1 Low-³He/⁴He sublithospheric mantle source for the most magnesian magmas of the
2 Karoo large igneous province

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11
12 **Abstract**

13 The massive outpourings of Karoo and Ferrar continental flood basalts (CFBs) ~180 Ma ago mark
14 the initial Jurassic rifting stages of the Gondwana supercontinent. The origin and sources of these
15 eruptions have been debated for decades, largely due to difficulties in defining their parental melt
16 and mantle source characteristics. Recent findings of Fe- and Mg-rich dikes (depleted ferropicrite
17 suite) from Vestfjella, western Dronning Maud Land, Antarctica, have shed light on the
18 composition of the deep sub-Gondwanan mantle: these magmas have been connected to upper
19 mantle sources presently sampled by the Southwest Indian Ocean mid-ocean ridge basalts (SWIR
20 MORBs) or to high ³He/⁴He plume-entrained non-chondritic primitive mantle sources formed early
21 in Earth's history. In an attempt to determine their He isotopic composition and relative
22 contributions from magmatic, cosmogenic, and radiogenic He sources, we performed in-vacuo
23 stepwise crushing and melting analyses of olivine mineral separates, some of which were abraded to
24 remove the outer layer. The best estimate for the mantle isotopic composition is given by a sample
25 with the highest amount of He released (> 50%) during the first crushing step of an abraded coarse

fraction. It has a $^3\text{He}/^4\text{He}$ of 7.03 ± 0.23 (2σ) times the atmospheric ratio (R_a), which is indistinguishable from those measured from SWIR MORBs ($6.3\text{--}7.3 R_a$; source $^3\text{He}/^4\text{He} \sim 6.4\text{--}7.6 R_a$ at 180 Ma) and notably lower than in the most primitive lavas from the North Atlantic Igneous Province (up to $50 R_a$), considered to represent the epitome magmas from non-chondritic primitive mantle sources. Previously published trace element and isotopic (Sr, Nd, and Pb) compositions do not suggest a direct genetic link to any modern hotspot of Indian or southern Atlantic Oceans. Although influence of a mantle plume cannot be ruled out, the high magma temperatures and SWIR MORB-like geochemistry of the suite are best explained by supercontinent insulation of a precursory Indian Ocean upper mantle source. Such a model is also supported by the majority of the recent studies on the structure, geochronology, and petrology of the Karoo CFBs.

Keywords: Karoo; large igneous province; continental flood basalts; helium isotopes; mantle source

1. Introduction

The Karoo continental flood basalts (CFBs) erupted between the landmasses of modern Africa and Antarctica when they were a part of the Gondwana supercontinent ~ 180 Ma ago (e.g., Jourdan et al., 2005). The Karoo CFBs, and temporally and spatially related Ferrar CFBs (Fig. 1; often called the Karoo-Ferrar province), exhibit strongly lithosphere-influenced geochemical characteristics (e.g., Hawkesworth et al., 1984; Sweeney et al., 1994; Molzahn et al., 1996; Luttinen and Furnes, 2000; Jourdan et al., 2007a; Neumann et al., 2011). This has led to difficulties in constraining their primary melt and mantle source characteristics, and, therefore, in defining their ultimate origins.

Recently, a group of rare Mg-rich dikes, the Vestfjella depleted ferropicrite suite, was described from the Vestfjella mountain range that forms an Antarctic extension of the Karoo CFB province (Fig. 2; Heinonen and Luttinen, 2008). These dikes exhibit no evidence of lithospheric

51 sources or contamination and thus provide important information on the composition of the
52 sublithospheric mantle during Karoo magmatism. Their MgO-rich (> 20 wt. %) parental magmas
53 require that a positive thermal anomaly existed in the upper mantle during their generation
54 (Heinonen and Luttinen, 2010). On the basis of the major element, trace element, and isotopic (Sr,
55 Nd, Pb, and Os) composition of the dikes, their origins have been linked to high-pressure (5–6 GPa)
56 melting of hot asthenospheric mantle that has since cooled and is presently sampled by the modern
57 mid-ocean ridge basalts of the Southwest Indian Ridge (SWIR MORBs; Heinonen et al., 2010), the
58 modern successor of the Africa-Antarctica rift zone (Fig. 1). The isotopic compositions also overlap
59 with non-chondritic primitive mantle reservoir recently suggested to be a major component in CFB
60 plume sources (Jackson and Carlson, 2011).

61 Helium isotopes are a powerful indicator of ancient relatively undegassed mantle sources,
62 having low (Th+U)/He and high $^3\text{He}/^4\text{He}$ (up to 50 times the atmospheric ratio, i.e., 50 R_a), and
63 often referred to as primitive or primordial (e.g., Lupton and Craig, 1975; Kurz et al., 1982).
64 Although such sources may be considered controversial in the context of whole mantle convection
65 models, most of the models suggest a lower mantle origin (e.g., Class and Goldstein, 2005). High
66 $^3\text{He}/^4\text{He}$ have been found in several hotspot-related rocks (e.g., Hawaii, Galapagos, and Iceland),
67 and in the high-Mg rocks from the North Atlantic Volcanic Province (NAIP; Graham et al., 1998;
68 Stuart et al., 2003; Starkey et al., 2009), which were used to estimate the trace element composition
69 of a non-chondritic primitive mantle source (Jackson et al., 2010). In contrast, MORBs show
70 relatively low $^3\text{He}/^4\text{He}$ values on average ($\sim 8 \pm 1 R_a$; Kurz and Jenkins, 1981; Graham, 2002), with
71 SWIR MORBs on the low end of the MORB spectrum ($^3\text{He}/^4\text{He} = 6.3\text{--}7.3 R_a$; Georgen et al.,
72 2003).

73 Helium isotopic compositions have not previously been reported for the Karoo CFBs or
74 related rocks. This is due to the scarcity of primitive samples with fresh olivine or glass containing
75 primary magmatic gas inclusions that would be suitable for He measurements, along with the

76 Jurassic age of the samples. Most existing global He data are from zero age basalts, which
77 minimizes the complexities of post-crystallization addition of nucleogenic (produced mainly by
78 ${}^6\text{Li}(n,\alpha){}^3\text{H}(\beta^-)$ reactions), cosmogenic (from spallation near the Earth's surface), and radiogenic
79 (generated by α -decay of U and Th) He, all of which can overprint the magmatic signature (see
80 Kurz, 1986; Graham et al., 1998).

81 Here we present the first He isotopic data for rocks related to the Karoo CFBs (Vestfjella
82 depleted ferropicrite suite). In order to estimate the effects of possible cosmogenic or radiogenic He,
83 some of the carefully hand-picked olivine fractions were abraded (to remove the outer layer at the
84 groundmass/phenocryst interface), step crushed in vacuum, and melted. In addition to He data, our
85 analysis of the ultimate mantle source of the Vestfjella depleted ferropicrite suite is corroborated by
86 detailed isotopic and trace element comparisons with relevant hotspot-related rocks.

87

88 **2. Sample selection and analytical methods**

89

90 *2.1. Samples*

91

92 Five ferropicritic and meimechitic dike rock samples ($\text{MgO} = 17\text{--}24$ wt. %; $\text{FeO}^{\text{tot}} = 13\text{--}15$ wt. %)
93 representing the well-characterized $\sim 185.5 \pm 1.8$ Ma (${}^{40}\text{Ar}/{}^{39}\text{Ar}$ plateau age; Luttinen et al., in press)
94 depleted ferropicrite suite of Vestfjella (Heinonen and Luttinen, 2008), western Dronning Maud
95 Land, were selected for He isotopic analysis. The dikes were emplaced through a heterogeneous
96 crust consisting of early-phase Karoo CFBs, Permian sedimentary rocks, and a crystalline basement
97 of Archean to Phanerozoic age (Fig. 2a; see Groenewald et al., 1995). The samples were selected
98 for the presence of fresh olivine and the field work emphasis was in obtaining unaltered,
99 representative, samples for whole-rock geochemistry without consideration of in-situ ${}^3\text{He}$
100 production by cosmic rays. All the samples contain olivine phenocrysts that are larger than 1 mm in

101 diameter, and exhibit very fine-grained groundmass indicative of crystallization relatively close to
102 the surface. The coordinates and field characteristics, and detailed petrographical and geochemical
103 descriptions of the samples are presented in Heinonen and Luttinen (2008, 2010) and Heinonen et
104 al. (2010).

105 Four of the samples are from Basen nunatak that is composed dominantly of near-horizontal
106 Karoo CFB flows and reaches the height of 580 m, about 400 m above the present ice surface ($h \approx$
107 200 m; Fig. 2b). It exhibits a vertical ~ 300 m cliff on its northwestern ($\sim$ distal) side and steeply
108 dipping partially ice- and snow-covered slope on its southeastern ($\sim$ proximal) side. Sample
109 AL/B16-98 represents an in-situ weathered dike from the exposed northeast corner of Basen ($h \approx$
110 480 m). Dike-derived boulders AL/B1b-03, AL/B7-03, and AL/B9-03 were sampled from the
111 coarse glacial drift deposits ($h \approx 450\text{--}470$ m) in the ‘backyard’ of the Aboa and Wasa research
112 stations close to the edge of the glacier on the southwestern slope of the nunatak (Fig. 2b). The
113 boulder sampled by AL/B1b-03 exhibits chilled margin against sandstone that is likely to represent
114 Permian sedimentary rock formations underlying the Vestfjella CFBs. It indicates that the boulders
115 have been glacially transported on top of the nunatak from the valleys that surround it.

116 Sample AL/WM1e-98 was collected from an in-situ weathered dike ($h \approx 600$ m) at West-
117 Muren, a ridge-like nunatak, composed of Karoo CFBs and related intrusive rocks and exposed
118 ~ 100 km southwest from Basen (Fig. 2c). The nunatak stands out only few meters above the ice
119 surface ($h \approx 600$ m) on its northwestern ($\sim$ proximal) side whereas its southeastern ($\sim$ distal) side dips
120 steeply towards the ice surface on the bottom of the ‘Muren valley’ ($h \approx 500$ m).

121

122 2.2. Analytical methods

123

124 The samples were crushed multiple times with jaw crusher and several size fractions were separated
125 from the crushed bulk sample with sieves. Although the largest olivine phenocrysts recorded on the

126 basis of petrographical studies were over 4 mm in diameter, fractions with grain size of more than 1
127 mm did not contain pure olivine grains. In most samples, 0.25–0.5 mm fraction was found to be
128 optimum in terms of pure olivine that was still large enough for hand-picking. The relatively small
129 grain size of pure olivine meant that at least ~1000 grains per sample had to be hand-picked to
130 obtain enough material for abrasion and step-crushing analyses (see below). Earlier petrographical
131 studies had revealed that many olivine grains contain large crystallized melt inclusions ($\varnothing$ up to 0.8
132 mm) with igneous amphibole (Heinonen and Luttinen, 2010), but preliminary SEM-screenings of
133 some of the fractions indicated that the olivines likely broke along these inclusions during crushing.
134 The size and amount of the melt inclusions that survived the crushing are small ($\varnothing < 0.05$ mm, ~5
135 inclusions per 100 grains in a cross-section).

136 In order to estimate the effect of radiogenic ^4He implanted to the olivines from the
137 surrounding groundmass after crystallization, at least one fraction from each sample was abraded to
138 remove the outer portions of the grains. The stainless steel abrasion chamber was of the design
139 traditionally used for zircon abrasion (Krogh, 1982), using pure nitrogen gas as the propellant. The
140 fractions were abraded as long as needed to achieve a mass loss of 25–40% (depending on the grain
141 size of the fraction), resulting in the removal of 20 μm thick ^4He -contaminated rims of the grains
142 and uniform oval to round grain shapes (see Aciego et al., 2007). In addition to olivine fractions,
143 one fraction consisting entirely of groundmass was separated from boulder sample AL/B9-03
144 (without abrasion).

145 The He isotopic measurements were performed at Woods Hole Oceanographic Institution
146 (WHOI) using a dual collection and statically operated He isotope mass spectrometer that has a
147 Nier-type ion source (locally referred to as MS2). The instrument and the procedure have been
148 described many times in the literature, e.g., by Kurz et al. (2004). The ^4He blank was always below
149 $3 \cdot 10^{-11} \text{ cm}^3$ (STP) and small relative to the gas released from the samples. In basaltic phenocryst
150 samples, He released by crushing in vacuum fractions is typically dominated by gas and melt

151 inclusions (e.g., Kurz et al., 1996). The Vestfjella samples are intrusive, however, and may not have
152 the same gas geochemistry. In most cases, the samples were crushed and analyzed in 20-stroke
153 steps to evaluate the efficiency of gas extraction and possible post-eruption contamination
154 (successive steps should increasingly release cosmogenic/radiogenic He from the olivine matrix).
155 After the crushing extraction, the remaining olivine powder was melted in a single heating step and
156 the gas analyzed to determine the bulk He isotopic composition.

157

158 **3. Results and interpretation of the He isotopic measurements**

159

160 *3.1. Results*

161

162 The results of the He measurements are shown in Table 1 and illustrated in Fig. 3. The released He
163 contents vary from 0.001 to 0.2 $\mu\text{cc/g}$ (STP) by crushing and from 0.02 to 19 $\mu\text{cc/g}$ (STP) by
164 melting: the highest amount of He was released from sample AL/WM1e-98 and from the AL/B9-03
165 groundmass sample having radiogenic isotopic signatures ($^3\text{He}/^4\text{He} < 2 R_a$). In the stepwise
166 crushing measurements, the amount of He decreased with every step, except in the case of
167 AL/WM1e-98 and the abraded fractions of AL/B16-98 and AL/B7-03, in all of which the 2nd step
168 exhibited lower He concentrations than the 3rd (last) step. The fraction of total He released by
169 crushing is generally below 40 %, except in the first crushing step of abraded 0.5–1 mm fraction of
170 AL/B1b-03 (51 % of total He released). The relatively low He amount released by crushing may be
171 the result of the rarity of melt inclusions in the olivine fractions (see section 2.2). The total He
172 released by melting of the AL/B9-03 groundmass fraction (4.8 $\mu\text{cc/g}$ (STP)) is within the range of
173 radiogenic ^4He production values (3.8–6.4 $\mu\text{cc/g}$ (STP)) calculated for an age of 180 Ma and using
174 whole-rock U (0.10 ppm) and Th (0.32 ppm) with and without subtraction of ~40 wt.% of olivine

175 phenocrysts (as suggested by the mineral mode of the sample; Heinonen and Luttinen, 2010) having
176 0 ppm of U and Th.

177 The measured $^3\text{He}/^4\text{He}$ varies from 0.14 (± 0.01 , 2σ) R_a to 7.03 (± 0.23) R_a by crushing and
178 from 0.04 (± 0.01) R_a to 15.38 (± 0.47) R_a by melting. The first crushing step of abraded 0.5–1 mm
179 fraction of AL/B1b-03 exhibits the highest crush value of 7.03 (± 0.23) R_a . In the stepwise
180 measurements, the samples that showed anomalous release of He (AL/WM1e-98 and abraded
181 AL/B16-98 and AL/B7-03) also yielded the highest $^3\text{He}/^4\text{He}$ in the 2nd step. Apart from the abraded
182 0.25–1 mm fraction of AL/B1b-03 that was crushed in five steps, the rest of the stepwise
183 measurements gave the highest $^3\text{He}/^4\text{He}$ in the first step, although in the case of AL/B16-98 all
184 crush $^3\text{He}/^4\text{He}$ values are within error.

185

186 3.2. Constraining the magmatic $^3\text{He}/^4\text{He}$ composition

187

188 Determination of magmatic $^3\text{He}/^4\text{He}$ in Jurassic samples is complicated by post-crystallization
189 addition of nucleogenic ^3He from $^6\text{Li}(n,\alpha)^3\text{H}(\beta^-)$, cosmogenic ^3He (by spallation), and radiogenic
190 ^4He (formed by the α -decay of U and Th) from the groundmass (e.g., Graham et al., 1998).

191 Nucleogenic ^3He addition in 180 Ma was calculated following the equations presented by
192 Andrews (1985) and using neutron capture cross-sections provided for 17 elements presented by
193 Andrews and Kay (1982). The neutron capture probabilities and neutron production rates were
194 calculated using whole-rock chemistry of the samples. This treatment yields an upper limit to
195 possible influence of nucleogenic ^3He , because pure olivine has considerably lower U and Th
196 contents compared to the whole-rock values and thus likely smaller α -particle flux for inducing the
197 Li reaction. The calculated effect of nucleogenic ^3He production is generally very small, but in the
198 few cases in which ^3He contents are low (3rd and further crushing steps of AL/B1b-03 fractions), the
199 net effect of estimated maximum nucleogenic ^3He production exceeds the 2σ error of the $^3\text{He}/^4\text{He}$

value (Table 1). In these analyses it is presented as a negative uncertainty (Table 1; Fig. 3), even though nucleogenic ^3He is likely to be released mainly during melting of the crushed powder.

The effect of cosmogenic ^3He addition is difficult to estimate, because all of the sampling sites have experienced at least some exposure to cosmic rays and exposure age constraints are lacking. Although the radiocarbon ages of snow petrel nesting sites at Vestfjella do not exceed 9000 BP (Lintinen and Nenonen, 1997), preliminary ^{10}Be age data indicates that at least some of the outcrops may have been exposed for tens of thousands of years (P. Lintinen, personal communication), and it is not possible to exclude multiple exposures produced by glacial advance and retreat. Highest melt $^3\text{He}/^4\text{He}$ (10–15 R_a) is shown by AL/B16-98 fractions (Table 1), collected from an in-situ weathered dike on Basen nunatak: their $^3\text{He}/^4\text{He}$ has likely been affected by at least some cosmogenic ^3He addition. Assuming initial $^3\text{He}/^4\text{He}$ of 7 R_a (see below) and no radiogenic input of ^4He , and using the mean global sea level high latitude cosmogenic ^3He production rate of 120 atoms $\text{g}^{-1} \text{yr}^{-1}$ (Goehring et al., 2010) and scaling factor of 1.25 for Antarctica (Stone, 2000), the melt R_a value of 15 R_a shown by the abraded AL/B16-98 fraction would have been produced in ~100 ka. Outcrop exposure ages of such magnitude within a few hundred meters above the present ice/bedrock contact have been reported from elsewhere in western Dronning Maud Land and East Antarctica (e.g., Liu et al., 2010). Relatively lower melt $^3\text{He}/^4\text{He}$ values (0–8 R_a) are shown by the boulders and dike sample AL/WM1e-98, all collected very close to the continental ice sheet. Although we cannot rule out cosmogenic ^3He influence in these cases either, it has likely been less than for AL/B16-98, which is exposed at higher elevation and farther from the glacier at Basen (Fig. 2b). Importantly, AL/B16-98 shows similar crush $^3\text{He}/^4\text{He}$ values with the boulder samples (Fig. 3), indicating that possible cosmogenic ^3He is primarily entrained in the olivine crystal lattice and has not significantly equilibrated with the gas-bearing inclusions.

The presence of radiogenic ^4He in the unabraded fractions is revealed by the fact that they show lower melt $^3\text{He}/^4\text{He}$ than the abraded fractions of the same sample (Table 1; Fig. 3). The

225 abrasion probably did not remove all of the implanted radiogenic ^4He , however. For example, the
226 very low crush and melt R_a (< 2) and high He contents of sample AL/WM1e-98 suggest strong
227 overprinting by radiogenic ^4He : the groundmass in this sample is very fine-grained and average
228 olivine grain size relatively small making the olivine fraction more vulnerable to groundmass
229 influence and abrasion treatment less effective. This hypothesis is substantiated in Fig. 3a, in which
230 the crushing step analyses of AL/WM1e-98 trend towards the groundmass reference sample.
231 Relatively low crush $^3\text{He}/^4\text{He}$ values ($\leq 4 R_a$) are also shown by some other fractions that include
232 small grain sizes ($\varnothing < 0.5$ mm) or were not abraded (Table 1; Fig. 3).

233 The abraded coarse ($\varnothing = 0.5\text{--}1$ mm) fraction of AL/B1b-03 shows the highest total crush
234 $^3\text{He}/^4\text{He}$ ($6.67 R_a$) and step crush $^3\text{He}/^4\text{He}$ ($7.03 R_a$; 1st step). It is also the only fraction that released
235 over 50% of He during the first crushing step (Fig. 3b), indicating that the olivines contained a
236 relatively high amount of gas-bearing inclusions. Although we cannot exclude some effect of
237 radiogenic ^4He addition in this case, we interpret the value of $7.03 (\pm 0.23) R_a$, given by the 1st
238 crushing step of an abraded coarse olivine fraction, as the best estimate for the magmatic $^3\text{He}/^4\text{He}$ of
239 the depleted ferropicrite suite. It is very unlikely that the variably treated sample fractions all
240 descended from $^3\text{He}/^4\text{He}$ values as high as shown by some of the CFB-related picrites from NAIP
241 ($\sim 50 R_a$; Stuart et al., 2003; Starkey et al., 2009) to a rather homogeneous average crush $^3\text{He}/^4\text{He}$
242 value of $4.79 R_a$ ($\sigma = 1.32$; excluding the strongly ^4He -overprinted AL/WM1e-98).

243

244 **4. Geochemical comparison with basalts from SWIR and relevant hotspots**

245

246 *4.1. He isotopes*

247 The closest estimate for the magmatic He isotopic composition ($7.03 R_a$) of the depleted ferropicrite
248 suite is compared with the magmatic He isotopic compositions of SWIR MORB, and oceanic

249 plateaus, seamounts, and oceanic islands of the southern Indian and Atlantic Oceans (Fig. 1) in Fig.
250 4.

251 It is important to note that the $^3\text{He}/^4\text{He}$ values presented for the oceanic lavas in Fig. 4 are
252 measured values and the data has not been age-corrected for radiogenic ingrowth of ^4He in the
253 mantle source. Preliminary estimates for the He isotopic evolution of SWIR MORB sources in 180
254 Ma can be calculated using U (0.0047 ppm), Th (0.0137 ppm), and ^3He ($44\text{--}170 \cdot 10^{-6} \mu\text{cc/g}$ (STP))
255 contents representative of depleted mantle (Salters and Stracke, 2004; Porcelli and Ballentine, 2002)
256 and present-day $^3\text{He}/^4\text{He}$ of 6.3–7.3 R_a (Georgen et al., 2003). Such sources would exhibit $^3\text{He}/^4\text{He}$
257 of 6.4–7.6 R_a at 180 Ma (i.e., 0.1–0.3 R_a higher relative to present-day values), which indicates
258 negligible effect for the age-correction at least in this case. Calculation of age-corrections for the
259 mantle sources of oceanic islands is hampered by the lack of knowledge on U, Th, and He
260 compositions of the different mantle components, but the above example indicates that changes of
261 more than $\sim 1 R_a$ in 180 Ma would require relatively U- and Th-enriched and/or He-depleted mantle
262 compositions.

263 Many of the oceanic hotspot-derived volcanic formations exhibit relatively high present-day
264 $^3\text{He}/^4\text{He}$ ($>8 R_a$). Values similar to the depleted ferropicrite suite (i.e. $\sim 6\text{--}8 R_a$) have been measured
265 from SWIR MORB, Comoros Islands, Gough Island, Prince Edward Islands, Tristan da Cunha,
266 Crozet Islands (Possession & East Islands), and Kerguelen Islands, although in the case of the latter,
267 many samples show comparatively high $^3\text{He}/^4\text{He}$ (Fig. 4). The He isotopic similarity to the Prince
268 Edward Islands is of particular interest, as the Marion hotspot presumed to feed the volcanism there
269 has been considered as a possible instigator of the Karoo volcanism as well (e.g., Richards et al.,
270 1989).

271 Picrites from NAIP, especially the ones that have been linked to non-chondritic primitive
272 mantle sources, show very high $^3\text{He}/^4\text{He}$ (up to 50 R_a ; Stuart et al., 2003; Starkey et al., 2009; not
273 shown in Fig. 4), considerably different from the magmatic value of the depleted ferropicrite suite.

274 Unfortunately, no data exist for the Madagascar CFBs or Paraná-Etendeka CFBs, so a helium
275 isotopic comparison is not possible.

276

277 4.2. *Sr, Nd, and Pb isotopes*

278 Given the uncertainties in defining the magmatic $^3\text{He}/^4\text{He}$ (see section 3.2.), it is also important to
279 consider other isotopic systems in the context of Indian Ocean and CFB magmatism. In Sr-Nd-Pb
280 isotopic space (Fig. 5–6; all data calculated at 180 Ma; none available for Discovery Seamounts,
281 Meteor Rise, or Shona Ridge), the great majority of the oceanic hotspots of the southern Indian and
282 Atlantic Oceans exhibit compositions (i.e., relatively high $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$, and low ϵ_{Nd})
283 that indicate relative long-term enrichment in incompatible trace elements, and are distinct from
284 those of the depleted ferropicrite suite and SWIR MORB. The depleted mantle source end-member
285 of the Etendeka CFBs exhibits similar $^{87}\text{Sr}/^{86}\text{Sr}$, ϵ_{Nd} , $^{206}\text{Pb}/^{204}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ to the depleted
286 ferropicrite suite, but $^{207}\text{Pb}/^{204}\text{Pb}$ is notably higher.

287 Similar age-corrected Sr, Nd, and Pb isotopic compositions with the depleted ferropicrite suite
288 are found from NAIP, Madagascar CFBs, and SWIR MORB (Fig. 5–6). In the case of Madagascar
289 CFBs, the high-Fe-Ti basalts and picrites having the highest ϵ_{Nd} and lowest $^{206}\text{Pb}/^{204}\text{Pb}$ (similar to
290 that of the depleted ferropicrite suite) have been linked to contribution from Indian Ocean upper
291 mantle sources (those of modern SWIR MORB), whereas the compositions of the more $^{206}\text{Pb}/^{204}\text{Pb}$ -
292 enriched basalts have been linked to contribution from the Marion plume (Storey et al., 1997). In
293 the case of SWIR MORB, the basalts that derive from sources isotopically most similar to the
294 depleted ferropicrite suite at 180 Ma are found between 22°E and 47°E (Fig. 1). It is important to
295 note that the SWIR MORBs that show influence of Marion hotspot (found between 36–39°E) or
296 DUPAL isotopic signature (i.e. high $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ at a given $^{206}\text{Pb}/^{204}\text{Pb}$; found
297 between 39–41°E) are not included in these samples (see Janney et al., 2005).

298

299 4.3. Trace elements

300 Direct comparison of trace element characteristics of the depleted ferropicrite suite with SWIR
301 MORBs and relevant CFBs and oceanic hotspots is hampered by the effects of possible secondary
302 alteration and, notably, variable degrees and pressures of mantle melting. The Nb/Y vs. Zr/Y
303 diagram (Fig. 7) introduced by Fitton et al. (1997), however, is not susceptible to discrepancies
304 imposed by these processes. Firstly, Nb, Zr, and Y are high-field strength elements that are not
305 easily mobilized by secondary processes. Secondly, although Nb/Y and Zr/Y show positive
306 correlation with increasing pressure and decreasing degree of melting, Nb/Y *at a given* Zr/Y should
307 not vary, unless the mantle source contained variable amounts of relatively Nb-enriched or -
308 depleted materials (Fitton et al., 1997).

309 Nb/Y vs. Zr/Y diagram has been found to be highly effective in distinguishing Nb-enriched
310 plume-sources from Nb-depleted MORB sources in Iceland (Fitton et al., 1997, 2003). In global
311 comparison, the great majority of OIBs, E-MORBs, and anomalous ridge segments, regardless of
312 whether they are related to thermochemical plumes or not, exhibit $\Delta Nb > 0$ [$\Delta Nb = 1.74 +$
313 $\log(Nb/Y) - 1.92\log(Zr/Y)$; Fitton et al., 1997; see Fig. 7], whereas virtually all N-MORBs exhibit
314 $\Delta Nb < 0$ (Fitton, 2007; see Fig. 7).

315 The uncontaminated depleted ferropicrite suite samples all have $\Delta Nb < 0$ and indicate that
316 their sources were not Nb-enriched (Fig. 7). In fact, their Nb-Zr-Y systematics can be explained by
317 high-pressure (+low-degree?) melting of MORB-like sources and they are connected to the field of
318 N-MORBs and SWIR MORBs by the Rooi Rand dikes of southern Lebombo that mark the first
319 phases of oceanization of the Africa-Antarctica rift (Jourdan et al., 2007b; Fig. 7; see also Heinonen
320 et al., 2010).

321 In contrast to the depleted ferropicrite suite, the CFB-related primitive samples from NAIP,
322 Etendeka, and Madagascar are composed of or include samples with $\Delta Nb > 0$. In the case of NAIP,
323 complex mixing of Nb-enriched sources with Nb-depleted sources is indicated by the trace element

324 and isotopic data (Stuart et al., 2003). The estimated trace element composition of the non-
325 chondritic primitive mantle source (Jackson and Jellinek, 2013), proposed as the main component
326 for the most $^3\text{He}/^4\text{He}$ -enriched NAIP picrites (Jackson and Carlson, 2011), exhibits high ΔNb , and
327 thus cannot represent a viable source for the depleted ferropicrite suite. The Nb-Zr-Y composition
328 of Madagascar CFBs indicate mixing with MORB and Marion hotspot sources, compatible with
329 what has been suggested for them on the basis of isotopic data (Storey et al., 1997).

330

331 **5. Implications for the Karoo CFBs and Indian Ocean magmatism**

332

333 The depleted ferropicrite suite exhibits considerable similarity to SWIR MORBs in terms of Sr, Nd,
334 Pb, and He isotopic compositions and incompatible trace element characteristics (Fig. 4–7). Major
335 element modeling indicates similar degrees of dominantly peridotitic mantle melting for both suites
336 (~10% and ~10–20%, respectively; Langmuir et al., 1992; Heinonen and Luttinen, 2010) so they
337 likely sampled mantle heterogeneity on a roughly equal scale. The presented results are thus
338 compatible with a common mantle source for these rocks that are separated by an age of 180 Ma
339 and were generated in distinct tectonic settings (see also Heinonen et al., 2010). In contrast, the He
340 isotopic results and trace element systematics (Fig. 7) preclude significant contributions from non-
341 chondritic primitive mantle sources, and Sr, Nd, and Pb isotopic data preclude significant
342 contributions from modern hotspots that are responsible for the generation of oceanic islands and
343 plateaus in the Indian and southern Atlantic Oceans (Figs. 5 and 6).

344 One notable feature that is not directly compatible with the SWIR MORB sources for the
345 depleted ferropicrite suite are the high crystallization temperatures ($> 1600\text{ }^{\circ}\text{C}$; Heinonen and
346 Luttinen, 2010) calculated on the basis of olivine-melt equilibria (method of Putirka et al., 2007).
347 The calculations imply generation from a source that was heated more than $200\text{ }^{\circ}\text{C}$ above ambient
348 (and modern Indian Ocean) upper mantle temperature, despite the presence of igneous amphibole

349 that indicates magmatic water contents of 1–2 wt. % for the parental melts (Heinonen and Luttinen,
350 2010). Recent study using thermometry that relies on exchange of Al between co-crystallizing
351 olivine and spinel, and is largely independent of pressure or melt composition, suggests that
352 estimations on the basis of olivine-liquid equilibria overshoot the actual crystallization (and
353 possibly also mantle source) temperatures by more than 100 °C, however (Heinonen et al., in press).
354 In any case, the mantle potential temperature required for the generation of the high-Mg depleted
355 ferropicrite suite is likely to be at least ~100 °C above that of ambient upper mantle.

356 If a mantle plume is invoked as the cause for the elevated temperatures, two different
357 scenarios could hypothetically explain the compositional similarity of the depleted ferropicrite suite
358 to SWIR MORB: 1) plume contains material with MORB source signature or 2) upper mantle
359 material was entrained into a rising plume (see Kerr et al., 1995). Scenario 1 would be the first
360 known Phanerozoic plume of such composition as modern plume-related rocks, even if having
361 MORB-like high ϵ_{Nd} , show either high ΔNb , high $^3\text{He}/^4\text{He}$, and/or anomalous Pb isotopic
362 compositions (e.g., Hawaii, Iceland; Kerr et al., 1995; Fitton, 2007; Figs. 6 and 7). Scenario 2 is
363 hampered by the lack of evidence on even more magnesian and hot and/or isotopically distinct
364 sublithospheric Karoo magmas that would have been expected to form within the plume itself (see
365 Kerr et al., 1995). On the other hand, it is important to note that the great majority of the Karoo
366 CFBs have been contaminated by or derive from the Gondwanan lithosphere (e.g., Hawkesworth et
367 al., 1984; Sweeney et al., 1994; Luttinen and Furnes, 2000; Jourdan et al., 2007a; Neumann et al.,
368 2011): possible plume-derived geochemical signatures may thus have been overprinted beyond
369 recognition (see Heinonen et al., 2014). Ellam et al. (1992) postulated mixing of lithospheric and
370 plume sources on the basis of Os isotopic composition of the Mwenezi picrites in Africa, but the
371 most recent studies on the Pb and Hf isotopic composition of these rocks led Ellam (2006) to favor
372 predominant contribution from variably enriched lithospheric mantle instead. In addition to depleted
373 ferropicrite suite and related Low-Nb magma type (Heinonen et al., 2010; Luttinen et al., in press)

374 of Vestfjella, the only Karoo magma types that are uncontaminated by the crust and are likely to
375 derive dominantly from sublithospheric sources are the Vestfjella enriched ferropicrite suite
376 (Heinonen and Luttinen, 2008; Heinonen et al., 2010), Group 3 dikes from Ahlmannryggen (Riley
377 et al., 2005), and some high- ϵ_{Nd} E-MORB Rooi Rand dikes from Africa (Armstrong et al., 1984;
378 Duncan et al., 1990). The emplacement of the Rooi Rand dikes post-dates the main CFB phase by
379 ~10 Ma and is related to the onset of oceanization of the rifted margin (Jourdan et al., 2007b; see
380 Fig. 7). The enriched ferropicrite suite and Group 3 dikes likely sampled recycled heterogeneities,
381 but these magma types are not similar in isotopic composition with relevant modern hotspots, are
382 not parental to Karoo CFBs, and are likely to pre-date (Group 3) and post-date (enriched ferropicrite
383 suite) the main phase of Karoo magmatism (Riley et al., 2005; Heinonen and Luttinen, 2008;
384 Heinonen et al., 2010, 2014). The suggested link between the modern Marion (or any other southern
385 Indian Ocean or Atlantic) plume and Karoo CFBs is thus not currently supported by the
386 geochemical data. Intriguingly, it could be argued that the generation of the Madagascar CFBs at
387 ~90 Ma marks the arrival of the Marion plume head to the Indian Ocean mantle lithosphere (Storey
388 et al., 1997).

389 The viability of the mantle plume hypothesis as an explanation for the generation of a specific
390 magmatic province should not be evaluated by geochemistry alone (e.g., Farnetani and Hofmann,
391 2014). Several recent studies on the structure, geochronology, and petrology of the Karoo province
392 are not readily compatible with the traditional mantle plume hypothesis (Le Gall et al., 2002, 2005;
393 Watkeys, 2002; Jourdan et al., 2004, 2005, 2006, 2007a, 2009; Coltice et al., 2009; Hastie et al.,
394 2014; cf. Ferraccioli et al., 2005; Curtis et al., 2008). One of the preferred alternative models for the
395 Karoo CFBs is the “internal mantle heating” model, which suggests an increase in the upper mantle
396 potential temperatures of up to ~100 °C beneath an insulating supercontinent (Coltice et al., 2009).
397 Favorable plate boundary assemblies prior to Gondwana breakup may have further enhanced the
398 effect of subcontinental heating (see Rolf et al., 2012; Hastie et al., 2014). The implication of this

399 study, that the hottest and most primitive sublithospheric magmas known from Karoo seem to
400 sample Indian Ocean upper mantle, is compatible with such propositions. Supercontinent insulation
401 could be further associated with drainage of slowly accumulated sublithospheric magma reservoir
402 (Silver et al., 2006) and lithospheric delamination (Elkins-Tanton and Hager, 2000) that could also
403 explain the large-scale metasomatism observed in the Indian Ocean upper mantle (Janney et al.,
404 2005; Hanan et al., 2013).

405 The NAIP may be one of the best examples of a plume-instigated CFB province (e.g., Brown
406 and Lesher, 2014), but the role of mantle plumes in the generation of Karoo CFBs remains an open
407 question. The hypothesis that all CFB provinces that were formed close to the margins of large low
408 shear velocity provinces (LLSVPs) in the lower mantle are plume-generated (Burke and Torsvik,
409 2004) may thus be an overgeneralization. After all, all CFB provinces related to the early stages of
410 Pangea and Gondwana breakup (i.e. Central Atlantic Magmatic Province, Ferrar, and Karoo) have
411 been presented as the best Phanerozoic candidates for internal mantle heating models (Coltice et al.,
412 2009; Hastie et al., 2014; Hole, 2015), and at least one of them hosts high-temperature picrites
413 generated from a low- $^3\text{He}/^4\text{He}$, long-term depleted mantle source.

414

415 **6. Conclusions**

416

417 The age of the Karoo CFB eruptions presents challenges to obtaining the mantle helium isotopic
418 compositions, given the presence of radiogenic and cosmogenic helium. The analytical strategy
419 utilized here, of abrasion and coupled crushing/melting in vacuum, shows that it is possible to
420 estimate the effects of different helium components to the total $^3\text{He}/^4\text{He}$ in these samples. The best
421 estimate for the magmatic $^3\text{He}/^4\text{He}$ isotopic composition (obtained from a fresh, abraded, and coarse
422 olivine fraction) of the most magnesian Karoo CFB magma type (Vestfjella depleted ferropicrite
423 suite) is $7.03 \pm 0.23 R_a$. This value is similar to those obtained from SWIR MORBs ($6.3\text{--}7.3 R_a$;

source $^3\text{He}/^4\text{He} \sim 6.4\text{--}7.6 R_a$ at 180 Ma) and notably lower than those measured from most of the southern Atlantic or Indian Ocean hotspots, or CFB-related high-Mg rocks thought to derive from early-formed non-chondritic primitive mantle sources. Trace element, Sr, Nd, and Pb isotopic comparisons further consolidate these relationships. The composition and high-T origin of the suite are best explained by supercontinent insulation of a precursory Indian Ocean upper mantle source, which is also preferred on the basis of the latest structural, geochronological, and petrological studies on the Karoo CFBs.

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Figure captions

Figure 1. Present-day location of Karoo-Ferrar CFBs (and related dikes), Etendeka CFBs, Madagascar CFBs, Southwest Indian Ridge, and relevant oceanic plateaus, seamounts, and oceanic islands (stars) depicted. Gray diamonds indicate SWIR MORB samples that are geochemically (namely on the basis of Sr, Nd, and Pb isotopes) most similar to the depleted ferropicrite suite (AGU22-8-1, AGU22-9-2, AGU53-3-3, MELPROT-5-11-17, MELPROT-5-14-81, MELPROT-5-

800 15-54, MELPROT-5-17-22, MELPROT-5-25-217, MELPROT-5-29-34, MELPROT-5-30-75,
801 MELPROT-5-42-11; Hamelin and Allegre, 1985; le Roex et al., 1989; Mahoney et al., 1989, 1992;
802 Janney et al., 2005).

803

804 **Figure 2.** Distribution of Jurassic CFBs in western Dronning Maud Land (a) and depleted
805 ferropicrite suite sampling sites relevant to this study (b, c) shown. Location of Aboa (Finnish) and
806 Wasa (Swedish) research stations shown in (b). Arrow marked with “G” indicates the approximate
807 primary ice flow direction and stippled lines indicate steep cliffs in (b) and (c). H.U.S. = H. U.
808 Sverdrupfjella.

809

810 **Figure 3.** Helium content and isotopic composition released by crushing of the depleted ferropicrite
811 suite olivine fractions shown in $^3\text{He}/^4\text{He}$ (R/R_a) vs. He ($\mu\text{ccSTP/g}$) (a) and $^3\text{He}/^4\text{He}$ (R/R_a) vs. He
812 fraction released (%) (b). He fraction has been calculated for total He released, including the melt
813 analysis. 2σ analytical error or maximum effect of nucleogenic ^3He production as the negative error
814 of relevant analyses (Table 1) shown for $^3\text{He}/^4\text{He}$. Successive step crushing analyses of each sample
815 fraction show decrease in He contents (i.e. proceed from right to left in the diagrams), except in the
816 case of abraded AL/B7-03, AL/B16-98, and AL/WM1e-98 fractions, for which the light dashed
817 lines connect results from step crushing for clarity (Table 1). Note that in the case of AL/B9-03
818 fractions and unabraded AL/B16-98 fraction, there are only data available for a single crush
819 analysis (Table 1). The dark arrows illustrate interpretation of magmatic and radiogenic components
820 in the olivine mineral separates. a = abraded fraction, ua = unabraded fraction

821

822 **Figure 4.** Magmatic $^3\text{He}/^4\text{He}$ (R/R_a ; the most reasonable estimate, see text) of the Vestfjella
823 depleted ferropicrite suite, SWIR MORB, and relevant oceanic plateaus, seamounts, and oceanic
824 islands compared (see Fig. 1; name of the associated mantle hotspot or plume mentioned in

parentheses). Strongly DM-influenced samples have not been included and a part of the data points represent anomalous ridge segments next to hotspots. Crozet Islands exhibit clear division into two distinct $^3\text{He}/^4\text{He}$ components (Possession & East Is. and Penguins Is.). Helium isotopic data for Madagascar CFBs, Etendeka CFBs, Conrad rise or Kerguelen plateau has not been published. The presented $^3\text{He}/^4\text{He}$ ratios are the measured ones and have not been age-corrected: the $^3\text{He}/^4\text{He}$ of the mantle sources is likely to change less than $\sim 1 R_a$ in 180 Ma (e.g., $\sim 0.1\text{--}0.3 R_a$ for the SWIR MORB source; see Section 4). Data sources: Kurz et al. (1982); Hilton et al. (1995); Graham et al. (1999); Sarda et al. (2000); Hanyu et al. (2001); Georgen et al. (2003); Class et al. (2005); Doucet et al. (2006); Nicolaysen et al. (2007); Fri et al. (2011); Breton et al. (2013).

Figure 5. Representative ϵ_{Nd} vs. $^{87}\text{Sr}/^{86}\text{Sr}$ compositions shown for the Vestfjella depleted ferropicrite suite and the estimated depleted sublithospheric sources of Etendeka CFBs (uncontaminated Horingbaai picrites), Madagascar CFBs (only samples with $\epsilon_{\text{Nd}} > 5$ included), NAIP (only uncontaminated samples included), SWIR MORBs, and relevant oceanic plateaus, seamounts, and oceanic islands (see Fig. 1; Prince Edward Islands and Kerguelen hotspot-related rocks shown separately for the sake of discussion) at 180 Ma. The age-corrected source compositions for rocks younger than ~ 180 Ma have been estimated by first calculating the initial isotopic composition of the rocks where relevant (NAIP at 60 Ma, Madagascar at 90 Ma, Kerguelen plateau at 110 Ma, and Etendeka at 132 Ma; corrections for < 30 Ma rocks neglected) and then using $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ calculated on the basis of depleted MORB mantle (DMM; Workman and Hart, 2005) and bulk silicate Earth (BSE; Workman and Hart, 2005) evolution trend at the initial moment to calculate the isotopic composition of the source at 180 Ma; for example, in the case of a modern mantle-derived volcanic rock, the equations for present-day $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ of the source are $37.439(^{87}\text{Sr}/^{86}\text{Sr})_i - 26.287$ and $105.25(^{143}\text{Nd}/^{144}\text{Nd})_i - 53.759$, respectively. Continuous depletion of DMM is neglected as it does not have a significant effect in

time scales relevant to this study. Data sources: Lightfoot et al. (1997); Storey et al. (1997); Graham et al. (1998); Melluso et al. (2001, 2005); Thompson et al. (2001); Stuart et al. (2003); Kent et al. (2004); Heinonen and Luttinen (2008); Starkey et al. 2009; Heinonen et al. (2010); data for SWIR MORBs compiled from PetDB (www.earthchem.org/petdb) and data for oceanic plateaus, seamounts, and oceanic islands compiled from GEOROC (www.georoc.mpch-mainz.gwdg.de).

Figure 6. Representative $^{207}\text{Pb}/^{204}\text{Pb}$ (a) and $^{208}\text{Pb}/^{204}\text{Pb}$ (b) vs. $^{206}\text{Pb}/^{204}\text{Pb}$ compositions shown for the Vestfjella depleted ferropicrite suite, Vestfjella enriched ferropicrite suite and the estimated depleted sublithospheric sources of Etendeka CFBs (uncontaminated Horingbaai picrites), Madagascar CFBs (only samples with $\epsilon_{\text{Nd}} > 5$ included), NAIP (only uncontaminated samples included), SWIR MORBs, and relevant oceanic plateaus, seamounts, and oceanic islands (Fig. 1; Prince Edward Islands and Kerguelen hotspot-related rocks shown separately for the sake of discussion) at 180 Ma. The age-corrected source compositions for rocks younger than ~180 Ma have been estimated in a similar way as explained in Fig. 5, except that $^{238}\text{U}/^{204}\text{Pb}$ and $^{232}\text{Th}/^{204}\text{Pb}$ of the sources were calculated on the basis of age-corrected D-DMM–DMM–E-DMM trend (Workman and Hart, 2005) and by assuming present-day $(^{208}\text{Pb}/^{204}\text{Pb})_{\text{DMM}}$ of 37.7 (Hart, 1988), $(^{208}\text{Pb}/^{204}\text{Pb})_{\text{D-DMM}}$ of 36.9 (on the basis of correlation with global MORB $^{207}\text{Pb}/^{204}\text{Pb}$), and $(^{208}\text{Pb}/^{204}\text{Pb})_{\text{E-DMM}}$ of 38.6 (on the basis of correlation with global MORB $^{207}\text{Pb}/^{204}\text{Pb}$). Data sources: Lightfoot et al. (1997); Storey et al. (1997); Graham et al. (1998); Thompson et al. (2001); Jourdan et al. (2007b); Heinonen et al. (2010); Jackson et al. (2010); data for SWIR MORBs compiled from PetDB (www.earthchem.org/petdb) and data for oceanic plateaus, seamounts, and oceanic islands compiled from GEOROC (www.georoc.mpch-mainz.gwdg.de).

Figure 7. Nb/Y vs. Zr/Y diagram of the Vestfjella depleted ferropicrite suite, Rooi Rand dikes of Karoo, primitive Etendeka CFBs (uncontaminated Horingbaai picrites), primitive Madagascar

875 CFBs (only samples with $\epsilon_{\text{Nd}} > 5$ included), primitive NAIP picrites (only samples with $\epsilon_{\text{Nd}} > 5$
876 included), Marion hotspot basalts, and normal SWIR MORBs (see Figs. 5 and 6). The Iceland
877 Plume Array and fields for N-MORB and OIB are after Fitton et al. (2003). NCPM marks the non-
878 chondritic primitive mantle source suggested for the NAIP picrites that exhibit the highest $^3\text{He}/^4\text{He}$
879 (Jackson and Jellinek, 2013). P and F arrows denote the approximate effects of increasing pressure
880 and degree of melting, respectively (Fitton et al., 1997). Lithospheric mantle contamination could
881 change Nb/Y at a given Zr/Y (Heinonen et al., 2010) so it is important to use this diagram only for
882 uncontaminated CFB samples. Data sources: Armstrong et al. (1984); Lightfoot et al. (1997);
883 Storey et al. (1997); Melluso et al. (2001, 2005); Thompson et al. (2001); Stuart et al. (2003); Kent
884 et al. (2004); Starkey et al. (2009); data for SWIR MORBs compiled from PetDB
885 (www.earthchem.org/petdb) and data for Prince Edward Islands compiled from GEOROC
886 (www.georoc.mpch-mainz.gwdg.de).

887

888

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Table 1

Table 1 Helium isotopic composition of the Vestfjella depleted ferropicrite suite

Sample	features*	olivine Ø [mm]	weight [g]	step	⁴ He [µccSTP/g]	⁴ He [%]	³ He/ ⁴ He (R/R _a)	2σ (R/R _a)	N- ³ He err.# (R/R _a)
AL/B1b-03	boulder MgO = 19 wt.% FeO ^{tot} = 13 wt.% U = 0.13 ppm Th = 0.43 ppm	0.5 □ 1 unabraded	0.263	1	0.0158	22.7	4.52	0.17	-0.07
				2	0.0083	11.9	5.01	0.20	-0.13
				3	0.0050	7.2	4.06	0.20	-0.21
				crush _{TOT}	0.0291	41.8	4.58	0.18	-0.04
				melt	0.0405	58.2	4.23	0.14	-0.03
				TOTAL	0.0696	100	4.38	0.16	-0.01
		0.25 □ 1 abraded	0.220	1	0.0095	23.9	4.11	0.18	-0.11
				2	0.0040	10.0	5.05	0.29	-0.26
				3	0.0028	7.0	4.86	0.31	-0.37
				4	0.0022	5.4	4.03	0.30	-0.48
				5	0.0012	3.0	5.15	0.47	-0.88
				6	0.0011	2.8	5.13	0.56	-0.93
				crush _{TOT}	0.0207	52.3	4.50	0.27	-0.05
			0.206	melt	0.0189	47.7	8.10	0.26	-0.06
				TOTAL	0.0395	100	6.22	0.27	-0.03
		0.5 □ 1 abraded	0.209	1	0.0303	51.1	7.03	0.23	-0.03
				2	0.0041	7.0	4.95	0.28	-0.25
				3	0.0023	3.9	5.03	0.33	-0.45
				crush _{TOT}	0.0367	62.0	6.67	0.24	-0.03
			0.194	melt	0.0225	38.0	7.25	0.24	-0.05
				TOTAL	0.0592	100	6.89	0.24	-0.02
AL/B7-03	boulder MgO = 19 wt.% FeO ^{tot} = 13 wt.% U = 0.14 ppm Th = 0.45 ppm	0.5 □ 1 unabraded	0.208	1	0.0151	18.2	4.70	0.18	-0.05
				2	0.0071	8.6	3.16	0.16	-0.11
				3	0.0039	4.7	3.19	0.21	-0.20
				crush _{TOT}	0.0261	31.5	4.06	0.18	-0.03
			0.199	melt	0.0568	68.5	3.97	0.13	-0.01
				TOTAL	0.0829	100	3.99	0.15	-0.01
		0.25 □ 0.5 abraded	0.315	1	0.0257	29.6	2.28	0.08	-0.03
				2	0.0034	3.9	4.49	0.24	-0.23
				3	0.0103	11.8	2.11	0.10	-0.07
				crush _{TOT}	0.0393	45.3	2.43	0.10	-0.02
			0.305	melt	0.0476	54.7	6.40	0.19	-0.02
				TOTAL	0.0869	100	4.60	0.15	-0.01
AL/B9-03	boulder MgO = 24 wt.% FeO ^{tot} = 13 wt.% U = 0.10 ppm Th = 0.32 ppm	0.25 □ 1 abraded	0.332	crush _{TOT}	0.0249	26.6	4.49	0.24	-0.02
				melt	0.0684	73.4	3.67	0.20	-0.01
				TOTAL	0.0933	100	3.89	0.21	0.00
		gm unabraded	0.300	crush _{TOT}	0.2001	4.0	0.14	0.01	0.00
				melt	4.7690	96.0	0.11	0.01	0.00
				TOTAL	4.9691	100	0.11	0.01	0.00
AL/B16-98	dike MgO = 17 wt.% FeO ^{tot} = 15 wt.% U = 0.21 ppm Th = 0.86 ppm	0.5 □ 1 unabraded	0.279	crush _{TOT}	0.0249	26.6	6.14	0.19	-0.02
				melt	0.0684	73.4	10.26	0.31	-0.01
				TOTAL	0.0933	100	8.86	0.27	-0.06
		0.25 □ 0.5 abraded	0.258	1	0.0171	23.4	5.57	0.29	-0.05
				2	0.0046	6.4	5.35	0.34	-0.20
				3	0.0064	8.8	5.22	0.30	-0.14
				crush _{TOT}	0.0281	38.6	5.46	0.30	-0.03
			0.245	melt	0.0449	61.4	15.38	0.47	-0.02
				TOTAL	0.0730	100	11.55	0.40	-0.01
AL/WM1e-98	dike MgO = 19 wt.% FeO ^{tot} = 15 wt.% U = 0.15 ppm Th = 0.49 ppm	0.25 □ 0.5 abraded	0.153	1	0.1223	0.6	0.44	0.03	-0.01
				2	0.0187	0.1	1.50	0.10	-0.07
				3	0.0845	0.4	0.21	0.02	-0.02
			0.138	crush _{TOT}	0.2255	1.2	0.44	0.03	-0.01
				melt	19.2000	98.8	0.04	0.01	0.00
				TOTAL	19.4255	100	0.04	0.01	0.00

* Whole-rock geochemical data from Heinonen and Luttinen (2008); # Maximum effect of nucleogenic ^3He growth on the R/R_a value calculated using the equations presented by Andrews (1985), neutron capture cross-sections provided for 17 elements presented by Andrews and Kay (1982), and whole-rock chemistry presented by Heinonen and Luttinen (2008). Li, B, Cl, and Co contents of the samples are unknown but required for the calculations, so they were estimated using the following trace element ratios: $\text{K/Li} = 118.64(\text{La/Yb}) + 17.77$ (see Ryan and Langmuir, 1993), $\text{B/K} = 0.001$ (see Ryan and Langmuir, 1993), $\text{Cl/K} = 0.04$ (see Stronick and Haase, 2004), and $\text{Ni/Co} = 6$ (see Le Roux et al., 2011).