institute for Geoscience Research (TIGeR), School of Earth and Planetary Sciences, Curtin
10 University, GPO Box U1987, Perth, WA 6845, Australia

11 ⁵ Institute of Tibet Plateau Research, Chinese Academy of Sciences, Beijing 100101, China

12 **Corresponding author: Feng Guo (guofengt@263.net)**

13 **Abstract**

14 The petrogenesis of high- μ (HIMU, $\mu = {}^{238}\text{U}/{}^{204}\text{Pb}$) basalt in a continental setting remains highly
15 controversial. Here we present a comprehensive geochemical (including major and trace
16 elements and isotopic compositions of whole-rock, olivine and melt inclusions) data on Jurassic
17 HIMU-like basalts from the interior (Antang) of South China. This study identified high- H_2O (up
18 to 5.0 wt%) olivine-hosted melt inclusions in these basalts. After stripping off the effects of
19 degassing, post-entrapment crystallization and kinetic diffusion, the primary magma of the
20 Antang basalts was estimated to contain ≥ 2.65 wt% H_2O and the corresponding mantle source
21 contains >1000 ppm H_2O , which is much higher than the water content of convecting
22 asthenosphere and lower mantle. In addition, the relative depletion of fluid-mobile elements and
23 positive Nb-Ta anomalies in bulk rocks and melt inclusions, and the distance far away from the
24 contemporaneous paleo-Pacific subduction zone, preclude an origin from a slab fluid-
25 metasomatized mantle. The strong depletion of CaO in both olivine phenocrysts and whole-rock
26 compositions further indicates that the basaltic magmas originated from an eclogitic or
27 pyroxenitic mantle source. The combined geochemical data suggest that the Antang basalts were
28 most likely derived from a hydrous source, which was composed mainly of ancient recycled
29 oceanic crust that had stagnated in the mantle transition zone (MTZ) for more than one billion
30 years. Our results therefore provide a new perspective on the MTZ origin of intracontinental
31 basalts and imply that the hydrous MTZ can preserve recycled crustal components for long
32 periods of time and form distinct mantle reservoirs for intraplate basalts.

33 **Keywords:** mantle transition zone; hydrous; melt inclusion; HIMU basalts; South China

34 **1. Introduction**

35 High- μ (HIMU, $\mu = {}^{238}\text{U}/{}^{204}\text{Pb}$) basalt refers to a group of oceanic island mafic lavas with
36 very high ${}^{206}\text{Pb}/{}^{204}\text{Pb}$ ratios, coupled with low ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ and moderate ${}^{143}\text{Nd}/{}^{144}\text{Nd}$ ratios,
37 suggesting an unusual mantle source characterized by long-term U and Th enrichment relative to
38 Pb without an associated increase in Rb/Sr (Zindler and Hart, 1986). HIMU basalts from
39 different localities have remarkably uniform trace element features, such as Nb and Ta
40 enrichment relative to other incompatible elements and depletion in highly incompatible
41 elements (e.g., Cs, Rb, and Ba) and fluid-mobile elements (e.g., K, Pb, and Sr) (e.g., Kogiso et
42 al., 1997a; Thirlwall, 1997). Debate continues regarding the origin of HIMU, with an origin from
43 either metasomatized lithospheric mantle (e.g., Pilet et al., 2008; Weiss et al., 2016) or
44 dehydrated altered oceanic crust (e.g., Weaver, 1991; Chauvel et al., 1992), which undergoes
45 long-term isolation from the convective mantle (Thirlwall, 1997; Stracke et al., 2003; Hanyu et
46 al., 2011). An origin from partial melting of subducted oceanic slabs that have been buried in the
47 deep mantle has been also proposed to explain the hotspot/plume-related HIMU basalts, such as
48 at St. Helena, Mangaia, and Tubuaii (e.g., Chauvel et al., 1992; Hofmann, 1997; Hanyu et al.,
49 2011). In contrast, the lack of evidence for the presence of plume activities (e.g., large-scale
50 doming, linear space-time trends of volcanic centers, and large volumes of erupted magmas) let
51 to argue against a plume origin for many HIMU-type alkaline magmas that erupted in continental
52 settings (e.g., Hoernle et al., 2006; Panter et al., 2006).

53 Seismic tomography shows that the intraplate HIMU-like basalts in the Mediterranean area
54 are related to a prominent low-velocity anomaly with a plume-like shape rooted within the
55 mantle transition zone (MTZ) (Vinnik and Farra, 2007). The upwelling of transition zone
56 materials, which have been stagnated for a long period of time, has been proposed to explain the

57 mantle source of lamproites from Gaussberg in the East Antarctic shield (Murphy et al., 2002)
58 and an Enriched Mantle-1 (EM1) feature of Cenozoic potassic basalts from NE China (Kuritani
59 et al., 2011, 2013; Wang et al., 2017). As mixing and petrological homogenization processes are
60 less efficient in the MTZ than in the upper and lower mantle, the MTZ may be heterogeneous
61 and contain discrete lithological domains of former subducted oceanic crust (e.g., Ringwood,
62 1994; Kuritani et al., 2011). The long-term storage of recycled ancient dehydrated oceanic crust
63 in the MTZ thus has the potential to represent a HIMU-like mantle reservoir. HIMU-like basalts
64 can therefore help to identify recycled crustal components in the deep Earth and to understand
65 the mechanism for the long-term preservation of mantle reservoirs.

66 Several petrological and geochemical indexes can be used to determine the mantle source of
67 HIMU-like basalts. For example, experimental studies on mantle melting have demonstrated that
68 melt derived from a metasomatized lithospheric mantle (amphibole/phlogopite-bearing
69 peridotite) has a distinct chemical composition from that derived from recycled oceanic crust
70 (eclogite and/or pyroxenite) (e.g., Pilet et al., 2008; Kogiso and Hirschmann, 2006). Moreover,
71 the distinct water distribution in Earth's mantle may provide a new constraint on the mantle
72 source of basalts. The mantle transition zone (MTZ) is proposed to be hydrated and acts as a
73 huge tank in Earth's deep-water cycle (Karato, 2010; Wang et al., 2015, 2016) in accordance
74 with geophysical observations (Meier et al., 2009), electrical-conductivity experiments (Huang et
75 al., 2005), and mineral inclusions in diamonds (Pearson et al., 2014). Whereas, the convecting
76 asthenosphere (Saal et al., 2002) and lower mantle (e.g., Bolfan-Casanova et al., 2002, 2003) are
77 considered to have much lower H₂O contents than the MTZ. Thus, basalts derived from different
78 mantle layers may contain distinct water contents. In addition, hydrogen isotopes will be
79 fractionated during oceanic slab subduction. Studies on hydrous minerals in subaerial arc rocks,

80 such as amphibole, point to a high δD source for magmatic water (ranging from -45 to -20‰)
81 (Miyagi and Matsubaya, 2003). Thus, lithospheric mantle metasomatized by dehydration of
82 subducting slab will have relatively higher hydrogen isotopes than the pristine mantle, which is
83 contrary to the dehydrated recycled slab (Shaw et al., 2008; Xu et al., 2014). Therefore, the water
84 content and hydrogen isotope of mantle-derived primary melts may be one of the important
85 indexes to identify a HIMU source.

86 Because olivine is an early crystallizing phase in basalts, olivine-hosted melt inclusions
87 should best represent primary magmas in equilibrium with their mantle sources (e.g.,
88 Danyushevsky et al., 2000;). In this paper, we conducted a comprehensive geochemical study of
89 the Mesozoic intraplate basalts with HIMU-like features in the interior (Antang) of South China,
90 including *in situ* element and H-O isotope analyses on olivines and their melt inclusions, together
91 with major, trace element and Sr-Nd-Pb-Hf-Os isotope analyses of whole-rock samples. This
92 new dataset enables us to estimate the water content and composition of primary mantle-derived
93 magma and to constrain the source lithology, and provides new perspectives on the origin of
94 intraplate basalts in the light of deep water cycling.

95

96 **2. Geological background and sample descriptions**

97 The South China Block (SCB) is divided into the Cathaysia Block in the southeast and the
98 Yangtze Craton in the northwest (Fig. 1). The Yangtze Craton consists of an Archean-
99 Paleoproterozoic basement, including the 3.3–2.9 Ga Kongling Complex in the Yangtze Gorge
100 area (Gao et al., 2011). Basement outcrops in the Cathaysian Block include the Badu, Chencai
101 and Baoban complexes, with formation ages from Paleoproterozoic to Mesoproterozoic (Yu et
102 al., 2007). The Yangtze and Cathaysia blocks were amalgamated to form the unified SCB during

103 the Neoproterozoic Sibao orogeny (~1.0–0.8 Ga) (Li et al., 2009).

104 Igneous activity in South China has been voluminous since the Mesozoic (Zhou et al.,
105 2006). The intense magmatism, lithospheric deformation, and metal mineralization initiated since
106 middle Permian, possibly related to subduction of the paleo-Pacific Oceanic Plate beneath East
107 Asia (Li et al., 2012). The Mesozoic volcanic lavas in South China are predominantly felsic in
108 composition (95%) with subordinate mafic magmas (<5%) (Zhou et al., 2006). The Jurassic
109 basaltic lavas are sporadically distributed across the Cathaysia Block, including the Ningyuan,
110 Antang, Daoxian, Yizhang, Chenlong, Huilongyu, Zhaibei, Fankeng, Baimianshan, and
111 Dongkeng (Fig. 1).

112 Antang village is located along the Chenzhou-Linwu Fault in the Cathaysia Block, south of
113 the Jiangshan-Shaoxing Fault (Fig. 1). The Jurassic Antang basalts (168 Ma) (Wang et al., 2004)
114 are fresh, dark green in colour with euhedral to subhedral olivine (~10%) and lesser
115 clinopyroxene phenocrysts (<4%) up to 1.0 mm in size in a dominantly aphanitic groundmass,
116 composed of microcrystalline olivine, clinopyroxene, plagioclase and Fe-Ti oxides. Melt
117 inclusions are preserved in olivine and occur as droplets <50 µm in size, comprising glass,
118 vapour bubbles and microcrystalline clinopyroxene and olivine (Fig. 2).

119

120 **3. Analytical techniques**

121 *3.1. Whole-rock major and trace element analyses*

122 Seven fresh basalt samples were collected from Antang village (27°0'43.2"N;
123 114°42'38.5"E) for whole-rock geochemical and isotopic analyses. These rock samples were cut
124 into millimeter-scale chips and removed weathered rims. After washed in 3% HCl and cleaned
125 with de-ionized water, these chips were crushed into small fragments (<0.5 cm in diameter) in a

126 corundum crusher and then reduced to 200 mesh in an agate ring mill. All analyses were
127 undertaken at the State Key Laboratory of Isotope Geochemistry, Guangzhou Institute of
128 Geochemistry, Chinese Academy of Sciences (GIGCAS). One randomly selected sample (JA16-
129 12R) was performed to monitor the data quality and reproducibility. Whole-rock major oxides
130 were determined using a Rigaku RIX 2000 X-ray fluorescence spectrometer (XRF) on fused
131 glass disks, with analytical errors less than 3% for SiO₂, Al₂O₃, Fe₂O₃, MgO, CaO, Na₂O and
132 K₂O, and less than 5% for TiO₂, MnO and P₂O₅. Trace element analyses were carried out using a
133 Perkin-Elmer ELAN 6000 inductively coupled plasma mass spectrometer (ICP-MS). The
134 analytical precision is generally better than 5% for element with concentrations > 200 ppm, and
135 5–10% when less than 200 ppm.

136

137 *3.2. Whole-rock Sr-Nd-Pb-Hf-Os isotope analyses*

138 Whole-rock Sr-Nd-Pb-Hf isotopic compositions were determined using a Neptune multi-
139 collector ICP-MS (MC-ICP-MS). The measured ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd ratios were
140 normalized to ⁸⁶Sr/⁸⁸Sr = 0.1194 and ¹⁴⁶Nd/¹⁴⁴Nd = 0.7219, respectively. Reference standards
141 were analyzed along with samples and gave ⁸⁷Sr/⁸⁶Sr = 0.710297 ± 8 (2σ, n = 7) for NBS987 and
142 ¹⁴³Nd/¹⁴⁴Nd = 0.512101 ± 5 (2σ, n = 7) for JNDi-1 (0.512115 ± 5). For Pb isotopic analyses, 150
143 mg of rock powder was placed into a Teflon vessel and dissolved in an HNO₃ + HF mixture for
144 three days. Pb was separated and purified by conventional anion exchange techniques (AG1X8,
145 200–400 resin) with diluted HBr as eluant. The internal correction performed for the Pb isotopic
146 analyses used a thallium solution. Repeated measurements of standard NBS981 yielded
147 ²⁰⁶Pb/²⁰⁴Pb = 16.9311 ± 4 (2σ, n = 8), ²⁰⁷Pb/²⁰⁴Pb = 15.4838 ± 5 (2σ, n = 8) and ²⁰⁸Pb/²⁰⁴Pb =
148 36.6757 ± 14 (2σ, n = 8). For Hf isotope analyses, the ¹⁷⁶Hf/¹⁷⁷Hf ratios were normalized to

149 $^{179}\text{Hf}/^{177}\text{Hf} = 0.7325$. The reported $^{176}\text{Hf}/^{177}\text{Hf}$ ratios were adjusted to the standard solution JMC-
150 475 of 0.282160. The reference standard gave $^{176}\text{Hf}/^{177}\text{Hf} = 0.282184 \pm 4$ (2σ , $n = 9$). Analyses
151 of the USGS reference material BHVO-2 yielded $^{87}\text{Sr}/^{86}\text{Sr} = 0.703511 \pm 9$ ($n = 5$), $^{143}\text{Nd}/^{144}\text{Nd} =$
152 0.512972 ± 5 ($n = 5$), $^{206}\text{Pb}/^{204}\text{Pb} = 18.6468 \pm 6$ ($n = 5$), $^{207}\text{Pb}/^{204}\text{Pb} = 15.5221 \pm 6$ ($n = 5$),
153 $^{208}\text{Pb}/^{204}\text{Pb} = 38.1957 \pm 17$ ($n = 5$) and $^{176}\text{Hf}/^{177}\text{Hf} = 0.283093 \pm 4$ ($n = 5$), being in agreement
154 with the recommended values of $^{87}\text{Sr}/^{86}\text{Sr} = 0.703481 \pm 20$, $^{143}\text{Nd}/^{144}\text{Nd} = 0.512983 \pm 10$,
155 $^{206}\text{Pb}/^{204}\text{Pb} = 18.6173 \pm 465$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.5355 \pm 54$ and $^{208}\text{Pb}/^{204}\text{Pb} = 38.2108 \pm 384$ and
156 $^{176}\text{Hf}/^{177}\text{Hf} = 0.283096 \pm 20$ (Weis et al., 2005). The blanks were generally <100 pg for Sr and
157 <50 pg for Nd, Pb and Hf.

158 The basaltic samples chosen for Re-Os isotope analyses were carefully fragmented into
159 chips (<0.5 cm), and then leached and cleaned in 3% HCl for half an hour. The Re concentrations
160 were measured by isotope dilution-inductively coupled plasma-mass spectrometry (ID-ICP-MS,
161 Thermo-Scientific X2). The isotopic abundance of OsO_3^- and ReO_4^- were measured on a thermal
162 ionization mass spectrometer (TIMS, Thermo TRITON). The detailed procedure was described
163 in Li et al. (2014). Total blank levels were 1.9 ± 4.2 pg (2σ) and 0.58 ± 0.02 pg (2σ) for Re and
164 Os, respectively, with an $^{187}\text{Os}/^{188}\text{Os}$ ratio of 0.1109 ± 0.005 (2σ). The reference basalt BIR-1 was
165 also analyzed in the analytical batch, yielding Re = 687 ± 25 ppt, Os = 362 ± 13 ppt, and $^{187}\text{Os}/^{188}\text{Os}$
166 = 0.1338 ± 2 , in good agreement with the published data for BIR-1 (Re = 675 ± 7 ; Os = 355 ± 20 ,
167 $^{187}\text{Os}/^{188}\text{Os} = 0.1337 \pm 4$; Ishikawa et al., 2014). The different Re and Os concentrations and
168 $^{187}\text{Os}/^{188}\text{Os}$ for repeated analyses of the same samples are possibly related to the nugget effect
169 (Reisberg and Meisel, 2002).

170

171 *3.3. In situ chemical and oxygen isotope analyses of olivine*

172 The chemical composition of olivine was determined by electron probe micro-analyses
173 (EPMA) using a JEOL JXA-8100 at GIGCAS. Olivine analyses followed the procedures of
174 [Sobolev et al. \(2007\)](#). The operating conditions for olivine analyses were an accelerating voltage
175 of 20 kV, a probe current of 3.0×10^{-7} A and a spot size of 3 μ m. An internal olivine standard
176 (from mantle xenolith, Hannuoba) was analyzed before and after each batch (every five samples)
177 of analyses. The relative errors for the olivine standard were less than 1% for major elements
178 (SiO_2 , MgO and FeO), and 4.5% for minor elements (CaO, MnO and NiO) ([Table S1a](#)).

179 Oxygen isotopic compositions of olivine were analyzed with a CAMECA HR-IMS-1280
180 ion microprobe at GIGCAS. The Cs^+ primary beam was accelerated at 10 kV with an intensity of
181 ca. 2.0 nA. The spot size was about 20 μ m in diameter (10 μ m beam diameter + 10 μ m raster).
182 An electron gun was used to compensate for sample charging during analyses. Secondary ions
183 were extracted with a -10 kV potential. Oxygen isotopes were measured in multi-collector mode
184 with two off-axis Faraday cups with each analysis consisting of 20 cycles \times 4 counting time. The
185 external reproducibility of the standard (San Carlos) for $\delta^{18}\text{O}$ was 0.4‰ (2 σ), which was set as
186 the error for individual measurements. Measured $^{18}\text{O}/^{16}\text{O}$ ratios were normalized using the
187 Vienna Standard Mean Ocean Water composition (VSMOW, $^{18}\text{O}/^{16}\text{O} = 0.0020052$), and then
188 corrected for instrumental mass fractionation (IMF). Then the $\delta^{18}\text{O}$ for samples were corrected by the
189 equation: $\delta^{18}\text{O}_{\text{sample}} = \delta^{18}\text{O}_{\text{measured}} + \text{IMF}$. Many studies have shown that the impact of mass
190 fractionation (IMF) of SIMS oxygen isotope analysis on olivine could be negligible providing
191 that the olivine has Fo number ranging from 70 to 92 (e.g., [Guo et al., 2013](#); [Wang et al., 2015](#)).
192 Thus, the potential IMF effect, especially for olivine with Fo >80, should be negligible.

193

194 *3.4. Major and trace element analyses of homogenized melt inclusion (HMI)*

195 To obtain the chemical composition of olivine-hosted melt inclusions, it is preferable to
196 analyze a homogenized glass rather than a mixture of various crystalline phases and residual
197 glass (Danyushevsky et al., 2000). Prior to analysis, a 1.0 atm gas-mixing furnace was used to
198 heat the olivine grains to 1250°C at the quartz-fayalite-magnetite buffer. The platinum packet
199 was then rapidly raised back to the top of the furnace tube after being heated for 10 min to
200 achieve homogenization of the glassy inclusions. Then the host olivine was embedded in epoxy
201 resin and polished on one side until the melt inclusions were exposed. Well-preserved melt
202 inclusions with a diameter larger than 30 µm were chosen for analysis. Major element
203 compositions of the homogenized melt inclusions were analysed by EPMA with a JEOL JXA-
204 8100 Electron Microprobe following the procedure reported by Wang and Gaetani (2008). The melt
205 inclusions were analyzed using a 15 kV accelerating voltage, 200 nA beam current and 3 µm beam
206 diameter. The volatile elements (Na, K) were analyzed with peak counting time of 10 s and
207 background counting time of 5 s. Repeated analyses on the internal glass standard (JB-2) showed that
208 the analytical uncertainty was less than 2% for SiO₂, Al₂O₃, FeO and CaO, and ~5% for TiO₂, MgO,
209 Na₂O and K₂O (Table S1b).

210 The trace element composition of the melt inclusions was analysed by a CAMECA HR-
211 SIMS-1280 ion microprobe at GIGCAS, using a 10 nA, -13kv primary beam of O₂⁻ ions and
212 detection of positive secondary ions with a nominal acceleration voltage of + 10 kV. The target
213 area on the sample was pre-sputtered for 100 s using a 25 µm × 25 µm rastering technique to
214 remove the gold coat. During the analysis, the contrast aperture of 400 µm and the field aperture
215 of 5000 µm×5000 µm were used. The magnification of the transfer system was configured as
216 ~100 (equivalent to 80 µm). The entrance slit and the exit slit was set at ~200 µm and 600 µm,

217 respectively. The energy window was 50 eV and 5 eV energy gap. Energy filter was employed -
218 55 V for most elements except for Rb and Pb (-75 V), and Er and Yb (-65 V). Each analysis
219 consisted of a 10-cycle running time and the following isotopes were analysed: ^{28}Si , ^{30}Si , ^{85}Rb ,
220 ^{88}Sr , ^{89}Y , ^{90}Zr , ^{93}Nb , ^{138}Ba , ^{139}La , ^{140}Ce , ^{141}Pr , ^{143}Nd , ^{149}Sm , ^{157}Gd , ^{165}Ho , ^{167}Er , ^{172}Yb , ^{208}Pb ,
221 ^{232}Th , and ^{238}U . A single analysis normally took 20 minutes. The trace element concentrations of
222 melt inclusions were normalized to ^{28}Si , which had already been measured by EPMA. Rb, Sr, Y,
223 Zr, Nb, Ba, La, Ce, and Nd calibrations were performed using the values obtained by repeated
224 analyses of standards BHVO-2G, BCR-2G, BIR-1G GSD-1G, and GSC-1G. Pr, Sm, Gd, Ho, Er,
225 Yb, Pb, Th, and U calibrations were performed using those analyzed values of standards BHVO-
226 2G, BCR-2G, BIR-1G, and GSC-1G. Repeated analyses of the monitor glass standard (TB-1G)
227 showed that the analytical errors were usually less than 15%, except for Th, U and Pb (~25%)
228 (Table S1c).

229

230 3.5. Major element analyses of glass and mineral in unhomogenized melt inclusion (UMI)

231 Major elements of glass and mineral in UMI were determined by a Shimadzu Electron
232 Probe Microanalyzer EPMA-1720 at Sun Yat-Sen University. The analytical conditions are 15
233 kV accelerating voltage, 20 nA beam current and 3 μm beam diameter. The JB-2 was used as the
234 internal standard for the glass analysis. Repeated analyses on the internal glass standard showed
235 that the analytical uncertainty was less than 2% for SiO_2 , FeO , CaO and K_2O , and 5% for TiO_2 ,
236 Al_2O_3 , MgO and Na_2O .

237

238 3.6. H_2O content and H isotope analyses of UMI

239 Previous studies of olivine-hosted melt inclusions showed that loss of hydrogen occurred

240 during laboratory heating of the host olivine, which could also lead to a decrease of the H₂O
241 content and an increase in the hydrogen isotope composition of the melt inclusions (Chen et al.,
242 2011; Gaetani et al., 2012). Thus, the H₂O and H isotope analyses in this study were performed
243 on the original melt inclusion without homogenization.

244 During sample preparation, the olivine grains were embedded in Crystal-bond resin to avoid
245 the contamination of epoxy in cracks. They were ground and polished on one side until the melt
246 inclusions were exposed. Then the olivine grains were cleaned in acetone in an ultrasonic bath to
247 remove Crystal-bond resin. The resin-free polished olivine grains were set into indium disks.
248 After coating with gold, the samples were loaded and stored in the vessel chamber of a Nano-
249 SIMS 50L for three days. The chamber was maintained at high vacuum ($< 5 \times 10^{-10}$ torr) during
250 the analytical session. More detailed descriptions of the sample preparation are reported in Hu et
251 al. (2015).

252 The hydrogen isotopic compositions and water content were measured with a Cameca
253 Nano-SIMS 50L at the Institute of Geology and Geophysics, Chinese Academy of Sciences.
254 Each 10 $\mu\text{m} \times 10 \mu\text{m}$ analyses area was rastered by a Cs⁺ beam of ~0.5 nA and a diameter of ~5
255 μm . An electron gun was applied for sample surface charge compensation, which maintains an
256 electron cloud over the area of analysis. Each analytical site was pre-sputtered for 2 min by
257 rastering a 10 $\mu\text{m} \times 10 \mu\text{m}$ with a beam current of 2 nA to eliminate surface contamination and
258 achieve stable yield rates of the secondary ions. A mass resolving power (MRP) of 1800–2000
259 ($M/\Delta M$, Cameca definition related to the width between 10% and 90% intensity of the peak) is
260 sufficient to resolve ²D⁻ from the interference of ¹H₂⁻ and ¹⁸O⁻. Each analysis contained 20
261 blocks of 20 cycles with a total counting time of ~3 min, and the ions were collected with
262 electron multipliers. The detailed analytical procedure is described in Hu et al. (2015, 2019).

263 During the analyses, the instrumental mass fractionation (IMF), matrix effect and water
264 concentration calibration curves were made by analyzing two apatite reference materials,
265 Durango apatite (0.0478 wt% H₂O) and Kovdor apatite (0.98 ± 0.07 wt% H₂O and δD = - 66 ±
266 21‰), and one MORB glass reference material SWIFT MORB glass (0.258 wt% H₂O). Previous
267 water and H isotope analyses have showed similar matrix effects for silicates with a range from
268 basaltic to rhyolitic composition, and there were no observable difference of matrix and IMF
269 effects in both apatite and silicate glasses by NanoSIMS (Hu et al., 2015, 2019). The H isotope
270 compositions were calibrated for instrumental mass fractionation (IMF), using $\alpha_{\text{IMF}} =$
271 $(\text{D}/\text{H}_{\text{measured}})/(\text{D}/\text{H}_{\text{recommended}})$. The hydrogen isotopes are presented in the form of $\delta\text{D} =$
272 $((\text{D}/\text{H})_{\text{sample}}/(\text{D}/\text{H})_{\text{SMOW}} - 1) \times 1000\text{‰}$, where SMOW is the standard mean ocean water with a
273 D/H ratio of 1.5576×10^{-4} (McKeegan et al., 2006). The water content was determined from the
274 ¹H intensities relative to ¹⁸O⁻ using the calibration curves of $[\text{H}^{18}\text{O}] = \alpha \times [\text{H}_2\text{O}] + \beta$. Repeated
275 analysis of those standards give $[\text{H}^{18}\text{O}] = 1.0039 \times [\text{H}_2\text{O}]$, $R^2 = 0.9944$, and $\alpha_{\text{IMF}} = 0.897 \pm$
276 0.012 (2σ) (Table S1d). One of the analytical errors is from the statistical uncertainty of the D/H
277 ratios (most < 30‰ for δD) and ¹H/¹⁸O ratios (<0.3 wt% for H₂O content). The other analytical
278 error is ascribed to the uncertainties in α_{IMF} values and the slopes of the calibration curves. Thus,
279 the reproducibility was monitored before and after each batch analysis through repeated
280 measurements of the MORB glass (SWIFT) using the obtained α_{IMF} and the calibration curves.
281 The measured water contents and H isotopic compositions of standard SWIFT are listed in Table
282 S1e.

283

284 4. Results

285 4.1. Chemical and isotopic composition of the bulk rocks

286 The Antang basalts have an alkaline affinity and span a narrow whole-rock compositional
287 range (Table S2), with SiO₂ = 46.4–46.8 wt%, MgO = 8.3–8.8 wt%, and CaO = 7.9–8.3 wt%.
288 Their mantle-normalized trace element patterns show enrichments in Nb-Ta, depletion of fluid-
289 mobile (e.g., Rb, K and Pb), and heavy rare earth elements (HREEs), analogous to typical HIMU
290 basalts from St. Helena (e.g., Weaver, 1991; Kogiso et al., 1997a; Hanyu et al., 2011) (Fig. 3).
291 These rocks also display unradiogenic Sr (⁸⁷Sr/⁸⁶Sr(i) = 0.7032–0.7040), moderately radiogenic
292 Nd (¹⁴³Nd/¹⁴⁴Nd(t) ~ 0.5127), Hf (¹⁷⁶Hf/¹⁷⁷Hf(t) ~ 0.2830), and Os (¹⁸⁷Os/¹⁸⁸Os(t) = 0.1653–
293 0.2740), and highly radiogenic Pb isotopic compositions, e.g., ²⁰⁶Pb/²⁰⁴Pb = 19.17–19.44,
294 ²⁰⁷Pb/²⁰⁴Pb = 15.55–15.59 and ²⁰⁸Pb/²⁰⁴Pb = 39.29–39.56 (Table S2; Fig. 5).

295

296 4.2. Chemical and O isotopic compositions of olivine

297 The olivine phenocrysts span a range of forsterite (Fo) compositions between 77.9 and 85.9.
298 They have lower CaO contents (0.07–0.26 wt%, Table S3) than olivines from MORBs and
299 carbonated peridotite-derived basalts, and most have CaO contents lower than olivine in the
300 plume-related Hawaii basalts (Fig. 6a). The morphologies, textures, and compositions of olivine
301 and the presence of melt inclusions strongly support a magmatic origin for the olivine
302 phenocrysts. Most of olivines yield mantle-like oxygen isotope compositions, with a δ¹⁸O range
303 of 5.04–5.50‰, except for some olivines with Fo_{82.0-85.9} showing slightly higher δ¹⁸O values up
304 to 6.01‰ (Table S3; Fig. 6b).

305

306 *4.3. Composition of HMI and estimated primary magma*

307 Olivine-hosted melt inclusions may re-equilibrate with their host olivine by Fe-Mg
308 exchange reaction (Danyushevsky et al., 2000). This re-equilibrium process is generally called
309 “Fe-loss” and usually results in an increase of MgO and a decrease of FeO^T (total iron oxides,
310 Danyushevsky et al., 2000). The negative correlation between FeO content of the melt inclusions
311 and Fo of the host olivine indicates that there was Fe-Mg exchange between them (Fig. S1).
312 Furthermore, the nearly constant CaO/Al₂O₃ ratio follows the range of Fo of olivine between
313 82.2 and 85.9, suggesting only olivine fractionation during magmatic differentiation (Fig. 4). We
314 thus use the PETROLOG software (Danyushevsky and Plechov, 2011) to adjust the measured
315 melt inclusion compositions using the “Fe-loss” correction so that we can obtain the composition
316 of the melt in equilibrium with the host olivine. This calculation requires an independent
317 estimation of the initial trapped melt FeO^T content. According to the FeO^T fractionation trend, 10
318 wt% is set as the FeO^T content in the initially trapped melt (Table S4a; Fig. S1). After applying
319 the “Fe-loss” correction, the homogenized melt inclusions in olivine show a range in SiO₂ of
320 44.6–47.3 wt%, in MgO of 6.6–8.3 wt% and in CaO of 8.8–10.7 wt% (Table S4b). The whole-rock
321 chemical composition can be roughly estimated as 93–96% melt inclusion plus 4–7% olivine,
322 consistent with the dominance of olivine phenocrysts in the basalt.

323 In the case of unique olivine fractionation, the primary magma composition of basalts can
324 be calculated through addition of olivine back into the melt until chemical equilibrium between
325 the estimated primary magma and mantle olivine is reached. We therefore selected those melt
326 inclusions within high-Fo (>82.2) olivine to estimate the primary magma composition of the
327 Antang basalts through addition of olivine in 0.1% increments (assuming that 10% of the total
328 iron is Fe³⁺ and $(\text{Fe}^{2+}/\text{Mg})_{\text{olivine}} / (\text{Fe}^{2+}/\text{Mg})_{\text{melt}} = 0.3$). The melt compositions are corrected so that

329 they are in equilibrium with an olivine composition of Fo₉₀. The estimated primary magmas span
330 a SiO₂ range of 44.3–46.1 wt%, a MgO range of 13.8–14.4 wt%, and a CaO range of 7.3–8.7 wt%
331 (Table S4c), which can be reconstructed by adding 14–19% olivine back into 81–86% corrected
332 melt inclusion. The magmas also have low CaO contents compared with the experimental melts
333 of peridotite, but show similar CaO contents to melts of eclogitic or pyroxenitic sources (Fig. 7).

334 The melt inclusions also show similar trace element patterns to the bulk rocks, confirming
335 the HIMU-like feature of Antang basalts (Table S5; Fig. 3b).

336

337 4.4. Composition, water content and H isotope ratio of UMI

338 Glasses in UMI show a range in SiO₂ of 56.2–62.9 wt%, in Al₂O₃ of 17.4–23.6 wt%, in CaO
339 of 1.11–5.04 wt%, and K₂O of 1.34–4.16 wt% (Table S7). The recrystallized mineral in the UMI
340 belongs to augite, spanning a SiO₂ range of 40.9–42.8 wt%, an Al₂O₃ range of 13.55–15.59 wt%,
341 a FeO range of 6.70–9.38 wt%, a MgO range of 9.23–12.66 wt%, and a CaO range of 12.11–23.49
342 wt% (Table S7).

343 The UMI spans a large range in H₂O from 0.34 to 4.97 wt% and a δD range between -102 and
344 141‰ (Table S6). The H₂O content shows a broadly negative correlation with δD (Fig. 8a). The
345 melt inclusion hosted in olivine with Fo number less than 82 has a relatively low H₂O content
346 and δD value (Fig. 8b).

347

348 5. Discussion

349 5.1. Effects of Alteration, Crustal Assimilation, and Fractional Crystallization

350 Basaltic rocks can undergo open-system compositional evolution through processes such as
351 crustal contamination or assimilation, magma mixing, and surface weathering or alteration. The

352 absence of secondary minerals (Fig. 2a) suggests an insignificant effect of alteration on the
353 Antang basalts. However, the whole-rock composition may be easily affected by the surface
354 alteration, especially for fluid-mobile elements such as K, Rb, Ba, and Sr. As shown in Fig. 3 and
355 Table S2, except for the sample JA16-14 showing a fluctuating Rb, the other incompatible
356 elements show consistent trace element patterns. This suggests that surface alteration should be
357 negligible for the most of the Antang basalts.

358 The Mesozoic Antang basalts and their olivine-hosted melt inclusions have high Nb/U and
359 Ce/Pb ratios and show positive Nb-Ta anomalies and prominent Rb, K and Pb depletion (Fig. 3),
360 indicating a negligible role for crustal contamination or assimilation. The relatively
361 homogeneous compositions of the olivine-hosted melt inclusions also preclude any significant
362 magma mixing in their generation.

363 Fractional crystallization of mantle-derived primary melts during magmatic evolution
364 would significantly change the chemical composition of the melts. The Antang basalts show
365 lower Mg# (65.4–66.7) than those of typical mantle-derived primary melts ($Mg\# = 73\text{--}81$),
366 implying a role of fractional crystallization. Melt inclusions are small portions of magma trapped
367 by growing crystals and represent snapshots of the crystallization environment (Danyushevsky et
368 al., 2000). The correlation between Fo of the host olivine and CaO/Al₂O₃ ratios of the melt
369 inclusions indicates that clinopyroxene fractionation took place at $Fo < 82.2$ (Fig. 4). Moreover,
370 the lack of Sr and Eu anomalies (Fig. 3a) implies that fractional crystallization of plagioclase was
371 negligible during the formation of Antang basalts. In summary, the geochemical features of
372 whole-rock and melt inclusion indicate the Antang basalts mainly experienced a role of olivine +
373 clinopyroxene fractionation, whereas the effect of alteration and crustal assimilation could be
374 negligible.

375

376 *5.2. Effects of magmatic evolution and post-entrapment processes on H₂O content of melt*

377 *inclusion*

378 Since H₂O is more readily exchanged than most incompatible elements during magmatic
379 evolution and post-entrapment processes, it is important to assess whether the magmatic H₂O
380 concentrations of the melt inclusions have been compromised by exotic fluid addition, magma
381 degassing, and post-entrapment diffusion and crystallization (e.g., [Chen et al., 2011](#); [Gaetani et](#)
382 [al., 2012](#)).

383 Most of olivine phenocrysts have mantle-like $\delta^{18}\text{O}$ values ([Fig. 6b](#)), confirming the
384 negligible influence of low-temperature alteration. Any addition of meteoric water or seawater
385 with high δD should produce a positive correlation between H₂O and δD , which is opposite to
386 the negative correlation in the UMI ([Fig. 8a](#)). Aqueous fluid addition will significantly change
387 concentrations of fluid-mobile elements of olivine-hosted melt inclusions ([Kogiso et al., 1997b](#)).
388 However, the melt inclusions in the Antang basalts show low Rb/Nb, Ba/Nb, and Sr/Nb ratios
389 and positive correlations of Zr with Rb, Ba, Th, and U, precluding the possibility of extra fluid
390 addition during olivine crystallization ([Fig. 9](#)).

391 Although we analysed the water content of UMI to avoid water loss during the heating and
392 homogenization processes, the loss of water (or H diffusion) may also have occurred during
393 eruption and cooling of the magma (e.g., [Saal et al., 2002](#); [Gaetani et al., 2012](#)). During magma
394 ascent, the solubility of water in magma is mainly controlled by pressure and may reach the
395 critical point at shallow depths. Magmatic degassing will reduce both the water content and δD .
396 The melt inclusion hosted in olivine with Fo number ranging from 82 and 86 exhibits a negative
397 correlation between H₂O and δD , which is contrary to the effect of magmatic degassing ([Fig. 8a](#)).

398 However, the UMIs hosted in olivine with Fo < 82 show a dramatical decrease of H₂O and δD
399 ratio, which could be affected by degassing (Fig. 8).

400 After entrapment in the host olivine, the H₂O content of melt inclusion may decrease via
401 hydrogen diffusion through olivine to keep dynamic equilibrium with the residual magmas (Chen
402 et al., 2011). It is nearly impossible to avoid diffusive loss of hydrogen during magma ascent,
403 because the timescale of magma ascent is close to that of diffusive water exchange as
404 experimentally determined (Hauri, 2002b; Gaetani et al., 2012). Since the diffusion of H is much
405 faster than D, hydrogen diffusion through host olivine will lead to an increase of δD in
406 association with a decrease of H₂O content in melt inclusion. A large number of UMIs from the
407 Antang basalts have fairly positive δD values, which were probably a result of post-entrapment
408 kinetic diffusion (Fig. 8a). By contrast, the UMI with the highest water content (up to 5.0 wt%)
409 and mantle-like δD value experienced insignificant hydrogen diffusion.

410 Crystallization of daughter mineral in melt inclusion will lead to an increase of water
411 content in the residual glass. This process does not reset δD value of the residual glass. The
412 petrographic examination (Fig. 2) shows that the water content of glass in the UMI cannot
413 represent the primary water content of the melt trapped by host olivine, because of the
414 crystallization of daughter mineral. In this respect, it is necessary to reconstruct the water content
415 of primary melt by analysing the relative volume of glass and recrystallizing mineral within the
416 UMI.

417 The recrystallized mineral in the UMI belongs to augite, which would dramatically decrease
418 the CaO/Al₂O₃ ratio of residual glass. Because the CaO/Al₂O₃ ratio of melt inclusion shows a
419 good correlation with Fo number of host olivine (Fig. 4), we calculate the proportion of glass and
420 mineral using the analysed CaO/Al₂O₃ ratios of glass and mineral in the UMI and the variation

421 trends of CaO/Al₂O₃ of HMI at given Fo component of host olivine. With the following equation,
422 we calculate the proportion of glass and mineral within each UMI:

$$423 \quad \text{CaO/Al}_2\text{O}_3 \text{ (HMI)} = a \times \text{CaO/Al}_2\text{O}_3 \text{ (glass in UMI)} + b \times \text{CaO/Al}_2\text{O}_3 \text{ (mineral in UMI)}$$

$$424 \quad \text{Where } a + b = 1$$

$$425 \quad \text{CaO/Al}_2\text{O}_3 \text{ (HMI)} = 0.56 \text{ (When the host olivine has Fo } > 82.2, \text{ Fig. 4)}$$

$$426 \quad \text{CaO/Al}_2\text{O}_3 \text{ (HMI)} = 0.1167 \times \text{Fo} - 9.03 \text{ (When the host olivine has Fo } < 82.2, \text{ Fig. 4)}$$

427 The results yield that the proportions of post-entrapment crystallization in these UMIs span a
428 range from 16 to 58% (Table S7). The melt inclusion with the highest water content (4.97 wt%)
429 and low δD experiences ~34% post-entrapment crystallization, yielding 3.28 wt% H₂O for the
430 bulk UMI, which should represent the minimum of its water content. Considering that the
431 Antang basalts might have experienced about 14–19% olivine fractionation, we finally estimate
432 that the primary magma of Antang basalts was hydrous with at least 2.65 wt% H₂O.

433

434 *5.3. Petrogenesis of the Antang basalts*

435 The HIMU-like whole-rock geochemistry and the H₂O-rich compositions indicate
436 derivation of the Antang basalts from a hydrous mantle reservoir. Possible mantle sources for
437 HIMU-like basalts include metasomatized lithospheric mantle (hornblendite, hornblende-bearing
438 mantle and carbonated peridotite) (e.g., Pilet et al., 2008; Weiss et al., 2016) and ancient recycled
439 oceanic crust (e.g., Hofmann, 1997; Hanyu et al., 2011). However, the low TiO₂, high SiO₂,
440 MgO and FeO contents and the lack of negative Zr-Hf anomalies distinguish the Antang basalts
441 from melts of hornblendite or hornblende peridotite (Fig. 3). In addition, the contemporaneous
442 HIMU-like Ningyuan basalts, which intrude the Cathaysia Block around 300 km to the
443 southwest (Wang et al., 2008) (Fig. 1), contain Mesozoic mantle xenoliths that are characterized
444 by the lack of hydrous minerals, nearly flat REE patterns, and depletion of Ba (Zhang et al.,

445 2008). These features of xenoliths imply that the lithospheric mantle beneath South China was
446 nearly anhydrous with little metasomatism from subducted slab-derived fluids. Furthermore, the
447 relatively low Os concentrations and high $^{187}\text{Os}/^{188}\text{Os}(t)$ ratios of the Antang basalts (Fig. 10a, b)
448 are inconsistent with melts derived from sub-continental lithospheric mantle (SCLM), which
449 typically has extremely high Os concentrations and low $^{187}\text{Os}/^{188}\text{Os}$ ratios. The mantle peridotite
450 xenoliths hosted in Jurassic Ningyuan basalts have higher Sr ($^{87}\text{Sr}/^{86}\text{Sr}(i) = 0.7048\text{--}0.7050$) and
451 lower Os ($^{187}\text{Os}/^{188}\text{Os}(t) = 0.1206\text{--}0.1293$) isotopic ratios than those of the Antang basalts
452 (Zhang et al., 2008; Liu et al., 2012) (Fig. 10c). If the Antang basalts were derived from the same
453 source as the mantle xenoliths, this could have required a large (>90%) continental crustal input
454 to match the Os-isotopic ratios of the primitive Antang magmas (Fig. 10d). This is inconsistent
455 with the HIMU-like trace element features of the Antang basalts that permit negligible crustal
456 contamination. We therefore conclude that the geochemical data from the Antang basalts argue
457 against derivation from the SCLM regardless of whether or not it experienced metasomatism.

458 A systematic study on the primary magmas has revealed that pyroxenite is the principal
459 lithology in the mantle source of Late Mesozoic basalts in SE China (Zeng et al., 2016).
460 Following the method in Thompson et al. (2005), the calculated primary melts were plotted on
461 the CIPW normative Di-Ol-Hy-(Ne+Lc)-Qz projection, along with experimentally determined
462 mantle-derived melts from peridotite, carbonated peridotite, carbonated eclogite, hornblendite,
463 clinopyroxene-hornblendite, silica-deficient pyroxenite and silica-excess pyroxenite (Fig. 11).
464 The Antang basalts plot within the compositional range of silica-deficient garnet-pyroxenite
465 melts. Because clinopyroxene is typically the major CaO-bearing mantle phase, leaving
466 clinopyroxene as a major residual phase is the only effective way of lowering the CaO content of
467 mantle-derived melts (Herzberg, 2011). The low CaO contents (7.3–8.7 wt%) of the primary

468 magmas (Fig. 7) also support a pyroxenitic residue rather than a peridotitic residue in the mantle
469 source. Moreover, the Antang basalts define a negative correlation of Th versus Nb/Ta and Nb/La
470 (Fig. S2), implying a bulk $D_{\text{Nb/Ta}} > 1$ and $D_{\text{Nb/La}} > 1$. This is a typical characteristic of an eclogitic
471 source because $D_{\text{Nb/Ta}}$ and $D_{\text{Nb/La}}$ for eclogitic garnet range from 1.2 to 2.0 and 1.3 to 7.5 (e.g.,
472 Klemme et al., 2002; Pertermann et al., 2004), respectively. Partition coefficients show that $D_{\text{Zr}} <$
473 D_{Hf} in clinopyroxene and $D_{\text{Zr}} > D_{\text{Hf}}$ in garnet (Klemme et al., 2002; Pertermann et al., 2004).
474 Thus, low degree melting of a clinopyroxene-rich eclogite can explain the elevated Zr/Hf ratios
475 (43–45) of the Antang basalts. The low CaO content of olivine (Fig. 6a) further indicates a
476 pyroxenitic source (Sobolev et al., 2007). In summary, the residual source region for the Antang
477 basalts was dominated by clinopyroxene and garnet.

478 Recycled upper oceanic mafic crust is known to form eclogitic or pyroxenitic mantle
479 reservoirs (e.g., Sobolev et al., 2007; Kogiso and Hirschmann, 2006). Extremely high $^{187}\text{Re}/^{188}\text{Os}$
480 ratios reported in oceanic crust (80 to 675) would lead to high radiogenic $^{187}\text{Os}/^{188}\text{Os}$ ratios over
481 1.0 Ga (Peucker-Ehrenbrink et al., 2012). A major contribution from such a mantle source is also
482 suggested by the relatively high whole-rock $^{187}\text{Os}/^{188}\text{Os}(t)$ ratios of the Antang basalts (Fig. 10)
483 and the relatively high $\delta^{18}\text{O}$ values in high-Fo (>82) olivine (Fig. 6b). In order to further
484 determine how the mantle source for the Antang basalts was formed, we compared their isotopic
485 compositions with modelled Sr-Nd-Pb-Hf isotopic evolution paths for both ancient subducted
486 oceanic crust and contemporary subducted sediment (Table S8, Stracke et al., 2003). Compared
487 with typical HIMU basalts, the Antang basalts have lower $^{206}\text{Pb}/^{204}\text{Pb}$ and higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios
488 (Fig. 5), and plot in the 1.0–1.5 Ga arrays defined by the mixing curves between contemporary
489 subducted oceanic crust and sediments. These results further suggest that the Antang basalts were
490 derived from a HIMU-like mantle reservoir composed principally of 1.0–1.5 Ga recycled oceanic

491 crust plus minor (<1 %) sediment. The involvement of minor sediment may also contribute to the
492 Ba enrichment observed in the Antang basalts (Fig. 3, Weaver, 1991).

493

494 5.4. Estimation of H₂O in the mantle source

495 Geochemically, H₂O behaves incompatibly during partial melting, with a partitioning
496 coefficient similar to that of Ce (Michael, 1995); thus, the H₂O/Ce ratio of magma has been
497 proven to be a useful geochemical tool to trace the hydrous state of mantle source (Dixon et al.,
498 2002; Xia et al., 2019). Considering the Ce concentrations of the HIMI between 73 and 135 ppm
499 and the primary H₂O content of melt inclusion is ≥ 3.28 wt%, the calculated H₂O/Ce ratios (243–
500 450) of the Antang basalts are higher than MORB (100–200, Michael, 1995; Saal et al., 2002)
501 and EM-type OIB (<150, Dixon et al., 2002), implying a hydrous mantle source.

502 In addition, the water content of the mantle source can be calculated at a given degree of
503 melting and the hydrogen partition coefficient. Using an inferred eclogitic/pyroxenitic source, we
504 then calculate the degree of partial melting based on the relationship between incompatible
505 elemental ratios (e.g., La/Yb versus Sm/Yb). As the La, Sm, and Yb concentrations of the
506 average depleted mantle (0.192 ppm, 0.239 ppm, and 0.365 ppm, respectively) are much less
507 than that of mafic oceanic crust (~3.83 ppm, ~3.11 ppm, and ~3.03 ppm, respectively) (Stracke et
508 al., 2003), some contribution of mantle peridotite could not greatly change the incompatible
509 elemental ratios. To simplify the modelling, we use the elemental ratios of recycled oceanic crust
510 as the starting component to calculate the degree of melting based on the relationship between
511 incompatible elemental ratios (Sm/Yb versus La/La and La/Yb versus La/Sm). The recycled
512 oceanic crust is assumed to consist of 99% mafic oceanic crust (25% N-MORB + 25% altered
513 MORB + 50% gabbro) and 1% subducted sediment represented by global subducted sediment

514 (GLOSS) (Reference can be seen in supplementary A1). Equilibrium non-modal melting
515 calculation indicates that 4.5–6.0% melting of the assumed source can produce melts with the
516 Antang basalt compositions (Fig. 12).

517 Previous SIMS analyses on H in nominally anhydrous minerals have facilitated direct
518 determination of mineral/melt hydrogen partitioning from coexisting mantle minerals and
519 hydrous silicate melts. A compilation of the published data over wide ranges of composition,
520 temperature and pressure yields the range of $D_H^{Grt/liq}$ from 0.001 to 0.003 and a $D_H^{Cpx/liq}$ range
521 from 0.01 to 0.03 (Hauri et al., 2006). The calculated bulk D_H value of eclogite with 80%
522 clinopyroxene plus 20% garnet ranges from 0.008 to 0.024. In combination with the estimated
523 primary magma that has $H_2O \geq 2.65$ wt%, we finally estimate that the mantle source for the Antang
524 basalts had H_2O contents between 0.14 and 0.22 wt%.

525

526 5.5. Origin from a hydrous MTZ

527 The lines of evidence from geochemical data of the bulk rock, olivine and melt inclusions
528 from the Antang basalts suggest that the basaltic magmas were derived from a hydrous mantle
529 source composed principally of eclogite/pyroxenite probably derived from an ancient recycled
530 oceanic slab. The Sr-Nd-Pb-Hf isotopic composition of this hydrous HIMU-like mantle reservoir
531 requires a long (1.0–1.5 Ga) mantle residence time and this requires at least two fundamental
532 conditions: (1) long-term isolation from the convective mantle to avoid chemical and isotopic re-
533 homogenization with the surrounding mantle, and (2) a relatively low temperature to avoid large-
534 scale partial melting and isotopic disturbance. Considering the asthenosphere is dry (< 200 ppm
535 H_2O) (Saal et al., 2002) and the convection would make the recycled oceanic crust hard to be
536 stagnated for a long time, these factors preclude the possibility that the Antang basalts were

537 derived from the low-viscosity convective asthenosphere. Having already ruled out derivation
538 from a metasomatized SCLM, the possible alternative sites include the MTZ, the lower mantle or
539 the D'' layer (i.e. the core-mantle boundary).

540 Cold oceanic slabs can penetrate into the lower mantle and may bring a certain amount of
541 water into the Earth's interior (Poli and Schmidt, 2002). The hydrous aluminosilicate phases,
542 such as δ -AlOOH and Al-rich phase D and phase H can be stable under lower mantle conditions
543 (Pamato et al., 2015). This suggests the possibility for the existence of locally hydrous sources
544 within the lower mantle. However, the following arguments exclude the possibility that the
545 Antang basalts were derived from the lower mantle: (1) Basalts associated with mantle plumes
546 containing recycled subducted slabs are relatively dry (e.g., Dixon et al., 2002; Shaw et al.,
547 2012). The geochemical features such as negative Pb anomalies (high U/Pb ratio or HIMU feature)
548 of basalts that originate in the lower mantle also indicate the intense dehydration of such recycled
549 subducted slabs (Kogiso et al., 1997b). The lower mantle is composed predominantly of Mg- and
550 Ca-perovskite and ferropericlase and has low water storage capacity (< 200 ppm) (e.g., Bolfan-
551 Casanova et al., 2002, 2003). Thus, the amount of water that can be transported into the lower
552 mantle would be minor; (2) Some OIBs have originated from the lower mantle indeed have low
553 water contents and δD values (Hauri et al., 2002b), indicating that most of the lower mantle is
554 relatively anhydrous; (3) Because the lower mantle is also convective (Chen, 2016), even if the cold
555 subducted slab brings a certain amount of water into this region, any locally hydrous source will be
556 rapidly homogenized. This means that it is difficult to form a long-term isolated hydrous source
557 within the lower mantle.

558 Compared with the normal mantle, a deep (e.g., derivation from the D'' layer) mantle plume
559 has an extremely high temperature. Based on the estimated primary magma composition and the

560 inferred hydrous eclogitic/pyroxenitic source of the Antang basalts, we calculate that the mantle
561 potential temperature (T_p) spans a range of 1331–1386°C (Table S4d; Supporting Information
562 A2). This T_p is similar to that estimated for the transition-zone derived hydrous mantle plume at
563 Changbaishan (Kuritani et al., 2019). Also, there is no geological evidence to support that a
564 mantle plume was active beneath South China during the Jurassic. Therefore, both the geological
565 records and the estimated lower T_p at that time imply that the D'' layer cannot be the source of
566 Antang basalts.

567 Alternatively, delamination and detachment of oceanic crust from the subducted slabs, with
568 subsequent long-term storage in the MTZ quarantined from the effects of convective stirring,
569 remains a viable option (Ringwood, 1994; Nebel et al., 2013). High-resolution tomographic
570 images in modern subduction zones demonstrate that subducted slabs in some regions do not
571 simply plunge directly through the 660 km discontinuity, deeply into the lower mantle (Kuritani
572 et al., 2011, 2013, 2019). The study of phase transformation in oceanic crust shows that the
573 basaltic crust component would be 0.1–0.2 g/cm³ lighter than its surrounding mantle pyrolite at
574 the depths of 660–800 km (Irifune and Ringwood, 1993). The buoyancy produced by internal
575 petrological differentiation may prevent basaltic crust entering the lower mantle and segregate it
576 to form a gravitationally stable layer overlying the 660 km discontinuity (Ringwood, 1994). The
577 MTZ thus may contain recycled oceanic crustal components, at least partially shielded from
578 mixing and petrological homogenization typical of the upper mantle. Therefore, the best
579 interpretation is that the source of the Antang basalts was a hydrous MTZ (0.14–0.22 wt% H₂O)
580 beneath the South China interior. This hydrous MTZ was isolated from mantle convection for
581 over one billion years.

582 The mechanism for upwelling of hydrous MTZ materials needs to be further discussed. The

583 estimated mantle potential temperature argues against thermal buoyancy is a primary cause for
584 the Antang basaltic magmatism. South China has been an active continental margin in
585 association with subduction of the paleo-Pacific Ocean beneath the Eurasian Continent during
586 Jurassic. Through 3-D numerical modelling, [Faccenna et al. \(2010\)](#) proposed that the focused
587 mantle upwelling can be generated both ahead of slab in the back-arc region (the disturbing
588 distance could be up to five times as that from trench to magmatic arc) and around the lateral
589 edges of the slab. Moreover, hydrous mantle peridotite is lighter than dry mantle peridotite, so a
590 wet focused mantle upwelling could be generated at the top of the MTZ ([Wang et al., 2016](#);
591 [Kuritani et al., 2019](#)). Generation of the Antang basalts might be ascribed to the co-effect of slab
592 subduction and hydrous mantle source that led to passive upwelling of hydrous MTZ materials in
593 South China interior.

594

595 *5.6. Implications for intraplate magmatism*

596 Subducted slabs in the present-day mantle display a wide variation in behaviour, where
597 some slabs penetrate into the lower mantle, for example beneath Peru, the Marianas and Central
598 America, whereas others seem to stagnate in the transition zone, for example beneath the Izu-
599 Bonin region, South-Kurile and Japan ([Agrusta et al., 2017](#)). While mantle heterogeneity may be
600 partly ascribed to internal mantle differentiation induced by partial melting and melt transport,
601 the recycled subducted components buried at the core-mantle boundary have been widely
602 accepted to form chemically heterogeneous reservoirs in the Earth ([Hofmann, 1997](#)). Regions of
603 buoyant upwelling mantle plume are considered to return some of this ancient oceanic crustal
604 material to the shallow mantle, where it melts to form compositionally variable OIBs ([Hofmann,](#)
605 [1997](#)). However, many intraplate OIB-type basalts form in splash, edge, or baby plumes, with the

606 lack of features characteristic of plume-related magmatism, and therefore cannot be attributed to
607 upwelling mantle plumes (Conrad et al., 2011). The compositional variation of these basalts may
608 thus need further explanation. Recently, several studies have proposed that the MTZ may be a
609 potential source for the generation of intraplate magmatism. Murphy et al. (2002) proposed that
610 subducted sediment sequestered in the MTZ for 2–3 Ga may be the mantle source of lamproites
611 from Gausberg. Kuritani et al. (2011) proposed that the MTZ, metasomatized by involvement of
612 ancient (>1 Ga) and recent sediment from a subducted slab, can explain the high Ba/Th and
613 $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of the Cenozoic basalts in NE China. Wang et al. (2015) found that
614 contribution of gabbroic components from stagnant slabs in the MTZ significantly affected the
615 chemical and oxygen isotopic values of continental flood basalts. Moreover, Wang et al. (2017)
616 proposed that recycled ancient carbonate-bearing sediments, subducted into the mantle transition
617 zone could form the EM1 reservoir of potassic basalts. Therefore, the present work, together with
618 previous studies, suggests that the MTZ can sequester subducted components without being
619 admixed into the convecting mantle for long periods of time, and thus form distinct mantle
620 reservoirs for the generation of intraplate basalts.

621

622 **6. Conclusions**

623 Our geochemical results from bulk rock, olivine and melt inclusions of Jurassic mafic lavas in
624 South China provide a new perspective on the origin of HIMU-like basalt. The following conclusions
625 can be summarized:

626 (1) The H_2O content of melt inclusion was mainly affected by post-entrapment crystallization and
627 kinetic diffusion through host olivine and can also be influenced by magmatic degassing for low-
628 Fo olivine-hosted melt inclusion. Using the least-affected melt inclusion that has the highest

629 water content (up to 5.0 wt%) and mantle-like δD value, the primary magma of the Antang basalts is
630 estimated to have $H_2O \geq 2.65$ wt% and the corresponding mantle source has H_2O concentration
631 between 0.14 and 0.22 wt%. The low δD values in the water-rich melt inclusions indicate that
632 water within the source is principally primordial.

633 (2) The Antang basalts were derived from a hydrous MTZ composed mainly of recycled oceanic
634 slabs that had stagnated for over 1.0 Ga. The dominant mantle lithology was eclogite or pyroxenite.

635 (3) The hydrous mantle transition zone is a potential source for generation of compositionally diverse
636 and unusually volatile-rich intraplate basalts.

637

638 **Acknowledgements**

639 We acknowledge H. Smithies for checking the English, suggestions and comments on an
640 earlier version of this paper. We also thank S. Hu., L. Zhang., X.L. Tu., Z.F. Zhang and X.P. Xia
641 for analytical assistance and Y.F. Deng for helpful discussions. Comments and suggestions from
642 two anonymous reviewers and the Editor help to greatly improve the manuscript. This study was
643 supported by the National Science Foundation for Outstanding Youth (Grant 41525006) and the
644 Strategic Priority Research Program (B) of the Chinese Academy of Sciences (Grant XDB
645 18000000) to F. Guo., and an Australian Research Council Future Fellowship (FT140100826) to
646 X.C. Wang.

647

648 **References**

649 Agrusta, R., Goes, S., van Hunen, J., 2017. Subducting-slab transition-zone interaction:
650 stagnation, penetration and mode switches. *Earth Planet. Sci. Lett.* 464, 10–23.

651 Bolfan-Casanova, N., Mackwell, S., Keppler, H., McCammon, C., Rubie, D.C., 2002. Pressure

652 dependence of H solubility in magnesiowustite up to 25 GPa: implications for the storage of
653 water in the Earth's lower mantle. *Geophys. Res. Lett.* 29, 1029–1032.

654 Bolfan-Casanova, N., Keppler, H., Rubie, D.C., 2003. Water partitioning at 660 km depth and
655 evidence for very low water solubility in magnesium silicate perovskite. *Geophys. Res.*
656 *Lett.* 30. <http://dx.doi.org/10.1029/2003GL017182>

657 Chauvel, C., Hofmann, A.W., Vidal, P., 1992. HIMU-EM: the French-Polynesian connection.
658 *Earth Planet. Sci. Lett.* 110, 99–119.

659 Chen, J.H., 2016. Lower-mantle materials under pressure. *Science* 351, 122–123.

660 Chen, Y., Provost, A., Schiano, P., Cluzel, N., 2011. The rate of water loss from olivine-hosted
661 melt inclusions. *Contrib. Miner. Petrol.* 162, 625–636.

662 Conrad, C.P., Bianco, T.A., Smith, E.I., Wessel, P., 2011. Patterns of intraplate volcanism
663 controlled by asthenospheric shear. *Nat. Geosci.* 4, 317–321.

664 Danyushevsky, L.V., Della-Pasqua, F.N., Sokolov, S., 2000. Re-equilibration of melt inclusions
665 trapped by magnesian olivine phenocrysts from subduction-related magmas: petrological
666 implications. *Contrib. Miner. Petrol.* 138, 68–83.

667 Danyushevsky, L., Plechov, P., 2011. Petrolog3: Integrated software for modelling crystallization
668 processes. *Geochem. Geophys. Geosyst.* 12. <http://dx.doi.org/10.1029/2011GC003516>.

669 Dixon, J.E., Leist, L., Langmuir, C., Schilling, J.G., 2002. Recycled dehydrated lithosphere
670 observed in plume-influenced mid-ocean-ridge basalt. *Nature* 420, 385–389.

671 Faccenna, C., Becker, T.W., Lallemand, S., Lagabrielle, Y., Funiciello, F., Piromallo, C., 2010.
672 Subduction-triggered magmatic pulses: a new class of plumes? *Earth Planet. Sci. Lett.* 299,
673 54–68.

674 Gaetani, G.A., O'Leary, J.A., Shimizu, N., Bucholz, C.E., Newville, M., 2012. Rapid

675 reequilibration of H₂O and oxygen fugacity in olivine-hosted melt inclusions. *Geology* 40,
676 915–918.

677 Gao, S., Yang, J., Zhou, L., Li, M., Hu, Z., Guo, J., Yuan, H., Gong, H., Xiao, G., Wei, J., 2011.
678 Age and growth of the Archean Kongling terrain, South China, with emphasis on 3.3 Ga
679 granitoid gneisses. *American Journal of Science* 311, 153–182.

680 Guo, F., Guo, J.T., Wang, C.Y., Fan, W.M., Li, C.W., Zhao, L., Li, H.X., Li, J.Y., 2013.
681 Formation of mafic magmas through lower crustal AFC processes: an example from the
682 Jinan gabbroic intrusion in the North China Block. *Lithos* 179, 157–174.

683 Hanyu, T., Tatsumi, Y., Senda, R., Miyazaki, T., Chang, Q., Hirahara, Y., Takahashi, T.,
684 Kawabata, H., Suzuki, K., Kimura, J.I., Nakai, S., 2011. Geochemical characteristics and
685 origin of the HIMU reservoir: a possible mantle plume source in the lower mantle.
686 *Geochem. Geophys. Geosyst.* 12. <http://dx.doi.org/10.1029/2010GC003252>.

687 Hauri, E., Wang, J.H., Dixon, J.E., King, P.L., Mandeville, C., Newman, S., 2002a. SIMS
688 analysis of volatiles in silicate glasses: 1. calibration, matrix effects and comparisons with
689 FTIR. *Chem. Geol.* 183, 99–114.

690 Hauri, E.H., 2002b. SIMS analysis of volatiles in silicate glasses, 2: isotopes and abundances in
691 Hawaiian melt inclusions. *Chem. Geol.* 183, 115–141.

692 Hauri, E.H., Gaetani, G.A., Green, T.H., 2006. Partitioning of water during melting of the Earth's
693 upper mantle at H₂O-undersaturated conditions. *Earth Planet. Sci. Lett.* 248, 715–734.

694 Herzberg, C., 2011. Identification of Source Lithology in the Hawaiian and Canary Islands:
695 implications for Origins. *J. Petrol.* 52, 113–146.

696 Hofmann, A.W., 1997. Mantle geochemistry: the message from oceanic volcanism. *Nature* 385,
697 219–229.

698 Hoernle, K., White, J.D.L., van den Bogaard, P., Hauff, F., Coombs, D.S., Werner R., Timm, C.,
699 Garbe-Schonberg, D., Reay, A., Cooper, A.F., 2006. Cenozoic intraplate volcanism on New
700 Zealand: Upwelling induced by lithospheric removal. *Earth Planet. Sci. Lett.* 248, 350–367.

701 Hu, S., Lin, Y.T., Zhang, J.C., Hao, J.L., Yang, W., Deng, L.W., 2015. Measurements of water
702 content and D/H ratio in apatite and silicate glasses using a NanoSIMS 50L. *J. Anal. At.*
703 *Spectrom.* 30, 967–978.

704 Hu, S., Lin, Y.T., Zhang, J.C., Hao, J.L., Yamaguichi, A., Zhang, T., Yang, W., Changela, H.,
705 2019. Volatiles in the martian crust and mantle: Clues from the NWA 6162 shergottite. *Earth*
706 *Planet. Sci. Lett.* <http://doi.org/10.1016/j.epsl.2019.115902>.

707 Huang, X.G., Xu, Y.S., Karato, S.I., 2005. Water content in the transition zone from electrical
708 conductivity of wadsleyite and ringwoodite. *Nature* 434, 746–749.

709 Ishikawa, A., Senda, R., Suzuki, K., Dale, C.W., Meisel, T., 2014. Re-evaluation of digestion
710 methods for highly siderophile element and ¹⁸⁷Os analysis: evidence from geological
711 reference materials. *Chem. Geol.* 384, 27–46.

712 Karato, S.I., 2010. Water distribution across the mantle transition zone and its implications for
713 global material circulation. *Earth Planet. Sci. Lett.* 301, 413–423.

714 Klemme, S., Blundy, J.D., Wood, B.J., 2002. Experimental constraints on major and trace
715 element partitioning during partial melting of eclogite. *Geochim. Cosmochim. Acta* 66,
716 3109–3123.

717 Kogiso, T., Hirschmann, M.M., 2006. Partial melting experiments of bimineraleclogite and the
718 role of recycled mafic oceanic crust in the genesis of ocean island basalts. *Earth Planet. Sci.*
719 *Lett.* 249, 188–199.

720 Kogiso, T., Tatsumi, Y., Shimoda, G., Barszczus, H.G., 1997a. High μ (HIMU) ocean island

721 basalts in southern Polynesia: new evidence for whole mantle scale recycling of subducted
722 oceanic crust. *J. Geophys. Res.* 102, 8085–8103.

723 Kogiso, T., Tatsumi, Y., Nakano, S., 1997b. Trace element transport during dehydration processes
724 in the subducted oceanic crust: 1. experiments and implications for the origin of ocean
725 island basalts. *Earth Planet. Sci. Lett.* 148, 193–205.

726 Kuritani, T., Kimura, J.I., Ohtani, E., Miyamoto, H., Furuyama, K., 2013. Transition zone origin
727 of potassic basalts from Wudalianchi volcano, northeast China. *Lithos* 156, 1–12.

728 Kuritani, T., Ohtani, E., Kimura, J.I., 2011. Intensive hydration of the mantle transition zone
729 beneath China caused by ancient slab stagnation. *Nat. Geosci.* 4, 713–716.

730 Kuritani, T., Xia, Q.K., Kimura, J., Liu, J., Shimizu, K., Ushikubo, T., Zhao, D.P., Nakagawa,
731 M., Yoshimura, S., 2019. Buoyant hydrous mantle plume from the mantle transition zone.
732 *Scientific Reports* 9. <http://dx.doi.org/10.1038/s41598-019-43103-y>.

733 Li, J., Jiang, X.Y., Xu, J.F., Zhong, L.F., Wang, X.C., Wang, G.Q., Zhao, P.P., 2014.
734 Determination of Platinum-Group Elements and Re-Os Isotopes using ID-ICP-MS and N-
735 TIMS from a Single Digestion after Two-Stage Column Separation. *Geostand. Geoanal.*
736 *Res.* 38, 37–50.

737 Li, X.H., Chung, S.L., Zhou, H.W., Lo, C.H., Liu, Y., Chen, C.H., 2004. Jurassic intraplate
738 magmatism in southern Hunan-eastern Guangxi: $^{40}\text{Ar}/^{39}\text{Ar}$ dating, geochemistry, Sr–Nd
739 isotopes and implication for tectonic evolution of SE China. In: Malpas, J., Fletcher, C.J.,
740 Aitchison, J.C., Ali, J. (Eds.), *Aspects of the tectonic evolution of China: Geological Society*
741 *Special Publication*, 226, pp. 193–216.

742 Li, X.H., Li, W.X., Li, Z.X., Lo, C.H., Wang, J., Ye, M.F., Yang, Y.H., 2009. Amalgamation
743 between the Yangtze and Cathaysia Blocks in South China: constraints from SHRIMP U–Pb

744 zircon ages, geochemistry and Nd–Hf isotopes of the Shuangxiwu volcanic rocks.
745 *Precamb. Res.* 174, 117–128.

746 Li, X.H., Li, Z.X., He, B., Li, W.X., Li, Q.L., Gao, Y.Y., Wang, X.C., 2012. The Early Permian
747 active continental margin and crustal growth of the Cathaysia Block: In situ U–Pb, Lu–Hf
748 and O isotope analyses of detrital zircons. *Chem. Geol.* 328, 195–207.

749 Litasov, K.D., Shatskiy, A., Ohtani, E., 2014. Melting and subsolidus phase relations in peridotite
750 and eclogite systems with reduced C–O–H fluid at 3–16 GPa. *Earth Planet. Sci. Lett.* 39, 87–
751 99.

752 Liu, C.Z., Liu, Z.C., Wu, F.Y., Chu, Z.Y., 2012. Mesozoic accretion of juvenile sub-continental
753 lithospheric mantle beneath South China and its implications: geochemical and Re–Os
754 isotopic results from Ningyuan mantle xenoliths. *Chem. Geol.* 291, 186–198.

755 Matthey, D., Lowry, D., Macpherson, C., 1994. Oxygen isotope composition of mantle peridotite.
756 *Earth Planet. Sci. Lett.* 128, 231–241.

757 McKeegan, K.D., Aleon, J., Bradley, J., Brownlee, D., Busemann, H., Butterworth, A.,
758 Chaussidon, M., Fallon, S., Floss, C., Gilmour, J., 2006. Isotopic compositions of cometary
759 matter returned by Stardust. *Science* 314, 1724–1728.

760 Meier, U., Trampert, J., Curtis, A., 2009. Global variations of temperature and water content in
761 the mantle transition zone from higher mode surface waves. *Earth Planet. Sci. Lett.* 282,
762 91–101.

763 Meibom, A., Sleep, N.H., Chamberlain, C.P., Coleman, R.G., Frei, R., 2002. Re–Os isotopic
764 evidence for long-lived heterogeneity and equilibration processes in the Earth's upper
765 mantle. *Nature* 419, 705–708.

766 Michael, P., 1995. Regionally distinctive sources of depleted MORB: evidence from trace

767 elements and H₂O. *Earth Planet. Sci. Lett.* 131, 301–320.

768 Miyagi, I., Matsubaya, O., 2003. Hydrogen isotopic composition of hornblende and biotite
769 phenocrysts from Japanese Island arc volcanoes; evaluation of alteration process of the
770 hydrogen isotopic ratios by degassing and re-equilibration. *J. Volcanol. Geotherm. Res.* 126,
771 157–168.

772 Murphy, D.T., Collerson, K.D., Kamber, B.S., 2002. Lamproites from Gaussberg, Antarctica:
773 possible transition zone melts of Archaean subducted sediments. *J. Petrol.* 43, 981–1001.

774 Nebel, O., Arculus, R.J., van Westrenen, W., Woodhead, J.D., Jenner, F.E., Nebel-Jacobsen, Y.J.,
775 Wille, M., Eggins, S.M., 2013. Coupled Hf-Nd-Pb isotope co-variations of HIMU oceanic
776 island basalts from Mangaia, Cook-Austral islands, suggest an Archean source component
777 in the mantle transition zone. *Geochim. Cosmochim. Acta* 112, 87–101.

778 Pamato, M.G., Myhill, R., Ballaran, T.B., Frost, D.J., Heidelbach, F., Miyajima, N., 2015.
779 Lower-mantle water reservoir implied by the extreme stability of a hydrous aluminosilicate.
780 *Nat. Geosci.* 8, 75–79.

781 Panter, K.S., Blusztajn, J., Hart, S.R., Kyle, P.R., Esser, R., McIntosh, W.C., 2006. The origin of
782 HIMU in the SW Pacific: evidence from intraplate volcanism in southern New Zealand and
783 subantarctic islands. *J. Petrol.* 47, 1673–1704.

784 Pearson, D.G., Brenker, E.E., Nestola, F., McNeill, J., Nasdala, L., Hutchison, M.T., Matveev, S.,
785 Mather, K., Silversmit, G., Schmitz, S., 2014. Hydrous mantle transition zone indicated by
786 ringwoodite included within diamond. *Nature* 507, 221–224.

787 Pertermann, M., Hirschmann, M.M., Hametner, K., Gunther, D., Schmidt, M.W., 2004.
788 Experimental determination of trace element partitioning between garnet and silica-rich
789 liquid during anhydrous partial melting of MORB-like eclogite. *Geochem. Geophys.*

790 Geosyst. 5. <http://dx.doi.org/10.1029/2003GC000638>.

791 Peucker-Ehrenbrink, B., Hanghoj, K., Atwood, T., Kelemen, P.B., 2012. Rhenium–osmium
792 isotope systematics and platinum group element concentrations in oceanic crust. *Geology*
793 40, 199–202.

794 Pilet, S., Baker, M.B., Stolper, E.M., 2008. Metasomatized lithosphere and the origin of alkaline
795 lavas. *Science* 320, 916–919.

796 Poli, S., Schmidt, M.W., 2002. Petrology of subducted slabs. *Annu. Rev. Earth Planet. Sci.* 30,
797 207–235.

798 Reisberg, L., Meisel, T., 2002. The Re–Os isotopic system: a review of analytical techniques.
799 *Geostand. Newslett.* 26, 249–267.

800 Ringwood, A.E., 1994. Role of the transition zone and 660 km discontinuity in mantle dynamics.
801 *Phys. Earth Planet. Inter.* 86, 5–24.

802 Saal, A.E., Hauri, E.H., Langmuir, C.H., Perfit, M.R., 2002. Vapour undersaturation in primitive
803 mid-ocean-ridge basalt and the volatile content of Earth's upper mantle. *Nature* 419, 451–
804 455.

805 Shaw, A.M., Hauri, E.H., Fischer, T.P., Hilton, D.R., Kelley, K.A., 2008. Hydrogen isotopes in
806 Mariana arc melt inclusions: implications for subduction dehydration and the deep-Earth
807 water cycle. *Earth Planet. Sci. Lett.* 275, 138–145.

808 Shaw, A.M., Hauri, E.H., Behn, M.D., Hilton, D.R., Macpherson, C.G., Sinton, J.M., 2012.
809 Long-term preservation of slab signatures in the mantle inferred from hydrogen isotopes.
810 *Nat. Geosci.* 5, 224–228.

811 Sobolev, A.V., Hofmann, A.W., Kuzmin, D.V., Yaxley, G.M., Arndt, N.T., et al., 2007. The
812 amount of recycled crust in sources of mantle-derived melts. *Science* 316, 412–417.

813 Sobolev, A.V., Hofmann, A.W., Sobolev, S.V., Nikogosian, I.K., 2005. An olivine-free mantle
814 source of Hawaiian shield basalts. *Nature* 434, 590–597.

815 Stracke, A., Bizimis, M., Salters, V.J.M., 2003. Recycling oceanic crust: quantitative constraints.
816 *Geochem. Geophys. Geosyst.* 4. <http://dx.doi.org/10.1029/2001GC000223>.

817 Sun, S.S., McDonough, W.F., 1989. Chemical and isotopic systematics of oceanic basalts:
818 implication for mantle composition and processes. In *Magmatism in the Ocean Basins*, 42
819 (eds. A. D. Saunderson, M. J. Norry). Geological Society of London, Special Publication,
820 London, pp. 313–345.

821 Thirlwall, M.F., 1997. Pb isotopic and elemental evidence for OIB derivation from young HIMU
822 mantle. *Chem. Geol.* 139, 51–47.

823 Thompson, R.N., Ottley, C.J., Smith, P.M., Pearson, D.G., Dickin, A.P., Morrison, M.A., Leat,
824 P.T., Gibson, S.A., 2005. Source of the quaternary alkalic basalts, picrites and basanites of
825 the Potrillo Volcanic Field, New Mexico, USA: lithosphere or convecting mantle? *J. Petrol.*
826 46, 1603–1643.

827 van der Meer, Q.H.A., Waight, T.E., Scott, J.M., Munker, C., 2017. Variable sources for
828 Cretaceous to recent HIMU and HIMU-like intraplate magmatism in New Zealand. *Earth*
829 *Planet. Sci. Lett.* 469, 27–41.

830 Vinnik, L., Farra, V., 2007. Low S velocity atop the 410-km discontinuity and mantle plumes.
831 *Earth Planet. Sci. Lett.* 262, 398–412.

832 Wang, X.C., Wilde, S.A., Li, Q.L., Yang, Y.N., 2015. Continental flood basalts derived from the
833 hydrous mantle transition zone. *Nat. Commun.* 6. <http://dx.doi.org/10.1038/ncomms8700>.

834 Wang, X.C., Wilde, S.A., Xu, B., Pang, C.J., 2016. Origin of arc-like continental basalts:
835 Implications for deep-Earth fluid cycling and tectonic discrimination. *Lithos* 261, 5–45.

- 836 Wang, X.J., Chen, L.H., Hofmann, A.W., Mao, F.G., Liu, J.Q., Zhong, Y., Xie, L.W., Yang, Y.H.,
837 2017. Mantle transition zone-derived EM1 component beneath NE China: geochemical
838 evidence from Cenozoic potassic basalts. *Earth Planet. Sci. Lett.* 465, 16–28.
- 839 Wang, Y.J., Liao, C.L., Fan, W.M., Peng, T.P., 2004. Early Mesozoic OIB-type alkaline basalt in
840 central Jiangxi Province and its tectonic implications. *Geochimica* 33, 109–117 (in Chinese
841 with English abstract).
- 842 Wang, Y.J., Fan, W.M., Cawood, P.A., Li, S.Z., 2008. Sr-Nd-Pb isotopic constraints on multiple
843 mantle domains for Mesozoic mafic rocks beneath the South China Block hinterland. *Lithos*
844 106, 297–308.
- 845 Wang, Z.R., Gaetani, G.A., 2008. Partitioning of Ni between olivine and siliceous eclogite partial
846 melt: experimental constraints on the mantle source of Hawaiian basalts. *Contrib. Miner.*
847 *Petrol.* 155, 661–678.
- 848 Weaver, B.L., 1991. The origin of ocean island basalt end-member compositions-trace element
849 and isotopic constraints. *Earth Planet. Sci. Lett.* 104, 381–397.
- 850 Weis, D., Kieffer, B., Maerschalk, C., Pretorius, W., Barling, J., 2005. High-precision Pb-Sr-Nd-
851 Hf isotopic characterization of USGS BHVO-1 and BHVO-2 reference materials. *Geochem.*
852 *Geophys. Geosyst.* 6. <http://dx.doi.org/10.1029/2004GC000852>
- 853 Weiss, Y., Class, C., Goldstein, S.L., Hanyu, T., 2016. Key new pieces of the HIMU puzzle from
854 olivines and diamond inclusions. *Nature* 537, 666–670.
- 855 Xia, Q.K., Liu, J., Kovács, I., Hao, Y.T., Li, P., Yang, X.Z., Chen, H., Sheng, Y.M., 2019. Water
856 in the upper mantle and deep crust of eastern China: concentration, distribution and
857 implications. *Nat. Sci. Rev.* 6, 125–144.
- 858 Xu, Z., Zheng, Y.F., Zhao, Z.F., Gong, B., 2014. The hydrous properties of subcontinental

859 lithospheric mantle: Constraints from water content and hydrogen isotope composition of
860 phenocrysts from Cenozoic continental basalts in North China. *Geochim. Cosmochim. Acta*
861 143, 285–302.

862 Yu, J., O'Reilly, Y.S., Wang, L., Griffin, W.L., Jiang, S., Wang, R., Xu, X., 2007. Finding of
863 ancient materials in Cathaysia and implication for the formation of Precambrian crust.
864 *Chinese Sci. Bull.* 52, 13–22.

865 Zeng, G., Chen, L.H., Xu, X.S., Jiang, S.Y., Hofmann, A.W., 2010. Carbonated mantle sources
866 for Cenozoic intra-plate alkaline basalts in Shandong, North China. *Chem. Geol.* 273, 35–
867 45.

868 Zeng, G., He, Z.Y., Li, Z., Xu, X.S., Chen, L.H., 2016. Geodynamics of paleo-Pacific plate
869 subduction constrained by the source lithologies of Late Mesozoic basalts in southeastern
870 China. *Geophys. Res. Lett.* 143, 10189–10197.

871 Zhang, H.F., Goldstein, S.L., Zhou, X.H., Sun, M., Zheng, J.P., Cai, Y., 2008. Evolution of
872 subcontinental lithospheric mantle beneath eastern China: Re-Os isotopic evidence from
873 mantle xenoliths in Paleozoic kimberlites and Mesozoic basalts. *Contrib. Miner. Petrol.* 155,
874 271–293.

875 Zhou, X.M., Sun, T., Shen, W.Z., Shu, L.S., Niu, Y.L., 2006. Petrogenesis of Mesozoic granitoids
876 and volcanic rocks in South China: a response to tectonic evolution. *Episodes* 29, 26–33.

877 Zindler, A., Hart, S., 1986. Chemical geodynamics. *Annu. Rev. Earth Planet. Sci.* 14, 493–571.

878 **Figure captions**

879 **Figure 1.** A geologic map showing the distributions of Mesozoic magmatic rocks and major
880 faults in Southeast China (Zhou et al. 2006; Wang et al., 2008). Inset figure shows a simplified
881 tectonic map of China with a speculated subduction zone of the Paleo-Pacific slab in the late
882 Mesozoic. The Antang basalt is located ~800 km west of the paleo-Pacific subduction zone.

883 **Figure 2.** Petrography of the Antang basalts and the melt inclusions in olivine. (a)
884 Photomicrograph of olivine basalt under crossed polarized light. (b) Olivine phenocryst in (a)
885 highlighting melt inclusion under reflected light. (c) A homogenized melt inclusion in olivine. (d)
886 and (e) Original (unhomogenized) melt inclusions in olivine. (f) Enlarged melt inclusion shown
887 in (e). Abbreviations: Ol: olivine; Cpx: clinopyroxene; MI: melt inclusion; Pl: plagioclase.

888 **Figure 3.** Primitive mantle-normalized trace element spidergram of bulk rocks (red lines in both
889 a and b) and melt inclusions (yellow lines in b) of the Antang basalts, and the melts derived from
890 carbonated peridotite (Zeng et al., 2010), hornblendite, clinopyroxene-hornblendite, and
891 hornblendite-peridotite (Pilet et al., 2008). HIMU basalts from St. Helena (Kawabata et al., 2011)
892 are also plotted for comparison. Normalized values for primitive mantle are from Sun and
893 McDonough (1989).

894 **Figure 4.** The $\text{CaO}/\text{Al}_2\text{O}_3$ value of melt inclusion versus the Fo number of host olivines.

895 **Figure 5.** Sr-Nd-Pb-Hf isotopic comparison between the Antang basalts and typical HIMU
896 basalts, modelling the evolution of recycled oceanic crust plus contemporary sediment (Stracke
897 et al., 2003). The solid blue line with solid circles shows the isotopic growth of recycled oceanic
898 crust. The black lines with crosses connect the recycled oceanic crust and the subducted sediment
899 of various ages (0–2.5 Ga). The numbers on the curves denote the mass fractions of sediment in

900 the mixed source. The isotopic compositions of St. Helena and Cook-Austral HIMU basalts are
901 compiled from the GEOROC Database (<http://georoc.mpch-mainz.gwdg.de/georoc/>). The typical
902 HIMU basalts with $^{206}\text{Pb}/^{204}\text{Pb} > 20.5$ can be interpreted as melts from 2.0–1.5 Ga recycled
903 oceanic crust, whereas those with $19.0 < ^{206}\text{Pb}/^{204}\text{Pb} < 20.5$ are related to younger recycled
904 oceanic crust (1.0–1.5 Ga) and/or to the addition of minor sediment (0–1%) to the source. The
905 isotopic composition of the Antang basalts indicates their derivation from a mantle source
906 comprising mainly 1.0–1.5 Ga recycled oceanic crust plus minor sediment. The data for other
907 HIMU-like basalts in SE China are from [Wang et al. \(2008\)](#). The detailed calculation process is
908 presented in [Table S8](#).

909 **Figure 6.** (a) Ca (ppm) versus Fo (mol%) of olivine, showing the divisions (blue dashed lines)
910 between carbonated peridotite, peridotite and eclogite/pyroxenite. The purple squares denote
911 olivines of the basalts from Mangaia and Tubuai that were derived from carbonated peridotitic
912 source ([Weiss et al., 2016](#)). The olivine data for MORB and Hawaii are compiled from [Sobolev](#)
913 [et al. \(2007\)](#). (b) Plot of $\delta^{18}\text{O}$ versus Fo of olivine for the Antang basalts. The shaded band
914 denotes the range of normal mantle olivine with $\delta^{18}\text{O}$ between 4.8 and 5.4‰ ([Mattey et al.,](#)
915 [1994](#)).

916 **Figure 7.** CaO (wt%) versus MgO (wt%) plot of primary magmas to discriminate between (blue
917 dashed line) the melts from peridotite and eclogite/pyroxenite. The primary magma compositions
918 of the Antang basalts have been corrected for olivine fractionation ([Table S4c](#)). Data sources for
919 melts of peridotite, SiO₂-excess eclogite, SiO₂-poor eclogite, and eclogite-peridotite hybrid are
920 presented in “[Supporting Information A1](#)”.

921 **Figure 8.** (a) δD versus H_2O of the melt inclusions. The yellow band denotes the ranges for
922 MORBs (Shaw et al., 2008). The range of Koolau melt inclusions is also shown for comparison
923 (Hauri et al., 2002b). (b) δD versus Fo of the host olivine.

924 **Figure 9.** Plots of (a) Rb vs. Zr, (b) Ba vs. Zr, (c) Th vs. Zr, and (d) U vs. Zr for the olivine-
925 hosted melt inclusions.

926 **Figure 10.** Re-Os isotopic systematics of the Antang basalts. (a) Re versus Os compared with
927 typical HIMU basalts from Cook-Austral (Hanyu et al., 2011) and mantle xenoliths (Liu et al.,
928 2012; Zhang et al., 2008) from the contemporaneous Ningyuan basalts in South China. (b) The
929 positive correlation between $^{187}Os/^{188}Os$ and $^{187}Re/^{188}Os$ indicates a nugget effect or dis-
930 equilibrium binary mixing (Meibom et al., 2002). (c) $^{87}Sr/^{86}Sr(i)$ versus $^{187}Os/^{188}Os(t)$ for the
931 Antang basalts and the mantle xenoliths. The broad negative correlation between $^{87}Sr/^{86}Sr(i)$ and
932 $^{187}Os/^{188}Os(t)$ argues against a role for continental crust contamination. (d) The binary mixing
933 lines (curves 1, 2 and 3) for melts derived from lithospheric mantle and crust. We assume that
934 melts derived from lithospheric mantle have $Os = 250$ ppt and $^{187}Os/^{188}Os$ (168 Ma) = 0.1231,
935 and that the continental crust has $^{187}Os/^{188}Os = 1.29$ and an Os concentration of 15 ppt in curve 1,
936 5 ppt in curve 2 and 1 ppt in curve 3.

937 **Figure 11.** CIPW normalization plotted in Ne + Lc (nepheline + leucitite), Ol (olivine), Di
938 (diopside), Hy (hypersphene), Q (quartz) space (Thompson et al., 2005). Melt compositions in a
939 range of partial melting experiments on various sources, including: hornblendite, cpx-
940 hornbleIndite, carbonated peridotite, carbonated eclogite, silica-deficient pyroxenite, silica-rich
941 pyroxenite, and peridotite. All data were calculated using the assumption that 10% of the total
942 iron as Fe_2O_3 . The related references are listed in “Supporting Information A1”.

943 **Figure 12.** La/Sm vs. Sm/Yb (a) and La/Yb (b) plots of the Antang basalts, modelling the
944 melting degrees (F) of eclogite. During the calculation, the non-modal batch melting equation
945 was used. The initial composition of the mantle source is assumed to comprise 99% basaltic
946 oceanic crust (25% N-MORB + 25% altered MORB + 50% gabbro) and 1% subducted sediment
947 represented by GLOSS, which has La = 3.88 ppm, Sm = 3.03 ppm, and Yb = 3.02 ppm. The
948 residual assemblage of 75% Cpx + 25% Grt, 80% Cpx + 20% Grt and 82% Cpx + 18% Grt is
949 assumed for Curves 1, 2 and 3, respectively. Numbers on the curves denote the degree of
950 melting. Mineral abbreviations: Grt = garnet; Cpx = clinopyroxene. Data sources for the related
951 parameters used in the calculation can be seen in “[Supporting Information A1](#)”.























