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# **Os isotopic constraints on crustal contamination in Auckland**

## **Volcanic Field basalts, New Zealand**

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### Keywords

Monogenetic basaltic volcanism, Os isotope, crustal contamination, Auckland Volcanic Field,  
magma ascent pathways

### Highlights

- Evidence of crustal contamination observed in Auckland Volcanic Field basalts.
- Highly radiogenic Os isotope ratios coupled with very low Os concentrations indicate a crustal metasedimentary contaminant.
- Slightly elevated Os isotope ratios with high Os concentrations indicate contamination from sulphide bearing olivines.

29 **Abstract**

30 The Auckland Volcanic Field (AVF) represents the youngest and northernmost of three  
31 subjacent Quaternary intraplate basaltic volcanic fields in the North Island, New Zealand.  
32 Previous studies on AVF eruptive products suggested that their major- and trace- element, and  
33 Sr-, Nd- and Pb-isotopic signatures primarily reflect their derivation from the underlying  
34 asthenospheric and lithospheric mantle. All AVF lavas however ascend through a ca. 20-30 km  
35 thick continental crust, and some do carry crustal xenoliths, posing the question whether or not  
36 crustal contamination plays a role in their formation. Here we present new Os and Pb isotopic  
37 data, and Os and Re concentrations for 15 rock samples from 7 AVF volcanic centres to  
38 investigate mantle and crustal petrogenetic processes. The samples include the most primitive  
39 lavas from the field (Mg# 59-69) and span a range of eruption sizes, ages, locations, and  
40 geochemical signatures. The data show a large range in Os concentrations (6-579 ppt) and  
41  $^{187}\text{Os}/^{188}\text{Os}$  isotope ratios from mantle-like (0.123) to highly radiogenic (0.547). Highly  
42 radiogenic Os signatures together with relatively low Os contents in most samples suggest that  
43 ascending melts experienced contamination primarily from metasedimentary crustal rocks with  
44 high  $^{187}\text{Os}/^{188}\text{Os}$  ratios (e.g., greywacke). We further demonstrate that <1% metasedimentary  
45 crustal input into the ascending melt can produce the radiogenic Os isotope signatures observed  
46 in the AVF data. This low level of crustal contamination has no measurable effect on the  
47 corresponding trace element ratios and Sr-Nd-Pb isotopic compositions. In addition, high Os  
48 contents (195-578 ppt) at slightly elevated but mantle-like Os isotopic compositions  
49 ( $^{187}\text{Os}/^{188}\text{Os} = 0.1374\text{-}0.1377$ ) in some samples suggest accumulation of xenocrystic olivine-  
50 hosted mantle sulphides from the Permian-Triassic ultramafic Dun Mountain Ophiolite Belt,  
51 which traverses the crust beneath the Auckland Volcanic Field. We therefore infer that the AVF  
52 Os isotopic compositions and Os contents reflect contamination from varying proportions of  
53 heterogeneous crustal components, composed of Waipapa and Murihiku terrane  
54 metasediments, and ultramafic rocks of the Dun Mountain Ophiolite Belt. This demonstrates,  
55 contrary to previous models that primitive lavas from the Auckland Volcanic Field do show  
56 evidence for variable interaction with the crust.

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## 62 1. Introduction

63 Monogenetic basaltic volcanic fields are the surface expression of small-scale magmatic  
64 systems and are found in a number of different tectonic settings, including extensional systems  
65 (e.g., Cascades, USA; Borg et al., 2000), subduction zones (e.g., Wudalianchi, China; Hwang et al.,  
66 2005), or intraplate settings relating to lithospheric rifting (e.g., Panter et al., 2006) and  
67 delamination of the lower lithosphere (e.g., Hoernle et al., 2006). The formation of individual  
68 monogenetic volcanic centres has mainly been attributed to isolated, often small-volume,  
69 batches of magma (<0.1 km<sup>3</sup>) (e.g., Connor and Conway, 2000; Németh, 2010; Kereszturi et al.,  
70 2013), which erupt quickly (e.g., Németh, 2010), creating volcanic fields which can encompass  
71 tens to hundreds of individual centres (e.g., Condit and Connor, 1996; Conway et al., 1998;  
72 Connor and Conway, 2000; Valentine et al., 2005). The fields show a range of surface  
73 expressions, dependent on the eruption style and magma-water interaction, including tuff  
74 rings, maars, scoria cones, and lava flows (e.g., Allen and Smith, 1994; Németh, 2010; Kereszturi  
75 et al., 2014). A link between eruptive volumes of monogenetic basaltic volcanoes, and  
76 mineralogical and geochemical composition of the underlying mantle has recently been  
77 proposed (McGee et al., 2015) highlighting the importance of understanding the characteristics  
78 of different sources contributing to magmatism beneath volcanic fields.

79 Petrological and geochemical studies have shown that most eruptive centres in  
80 monogenetic fields are composed of silica-undersaturated basanites, nephelenites and alkaline  
81 basalts that are akin to Ocean Island Basalts (OIBs) (e.g., Huang et al., 1997; Cook et al., 2005;  
82 Valentine and Gregg, 2008). Nevertheless, eruption products can show considerable  
83 geochemical variations both within volcanic fields (e.g., Valentine and Hirano, 2010; Timm et al.,  
84 2010; McGee et al., 2013) and within individual centres, the origins of which remain  
85 controversial (e.g., Bradshaw and Smith, 1994; Valentine and Gregg, 2008; Needham et al.,  
86 2011; Brenna et al., 2010, 2011; McGee et al., 2012). Several studies have attributed these  
87 variations in geochemical and isotopic signatures, both for individual eruptions and field-wide  
88 scales, to heterogeneities in the underlying mantle (e.g., Huang et al., 1997; McBride et al., 2001;  
89 Cook et al., 2005; McGee et al., 2013), magma modification by lithospheric contamination  
90 (mantle and crust) or fractional crystallisation (e.g., Lassiter and Luhr, 2001; Alves et al., 2002;  
91 Chesley et al., 2002; Jamais et al., 2008; Timm et al., 2009). Although crustal assimilation and  
92 contamination may be facilitated by storage or ponding of magma within the crust (e.g.,  
93 Bohrson et al., 1997), eruptive products in monogenetic volcanic fields are generally mafic (>8  
94 wt.% MgO) thus arguing for limited fractional crystallization and against prolonged crustal  
95 magma storage. Therefore, the variations in major, and trace element signatures, and Sr-, Nd-,  
96 and Pb-isotopic compositions of rocks from monogenetic volcanic fields are proposed to

97 primarily reflect the composition of the underlying mantle (e.g., Valentine and Perry, 2007;  
98 McGee et al., 2012, 2013, 2015). Nevertheless, even mafic magmas in monogenetic fields are not  
99 primary melts and therefore are likely to have undergone some interaction with the crust  
100 through which they ascend, potentially affecting the physical and chemical properties of the  
101 melts. Thus, it is a prerequisite to distinguish between the geochemical signatures of mantle  
102 heterogeneity versus crustal contamination (e.g., Blondes et al., 2008; Jung et al., 2011).

103 To understand the role of the crust and mantle petrogenesis beneath an archetypal  
104 continental monogenetic volcanic field we use the  $^{187}\text{Re}$ - $^{187}\text{Os}$  decay system. This system is  
105 highly sensitive to crustal contamination due to the large difference in incompatibility between  
106 parent (Re) and daughter (Os) elements during partial melting of mantle lithologies. During  
107 partial melting Os behaves compatibly in mantle sulphides and thus dominantly remains in the  
108 mantle, whereas Re is moderately incompatible and preferentially enters the melt. This  
109 contrast in behaviour of Re and Os therefore result in extreme fractionation (e.g., high Re/Os in  
110 melt and crust, and low Re/Os in the mantle) which, with time, leads to an overall significant  
111 contrast between the  $^{187}\text{Os}/^{188}\text{Os}$  ratios of continental crust (up to  $^{187}\text{Os}/^{188}\text{Os} = 5.0$ ; McBride et  
112 al., 2001) versus typical mantle sources (ca. 0.12; Meisel et al., 1996).  $^{187}\text{Os}/^{188}\text{Os}$  isotopic values  
113 in the eruptive products of volcanic systems therefore provide a unique tool to decipher the  
114 influences of crustal contamination (e.g., Central European Volcanic Province: Jung et al., 2011;  
115 Newer Volcanics Province: McBride et al., 2001) and mantle source heterogeneity.

116 Here we present new Os and Re concentrations, as well as Os and Pb isotope data for 15  
117 mafic samples from the Auckland Volcanic Field, New Zealand. These new data are aimed at  
118 identifying the relative importance of mantle heterogeneity, including crustal recycling, and  
119 crustal contamination into ascending melts, in order to give new insights into dynamics of melt  
120 generation and ascent beneath the city of Auckland.

121

## 122 **2. The Auckland Volcanic Field**

123 The monogenetic basaltic Auckland Volcanic Field (AVF) is located 400 km west of the  
124 currently active arc, the Hikurangi Margin (e.g., Seebeck et al., 2014)(**Fig. 1.A**). The AVF is the  
125 northernmost of three intraplate monogenetic volcanic fields that become progressively  
126 younger towards the north (Ngatutura and Okete ca. 2.7-1.5 Ma; Briggs et al., 1994, and South  
127 Auckland Volcanic Field (SAVF) ca. 1.59-0.51 Ma; Cook et al., 2005) (**Fig. 1.A**).

128 The AVF consists of 53 individual centres (**Fig. 1.B**; Hayward et al., 2011) that  
129 collectively cover ca. 360 km<sup>2</sup> (**Fig. 1**; Allen and Smith, 1994; Kermode, 1992). The centres

130 include one or more explosion craters, tuff rings (some now occupied by lakes), scoria cones  
131 and lava flows. The individual centres are interpreted to have formed through single eruptions  
132 of small magma batches (e.g., Allen and Smith, 1994), except for Rangitoto, the youngest and  
133 largest volcano where two eruption episodes have been identified (Rangitoto 1 [553±7 cal. yrs.  
134 BP] and Rangitoto 2 [504±5 cal. yrs. BP]; Needham et al., 2011). The total volume of the AVF  
135 field is estimated at 1.7 km<sup>3</sup> dense rock equivalent (DRE<sup>tot</sup>), ca. 41% (0.7 km<sup>3</sup>) of which is  
136 represented by Rangitoto (Kereszturi et al., 2013). Basaltic volcanism commenced at ca. 190 ka  
137 (Lindsay et al., 2011) and shows some distinct changes in eruption frequency over time (e.g.,  
138 Molloy et al., 2009; Hopkins et al., 2015), however the ages of many of the individual eruptive  
139 centres are currently poorly constrained (Lindsay et al., 2011).

140 The crust underlying the AVF is 20-30 km thick, and composed of Waipapa and the  
141 Murihiku terranes (**Fig. 1.C**) (e.g., Kermode 1992; Eccles et al., 2005; Horspool et al., 2006;  
142 Mortimer et al., 2014), which are overlain by up to 1-2 km of Miocene Waitemata terrane  
143 sediments. The western Waipapa and eastern Murihiku terranes mainly consist of late Triassic  
144 to late Jurassic low-grade meta-sediments, separated by the Dun Mountain terrane. This terrane  
145 represents a Permian to late Cretaceous oceanic arc ophiolite obduction event at the eastern  
146 Gondwana margin (e.g., Kimbrough et al., 1992) and mainly consists of ultramafic rocks,  
147 including dunites, lherzolites, harzburgites and werhlites (cf. McCoy-West et al., 2013). This  
148 terrane is reflected in a distinct positive magnetic anomaly known as the Junction Magnetic  
149 Anomaly (JMA), which can be traced continuously through the North and South islands of New  
150 Zealand (Hatherton and Sibson, 1970). The JMA traverses the Auckland area as a narrow (ca. 2-  
151 5 km), linear series of positive magnetic anomalies interpreted to be eastward-dipping  
152 serpentinised shear zones, extending throughout the 20-30 km thick crust beneath the AVF  
153 (Eccles et al., 2005). Direct evidence of the Dun Mountain Ophiolite Belt (DMOB) beneath the  
154 AVF occurs in the form of serpentinite xenoliths within volcanic rocks in tuff rings from Pupuke,  
155 St. Heliers and Taylors Hill volcanoes (e.g., Searle, 1959; Bryner, 1991; Jones, 2007; Spörli et al.,  
156 2015). The basement rocks that make up the region of the study area have been extensively  
157 investigated (e.g., Bryner, 1991; Kermode, 1992; Eccles et al., 2005; McCoy-West et al., 2013),  
158 allowing potential sources of crustal contamination to be well characterised.

159 The Auckland Volcanic Field has been the focus of previous geochemical and  
160 petrological studies (Huang et al., 1997; Smith et al., 2008; Needham et al., 2011; McGee et al.,  
161 2011, 2012, 2013, 2015). Low degrees of partial melting (≤6%), a limited proportion of  
162 fractional crystallisation, in addition to a limited range of Sr-, Nd-, and Pb-isotopic compositions  
163 of the AVF lavas were interpreted to reflect the mantle origin of the AVF lavas. Huang et al.  
164 (1997) furthermore attributed the more radiogenic <sup>206</sup>Pb/<sup>204</sup>Pb isotopic compositions (>19.2) of

165 the AVF rocks to the presence of 'young' HIMU signature formed in  $\leq 0.2$  Ga in the underlying  
166 mantle (cf. Thirlwall, 1997). Alternatively, McGee et al. (2013) suggested that AVF lavas with  
167 more radiogenic  $^{206}\text{Pb}/^{204}\text{Pb}$  isotopic compositions, coupled with low  $\text{SiO}_2$  (<48 wt.%) and  
168 Nb/U, but high CaO/ $\text{Al}_2\text{O}_3$ , Na<sub>2</sub>O and K<sub>2</sub>O, and elevation Ce/Pb, Nb/Ce and U/Pb represent melts  
169 derived from carbonated garnet peridotite domains hosted in a depleted mantle-like peridotitic  
170 asthenosphere. The more siliceous AVF alkali basalts ( $\text{SiO}_2 \geq 48$  wt.%) were in contrast  
171 attributed to interaction of ascending lithospheric melts with (or direct melting of) an EMII-type  
172 lithospheric mantle, previously metasomatised by subduction-related fluids. These  
173 interpretations of the AVF major and trace element, and Sr-, Nd-, and Pb- isotopic systematics  
174 rely on the assumption that the AVF eruption products are directly representative of their  
175 mantle sources, as proposed by McGee et al. (2013), and are not affected by contamination.  
176 However, a number of studies have noted xenolithic materials (e.g., crustal schistose and non-  
177 schistose fragments, meta-igneous rocks, and abundant large olivines) within some AVF centres,  
178 for example at St Heliers, Taylors Hill, and Mangere Mt (Spörli et al., 2015), providing direct  
179 evidence for the interaction of ascending magmas with the underlying crustal and mantle  
180 lithologies (e.g., Bryner, 1991). Therefore, traditionally used major and trace element contents  
181 and ratios, and Sr-, Nd-, and Pb-isotopes may not be sensitive enough to distinguish between  
182 crustal contamination and mantle heterogeneity. We address this problem here using the more  
183 sensitive Os isotope system.

184

### 185 **3. Methods**

#### 186 **3.1. Sample selection**

187 Specific samples were chosen for this study in order to complement existing data and to  
188 cover not only a range of geochemical compositions of the field (c.f. McGee et al., 2011, 2012,  
189 2013, 2015) but also a range in ages (Needham et al., 2011; Lindsay et al., 2011), locations,  
190 eruptive volumes and types (Kereszturi et al., 2013, 2014). Based on previous analyses, fifteen  
191 primitive samples were chosen (Mg# =59-69) from Rangitoto, Mt Wellington, Purchas Hill,  
192 Three Kings, Wiri, Puketutu, and Pupuke. New Pb isotope data are added by this study for 7  
193 samples, and new major and trace element data are added for 5 samples (see **Table 1** for  
194 classification). For all fifteen samples new Re and Os concentrations, and Os isotope ratios are  
195 determined. Generally the selected samples contain minor olivine phenocrysts ( $\leq 3$  mm across)  
196 in a plagioclase, pyroxene, and olivine-bearing groundmass, except for samples from Pupuke,  
197 which contain abundant large olivines  $\geq 5$  mm across.

198

199 **3.2. Analytical techniques**

200 Sample preparation for chemical and isotopic analyses were conducted at Victoria  
201 University of Wellington, New Zealand (VUW). Samples were chipped using a Rocklabs Boyd  
202 crusher to <15 mm, then reduced to powder in an agate ring mill. Major elements were analysed  
203 by XRF analysis at the Open University, Milton Keynes, United Kingdom (UK), on an ARL® 8420+  
204 dual goniometer spectrometer. Powdered samples were fused with lithium metaborate and  
205 analysed following methods of Ramsey et al. (1995). Whin Sill dolerite was run as an internal  
206 standard with associated accuracy of <1% except for Na<sub>2</sub>O (2.37%) and P<sub>2</sub>O<sub>5</sub> (1.59%) and  
207 precision of <2%. For trace element concentrations 50 mg of sample powder was digested in  
208 hot concentrated HF + HNO<sub>3</sub> for 4 days, then dried to incipient dryness and taken up in  
209 concentrated HCl. Following this, samples were converted back into HNO<sub>3</sub> and left for 3 days in  
210 hot 1M HNO<sub>3</sub> to form the final analytical solution. Centrifuged sample dilutions were measured  
211 on an Agilent 7500CS ICP-MS at Victoria University, Wellington, using BHVO-2 as a primary  
212 standard, and BRCR-2 as a secondary internal standard. All results and standard values are  
213 reported in the supplementary material (SM) (**SM.1 and SM.2**). Precision on BCR-2 (n=15 from  
214 five digestions) was <6.5% 2sd% except for Nb, Cs, and Ba (≤8 %) and Ta, Pb, and Nb (≤20.5 %) and  
215 values were all within <6 % of the reference value, except for Cu, Cs, and Ta. All standard  
216 values are outlined in the supplementary material.

217 Pb isotope samples were prepared and purified in an ultra-clean chemical separation  
218 laboratory at VUW. Powdered sample was leached in ultrapure 6M HCl for 1 hour at 120°C,  
219 rinsed with MilliQ water, and digested in ultrapure conc. HNO<sub>3</sub> + ultrapure conc. HF, then turned  
220 into solution with 0.8M HBr. The solution was centrifuged, loaded onto 5 mm columns equipped  
221 with AG1-X8 resin, and Pb was extracted in a double-pass using 6M HCl. Pb isotopic  
222 compositions were analysed using a Neptune MC-ICP-MS at Durham University, UK, results and  
223 errors are reported in **Table 1**. International standard NBS-981 was used to monitor machine  
224 drift, with internal precisions (2SE) of <sup>206</sup>Pb/<sup>204</sup>Pb <±0.0012, <sup>207</sup>Pb/<sup>204</sup>Pb <±0.0013, <sup>208</sup>Pb/<sup>204</sup>Pb  
225 <±0.0044. All data are normalised to NBS-981 standard values reported by Baker et al. (2004)  
226 (<sup>206</sup>Pb/<sup>204</sup>Pb = 16.9416, <sup>207</sup>Pb/<sup>204</sup>Pb = 15.4998, <sup>208</sup>Pb/<sup>204</sup>Pb = 36.7249), all standard  
227 measurements can be found in supplementary material (**SM2.3**).

228 Os isotope compositions and Re and Os contents were determined at Geotop, Université  
229 du Québec à Montréal, Canada, following the method of Meisel et al. (2003). For these analyses  
230 0.8 g aliquots of whole rock powder were spiked with a known enriched tracer solution of  
231 <sup>190</sup>Os/<sup>185</sup>Re, and digested in Teflon-sealed quartz tubes with 3 ml 6M HCl and 3 ml conc. HNO<sub>3</sub> at  
232 300 °C and 100 bars in a high-pressure asher unit (HPA-S, Anton-Parr). Following this, Os was



233 extracted using the Paris Br<sub>2</sub> technique (Birck et al., 1997). 2 ml of chilled Br<sub>2</sub> was added to the  
234 digested sample and left on a hot plate at 90 °C for 2 hours. This scavenges the already oxidised  
235 Os (OsO<sub>4</sub>) from the aqueous solution into the liquid Br<sub>2</sub>, leaving Re (and PGEs) within the  
236 aqueous solution. 20 drops of HBr was added to the isolated liquid Br<sub>2</sub> (including Os) to reduce  
237 Os from volatile Os<sup>8+</sup> (OsO<sub>4</sub>) to non-volatile Os<sup>4+</sup> (OsBr<sub>6</sub><sup>2-</sup>), and then evaporated down. A final  
238 step of micro-distillation using Cr<sup>VI</sup> containing H<sub>2</sub>SO<sub>4</sub> was used to purify the Os, and make sure it  
239 was quantitatively separated from isobaric Re (after Birck et al., 1997). Samples were then  
240 measured using a Triton TIMS in negative-ion mode for Os (Creaser et al., 1991), and sector  
241 field (SF)-ICP-MS for Re. Os and Re concentrations were calculated by isotope dilution from  
242 known spike solution values. Os blanks for total procedure are 0.3 pg, and 7 pg for Re, these  
243 value are subtracted from the sample totals in data processing along with oxygen interface and  
244 sample-spike unmixing corrections. Results and errors for Os and Re analysis are reported in  
245 **Table 1**. The widely used standard Durham Romil Osmium solution (DROsS) was used with an  
246 internal precision (2SE) for <sup>187</sup>Os/<sup>188</sup>Os of average measurements of <±0.00012 (Luguet et al.,  
247 2008; Nowell et al., 2008, see supplementary details for measurements (SM2.4)), and for Re  
248 standard NIST SRM 3143 was used, with internal precision (2SE) for <sup>187</sup>Re/<sup>185</sup>Re of average  
249 measurements of <±0.05 (see SM.2.4). Duplicate analyses were undertaken in two ways, on  
250 aliquots of one sample solution post-digestion to test consistency of digestion techniques, and of  
251 sample powders to check the internal consistency of samples.

252

## 253 **4. Results**

### 254 **4.1. Major and trace elements**

255 Volcanic rocks from the AVF range in composition from sub-alkaline silica under-  
256 saturated basanites to alkali basalts (e.g., SiO<sub>2</sub> = 39.8-48.8 wt.%, **Fig. 2.**) following the rock  
257 classification of LeMaitre (2002). In general, the data for the AVF samples form a broad negative  
258 trend with MgO vs. SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub> and positive trends with CaO Fe<sub>2</sub>O<sub>3</sub><sup>t</sup>, TiO<sub>2</sub>, and P<sub>2</sub>O<sub>5</sub> indicative  
259 of olivine and pyroxene fractionation (**Fig. 2.**). The exceptions are, two samples from Rangitoto,  
260 which have lower Fe<sub>2</sub>O<sub>3</sub><sup>t</sup>, TiO<sub>2</sub>, CaO and P<sub>2</sub>O<sub>5</sub> values and higher SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub> values at a given  
261 content of MgO, and two samples from Pupuke, which have lower values for CaO, TiO<sub>2</sub> and  
262 Fe<sub>2</sub>O<sub>3</sub><sup>t</sup>, but higher values of MgO than the other AVF rocks.

263 All AVF samples used in this study show primitive mantle-normalised trace element  
264 distributions broadly similar to ocean island basalt (OIB) (normalisation values after  
265 McDonough and Sun, 1995), with typical positive Nb-Ta, and negative K anomalies (**Fig. 3.**). As

266 previously identified by McGee et al. (2013), AVF samples with low SiO<sub>2</sub> contents have higher  
267 Nb and Ta relative to the large ion lithophile elements (LILE; Rb, Ba, K) (e.g., Nb/Ba ≥ 0.2), and  
268 higher Light Rare Earth Element (LREE; La, Gd, Nb) concentrations. Furthermore, these samples  
269 (those with low SiO<sub>2</sub> content) have high LREE relative to their heavy rare earth elements (HREE;  
270 Yb), resulting in high LREE/HREE ratios (e.g., (La/Yb)<sub>N</sub> >20), and elevated ratios of highly to  
271 moderately incompatible elements, such as Th/Yb ≥2. In comparison to the low SiO<sub>2</sub>, two of the  
272 Rangitoto samples with high SiO<sub>2</sub> values have lower Nb and Ta concentrations relative to LILE  
273 (e.g., Nb/Ba ≤ 0.2) and lower LREE but similar to slightly higher HREE (e.g., (La/Yb)<sub>N</sub> <20).  
274 These high SiO<sub>2</sub> Rangitoto samples also have higher Th relative to HREE (e.g., Th/Yb <1), lower  
275 Ce/Pb (<20), yet higher K/Nb (>300) and Zr/Nb (>7). These Rangitoto samples also show a  
276 positive Sr-anomaly (**Fig. 3**), which is less prominent in all other samples; conversely all other  
277 samples show a prominent negative K-anomaly (**Fig. 3**), which is not observed within the  
278 Rangitoto samples.

279 These findings are consistent with the previous work of McGee et al. (2013, 2015), and  
280 are also comparable to samples from the South Auckland Volcanic Field (SAVF; Cook et al.,  
281 2005). Cook et al. (2005) noted higher Nb, (La/Yb)<sub>N</sub> and Th/HREE values for samples with low-  
282 SiO<sub>2</sub> (their Group B) and low Nb, (La/Yb)<sub>N</sub> and Th/HREE for samples with high-SiO<sub>2</sub> (their  
283 Group A) and contrasting incompatible trace element ratios (e.g., K/Nb, Zr/Nb and Ce/Pb).  
284 These observations are also consistent with those of Hoernle et al. (2006) and Timm et al.  
285 (2009, 2010) who divided their data from New Zealand intraplate volcanic centres and fields  
286 into 'low-SiO<sub>2</sub>' and 'high-SiO<sub>2</sub>' groups, showing similar major and trace element geochemical  
287 compositions to the AVF low- and high-SiO<sub>2</sub> samples. For the purpose of this study we have  
288 chosen not to group our data due to the limited number of samples, but highlight the similarity  
289 in geochemical signatures for the AVF samples.

290

#### 291 **4.2. Pb isotopes (combined with published Sr-, Nd-, and Pb-isotopes)**

292 Pb isotopic data for the AVF rocks are reported in **Table 1**. Seven new Pb isotope  
293 determinations have been generated to supplement the data of McGee et al. (2013). In general  
294 the AVF Pb isotopic compositions show limited variation, with <sup>206</sup>Pb/<sup>204</sup>Pb = 19.064-19.383,  
295 <sup>207</sup>Pb/<sup>204</sup>Pb = 15.576-15.615, and <sup>208</sup>Pb/<sup>204</sup>Pb = 38.485-38.931. On Pb isotope variation  
296 diagrams the AVF data plot between the mid ocean ridge basalt (MORB), HIMU (high μ = high  
297 time integrated U/Pb) and enriched mantle (EM I or II) end-members (following the  
298 classification of Zindler and Hart, 1986; **Fig. 4. A&B.**). Purchas Hill sample AU44711 shows the  
299 highest <sup>206</sup>Pb/<sup>204</sup>Pb (19.383) and high <sup>208</sup>Pb/<sup>204</sup>Pb (38.929), but <sup>207</sup>Pb/<sup>204</sup>Pb similar to rocks

300 from other AVF centres (15.613). Conversely, the two Rangitoto samples show the lowest  
301  $^{206}\text{Pb}/^{204}\text{Pb}$  (19.064), low  $^{208}\text{Pb}/^{204}\text{Pb}$  (38.485), and some of the highest  $^{207}\text{Pb}/^{204}\text{Pb}$  values  
302 (15.615) of all AVF volcanic rocks (**Fig. 4.**). Most other AVF samples plot between the Purchas  
303 Hill and Rangitoto samples, with the possible exception of the samples from Wiri Mt, which have  
304 the lowest  $^{207}\text{Pb}/^{204}\text{Pb}$  (15.576-15.579) at intermediate  $^{206}\text{Pb}/^{204}\text{Pb}$  (19.219-19.225) and  
305  $^{208}\text{Pb}/^{204}\text{Pb}$  (38.791-38.795). These Pb isotope compositions observed in AVF rocks generally  
306 overlap with those from the SAVF ( $^{206}\text{Pb}/^{204}\text{Pb} = 18.959-19.332$ ;  $^{207}\text{Pb}/^{204}\text{Pb} = 15.579-15.624$ ,  
307 and  $^{208}\text{Pb}/^{204}\text{Pb} = 38.752-38.953$ : data from Cook et al. (2005), values are re-normalised to  
308 standard SRM-981 values reported in Baker et al. (2004), to maintain consistency with values  
309 from McGee et al. (2013), and this study; **Fig. 4.**).

310 Previously published Sr and Nd isotope data for the same samples analysed in this study  
311 show a restricted range from  $^{87}\text{Sr}/^{86}\text{Sr} = 0.702710$  to  $0.703125$  and from  $^{143}\text{Nd}/^{144}\text{Nd} =$   
312  $0.512939$  to  $0.512956$  (McGee et al., 2013; **Fig. 4. C&D.**, reported in supplementary material  
313 SM.1.). Although no obvious trend has been identified between Sr and Nd isotopic compositions,  
314 some of the AVF data (McGee et al., 2013) show similar trends to those observed by Cook et al.  
315 (2005) in the SAVF samples. Some samples have distinctly higher  $^{87}\text{Sr}/^{86}\text{Sr}$  (ca. 0.7032) than the  
316 samples from the other centres at similar  $^{143}\text{Nd}/^{144}\text{Nd}$  values (ca. 0.5129) (**Fig. 4.C&D.**).  
317 Similarly there is no clear correlation observed between  $^{206}\text{Pb}/^{204}\text{Pb}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$ , although  
318  $^{206}\text{Pb}/^{204}\text{Pb}$  and  $^{87}\text{Sr}/^{86}\text{Sr}$  show a poorly defined negative trend (Cook et al., 2005; McGee et al.,  
319 2013; **Fig. 4.C.**). Similar to most AVF samples, the SAVF 'Group B' type (Cook et al., 2005) are  
320 characterised by lower  $^{87}\text{Sr}/^{86}\text{Sr}$  values ( $\leq 0.7028$ ) at similar  $^{143}\text{Nd}/^{144}\text{Nd}$  values (0.51296-  
321 0.51298), coupled with higher  $^{206}\text{Pb}/^{204}\text{Pb}$  (19.210-19.332), higher  $^{208}\text{Pb}/^{204}\text{Pb}$  (38.862-38.935)  
322 and lower  $^{207}\text{Pb}/^{204}\text{Pb}$  values (15.579-15.600). In comparison, SAVF 'Group A' type (Cook et al.,  
323 2005), has higher  $^{87}\text{Sr}/^{86}\text{Sr}$  values ( $\geq 0.7029$ ) and similar  $^{143}\text{Nd}/^{144}\text{Nd}$  values (0.51294-0.51298),  
324 coupled with lower  $^{206}\text{Pb}/^{204}\text{Pb}$  (18.959-19.286), lower  $^{208}\text{Pb}/^{204}\text{Pb}$  (38.752-38.905), and higher  
325  $^{207}\text{Pb}/^{204}\text{Pb}$  values (15.597-15.624) similar to samples from Rangitoto (See **Fig. 4.**).

326

### 327 **4.3. Re, Os, and $^{187}\text{Os}/^{188}\text{Os}$**

328  $^{187}\text{Os}/^{188}\text{Os}$  values and Re and Os concentrations are given in **Table 1**. The AVF samples  
329 show a wide range in Os and Re concentrations (Os = 5.8-578 ppt, Re = 25.9-411.9 ppt; **Fig. 5.**),  
330 but concentrations fall in the range of contents reported from other OIB-type lavas (Os = 1-600  
331 ppt: Schiano et al., 2001; Re = 100-642 ppt: Hauri and Hart, 1997). Similarly, the AVF volcanic  
332 rocks have a significant range in Os isotope compositions from mantle-like  $^{187}\text{Os}/^{188}\text{Os}$  of 0.1230  
333 (cf. Meisel et al., 2000) to radiogenic values up to  $^{187}\text{Os}/^{188}\text{Os} = 0.5470$  (**Fig. 6.**). The majority of

334 samples (n = 19; including duplicates) have  $^{187}\text{Os}/^{188}\text{Os} \geq 0.15$ , higher than the range inferred for  
335 mantle-derived magmas (Lassiter and Hauri, 1998; Widom et al., 1999; Rasoazanamparany et  
336 al., 2015). In general no obvious correlations between Os and Re concentrations,  $^{187}\text{Os}/^{188}\text{Os}$   
337 values and MgO, Ni, Cu or Zr concentrations or trace element ratios are observed. The  
338 exceptions are samples from Pupuke, which have high MgO, Ni, and Os concentrations (**Fig. 5.**),  
339 but relatively unradiogenic  $^{187}\text{Os}/^{188}\text{Os}$  values (**Table 1.**).

340 The overall range in  $^{187}\text{Os}/^{188}\text{Os}$  values in the AVF rocks (0.1230 to 0.5470) is larger  
341 than those observed in most OIB-like basalts (ca. 0.11 to 0.14; Reisberg et al., 1993; Hauri and  
342 Hart, 1993; Roy-Barman and Allègre, 1995; Marcantonio et al., 1995; Widom and Shirey, 1996;  
343 Lassiter and Hauri, 1998; Schiano et al., 2001; Day et al., 2010), and overlap with those in alkali  
344 basaltic and tholeiitic lavas from other continental intraplate volcanic fields. These include  
345 Newer Volcanic Province, SE Australia ( $^{187}\text{Os}/^{188}\text{Os} = 0.1342\text{-}0.4456$ ; McBride et al., 2001);  
346 Central European Volcanic Province, Germany ( $^{187}\text{Os}/^{188}\text{Os} = 0.1487\text{-}0.7526$ ; Jung et al., 2011)  
347 and the East African Rift System ( $^{187}\text{Os}/^{188}\text{Os} = 0.1239\text{-}0.4366$ ; Nelson et al., 2012) (**Fig 6.**).  
348 Samples with more radiogenic Os isotopic composition generally contain less Os (e.g., Rangitoto  
349 sample AU59309:  $^{187}\text{Os}/^{188}\text{Os} = 0.5470$ , Os = 5.8 ppt), and those with lower  $^{187}\text{Os}/^{188}\text{Os}$  contain  
350 more Os (e.g., Wiri Mt. sample AU43931:  $^{187}\text{Os}/^{188}\text{Os} = 0.1283$ , Os = 194 ppt). In addition, low Os  
351 concentration samples (<40 ppt) generally have more variable Os isotope signatures (0.1623-  
352 0.5470), becoming less variable (0.1230-0.1374) with increasing Os contents (>100 ppt). On an  
353  $^{187}\text{Os}/^{188}\text{Os}$  versus Os diagram (**Fig. 6.**) the AVF data generally plot between crustal and mantle  
354 end-member fields, similar to other Pacific OIB fields (**Fig. 6.**)

355 Duplicate analyses of the same sample using the same digestion, and same sample  
356 quantities, show highly reproducible results (e.g., Purchas Hill and Wiri samples; Table 1),  
357 however, duplicate analyses of the same sample using different digestions show results which  
358 plot along a positive trend between high Os with low  $^{187}\text{Os}/^{188}\text{Os}$ , and low Os with high  
359  $^{187}\text{Os}/^{188}\text{Os}$  (**Fig. 6.**). These include samples from AVF centres Rangitoto, Three Kings, Purchas  
360 Hill, Mt Wellington, and Puketutu. This effect has previously been observed (e.g., Allègre and  
361 Luck, 1980; Potts, 1987; Alves et al., 2002) and termed the 'nugget effect'. This suggests that the  
362 variable  $^{187}\text{Os}/^{188}\text{Os}$  in an individual digestion is caused by sampling differing amounts of  
363 mineral-hosted sulphides, or oxide micro-inclusions, with either inherited crustal, or mantle  
364  $^{187}\text{Os}/^{188}\text{Os}$  signatures in a single powdered sample. The high reproducibility of Os contents and  
365  $^{187}\text{Os}/^{188}\text{Os}$  for a split of the same digestion (c.f. **Table 1**) demonstrates effective digestion of the  
366 sample, and argues against the variability in the results from different digestions of the same  
367 sample being due to incomplete sample dissolution. Single samples are likely to contain

368 sulphide populations that have different Os contents and  $^{187}\text{Os}/^{188}\text{Os}$  values, which are variably  
369 sampled by duplicate sample splits.

370

## 371 **5. Discussion**

### 372 **5.1. The effects of mantle source heterogeneity and fractional crystallisation on Os** 373 **signatures**

374 The petrogenesis of the AVF basalts has been the subject of a number of geochemical  
375 and isotopic studies detailed in **section 2** (Huang et al., 1997; Needham et al., 2011; McGee et  
376 al., 2011, 2012, 2013, 2015). The isotopic data, with specific emphasis on Pb isotope ratios, have  
377 been used previously (Huang et al., 1997; McGee et al., 2013, 2015) to identify and fingerprint  
378 three heterogeneous mantle sources, which are variably mixed to produce the isotopic and  
379 geochemical variability observed within the AVF eruptive products. New Pb isotope data added  
380 by this study (**Fig. 4 A&B.**) are consistent with those published for the AVF (McGee et al., 2013)  
381 and the SAVF (Cook et al., 2005), however, new Os isotope data obtained by this study show no  
382 correlation to the Pb-Sr-Nd isotope systems (**Fig. 7**), suggesting that different processes control  
383 these isotopic signatures within the AVF melts. This hypothesis is further supported by the lack  
384 of correlation between Os and Re contents and the major and trace element concentrations of  
385 the AVF rocks (**Fig. 5.**). In addition, no correlation is observed between the key trace element  
386 ratios, and trace element anomalies (K and Sr) that have previously been used to determine the  
387 mantle source signatures of the AVF rocks (e.g., La/Yb ratios; McGee et al., 2013, 2015). The  
388 different behaviours of the major and trace element concentrations and isotope systems  
389 therefore question whether or not a heterogeneous mantle source alone is responsible for the  
390 AVF Os signatures.

391 The generally low and variable Os concentrations, and lack of correlation between Os  
392 and Re and the other major and trace element concentrations could be explained through  
393 partial melting and fractional crystallisation. However, the concentrations of Os and Re in most  
394 AVF samples show no obvious correlation between  $^{187}\text{Os}/^{188}\text{Os}$  and MgO or, Ni, Cu or Zr (**Fig 5.**),  
395 suggesting that fractional crystallisation of the major phases found within the AVF rocks  
396 (olivine and pyroxene  $\pm$ plagioclase) has an insignificant influence on the Os and Re budgets.  
397 However, Platinum Group Elements (PGE), including Os, in addition to being siderophile are  
398 also chalcophile, and thus are relatively concentrated in sulphide minerals or PGE-rich metal  
399 alloys (e.g.,  $D_{\text{Os}} \sim 10^4$  between sulphides and silicate melt; Roy-Barman et al., 1998; Jamais et al.,  
400 2008; Park et al., 2013; McCoy-West et al., 2015). Very minimal amounts of sulphide

401 fractionation will therefore preferentially remove Os from a melt, without a strong co-variance  
402 with other elements. This is likely to contribute to the low Os concentrations that are seen in  
403 the majority of AVF samples. Conversely, Re is less compatible to moderately incompatible in  
404 sulphides and thus is not preferentially incorporated into them. Sulphides formed through  
405 fractional crystallisation are either retained in residual minerals in the mantle (e.g., olivine;  
406 Lorand et al., 2010) or segregate during magma transport and emplacement (e.g., Bézou et al.,  
407 2005). The PGE concentrations in the melt therefore depend on the degree of partial melting, as  
408 only high degrees of partial melting (ca.  $\geq 20\%$ ) will completely exhaust Base Metal Sulphides  
409 (BMSs) from the mantle source, resulting in a PGE-enriched melt (e.g., Rehkämper et al., 1999;  
410 Lorand et al., 2010; Dale et al., 2012). Accordingly, lower-degree partial melts will result in  
411 lower and more variable HSE contents due to incomplete breakdown of sulphide in the mantle  
412 source (e.g., Day et al., 2010). As low degrees of partial melting have been suggested to occur  
413 beneath the AVF ( $\leq 6\%$ ; McGee et al., 2013) variable and generally low Os contents are  
414 therefore expected in the AVF melts. However, the absence of co-variance between the above-  
415 mentioned elements and Os suggest that partial melting and silicate + sulphide phase  
416 fractionation cannot alone cause the variability in Os contents observed in the AVF rocks. Some  
417 of the AVF samples have high Os contents together with high MgO and Ni contents (**Fig. 5.**),  
418 suggestive of the accumulation of xenocrystic olivine ( $\pm$ pyroxene)-hosted sulphides into the  
419 AVF melts during ascent (e.g., Alard et al., 2002). However, PGEs reveal an even more complex  
420 behaviour depending on  $fO_2$ ,  $fS_2$ , temperature and pressure prevailing during mantle melting,  
421 crystal fractionation and during melt ascent and eruption. Regardless of the behavioural  
422 complexity of BMSs, due to their low melting temperatures and low viscosities, they are soluble  
423 in basaltic melts and will therefore be partially incorporated into the ascending magmas (Alard  
424 et al., 2002). It is therefore likely that ascending melt will become contaminated with mantle or  
425 crustal sulphides if the magma passes through an appropriate sulphide-bearing source. We  
426 therefore propose that the Os systematics of the AVF are not controlled by mantle source  
427 heterogeneity or fractional crystallisation, and instead are most likely controlled by  
428 contamination and assimilation processes. This possibility is further investigated below in  
429 regards to the Os isotopes and Os concentrations of the AVF basalts.

430

## 431 **5.2. Contamination and assimilation**

432 On an  $^{187}\text{Os}/^{188}\text{Os}$  versus Os diagram (**Fig. 6**) most data from the AVF samples plot on a  
433 curved array between mantle-like unradiogenic  $^{187}\text{Os}/^{188}\text{Os}$  with elevated Os concentrations,  
434 and radiogenic  $^{187}\text{Os}/^{188}\text{Os}$  with low Os concentrations. This array is indicative of a mixing end

435 member (into a mantle-like source) with an elevated Os isotope ratio but a low Os content (**Fig.**  
436 **6.**). However,  $^{187}\text{Os}/^{188}\text{Os}$  vs. Sr-, Nd-, or Pb- isotopic systems (**Fig. 4.**) show no indication of a  
437 source for this signature, with no obvious trends towards either a combined mantle (e.g., young  
438 HIMU, EMII-type pelagic sediments, or carbonatite) or continental crustal signature (**Fig. 7.**).  
439 The source of this mixing end member therefore remains unclear, and so potential contributors  
440 to contamination or assimilation in the mantle and crust are discussed below.

#### 441 **5.2.1. Mantle Source**

442 Typical primitive mantle values for  $^{187}\text{Os}/^{188}\text{Os}$  are  $\leq 0.1290$  (Meisel et al., 1996), with  
443 variable Os concentrations between 1-600 ppt (Schiano et al., 2001). However, in subduction  
444 zones the Os isotopic signatures of lithospheric mantle can be increased through the  
445 incorporation of highly radiogenic oceanic crust (e.g., Alves et al., 1999; Borg et al., 2000; Dale et  
446 al., 2007; Suzuki et al., 2011). In addition, fluids derived from the subducting plate will cause  
447 metasomatism of the subcontinental lithosphere, leading to more oxidising conditions in the  
448 mantle by increasing the  $f\text{O}_2$  and  $f\text{S}_2$ . Mantle hosted sulphides, which are more stable under  
449 reducing conditions, will destabilise and oxidise to sulphates (e.g., Carroll and Rutherford,  
450 1985), releasing sulphide-bonded metals, which include high concentrations of Os, and PGEs  
451 (e.g., Jugo, 2009; Suzuki et al., 2011). Subduction-metasomatised lithosphere can therefore host  
452 a more radiogenic Os isotope signature than the ambient mantle, and has the potential to  
453 contain higher concentrations of Os in partial melts. However, because Os is compatible with  
454 residual mantle sulphides during low degree partial melting, any erupted lavas will likely have  
455 elevated radiogenic isotope signature and a low Os concentration (e.g., Widom et al., 1999).

456 The lithosphere beneath the AVF was exposed to multiple episodes of subduction at  
457 least from the Cretaceous (e.g., Mortimer et al., 2006) and from Oligocene to Miocene (30 to ca.  
458 20 Ma: Seebeck et al., 2014). In addition, studies on peridotitic xenoliths from the South Island  
459 volcanic groups (e.g., Scott et al., 2014; McCoy-West et al., 2015) present evidence for  
460 subduction-related carbonatitic metasomatism. Carbonatites typically have low Os  
461 concentrations and elevated Os isotope ratios (Widom et al., 1999; Escrig et al., 2005), and  
462 therefore could represent a mantle end-member causing the radiogenic signature seen for the  
463 AVF samples. However, with the exception of the high La/Yb ratio for the Purchas Hill samples,  
464 the AVF samples do not show the geochemical signatures (e.g., low Ti/Eu or low wt.%  $\text{Al}_2\text{O}_3$ )  
465 that have been attributed to carbonatite metasomatism (e.g., Scott et al., 2014). In addition, the  
466  $^{206}\text{Pb}/^{204}\text{Pb}$  and  $^{87}\text{Sr}/^{86}\text{Sr}$  isotope values in the AVF rocks are lower than the HIMU-type South  
467 Island xenoliths ( $^{206}\text{Pb}/^{204}\text{Pb} > 20$ ). Using  $^{187}\text{Os}/^{188}\text{Os}$  values from Widom et al. (1999) (e.g.,  
468  $^{187}\text{Os}/^{188}\text{Os} = 0.6$ ), and Os concentration from Escrig et al. (2005) (e.g., Os = 15 ppt) modelling

469 input of carbonatite into the system cannot readily explain the complete spread in Os (ppt)  
470 versus  $^{187}\text{Os}/^{188}\text{Os}$  data spread of the AVF samples used in this study (**Fig. 8.**).

471 Alternative models to explain the radiogenic  $^{187}\text{Os}/^{188}\text{Os}$  geochemical composition  
472 involve the presence of eclogitic or pyroxenitic or carbonated peridotite domains in the  
473 asthenospheric or lithospheric mantle (e.g., Hoernle et al., 2006; Sprung et al., 2007; Timm et al.,  
474 2009; McCoy-West et al., 2010; McGee et al., 2013, 2015). This poses the question whether or  
475 not partial melting of different mantle domains (peridotite vs. pyroxenite  $\pm$   $\text{CO}_2$  beneath the  
476 AVF, as originally proposed by McGee et al. (2013), may lead to the radiogenic Os isotope  
477 signature of the AVF samples. In general neither the AVF Os isotope ratios, nor the Os  
478 concentrations correlate with the typical geochemical tracers for carbonated peridotite,  
479 carbonatite metasomatism (e.g., low wt.%  $\text{SiO}_2$  and low  $\text{MgO}\#$ , or elevated  $\text{CaO}/\text{Al}_2\text{O}_3$ ,  $\text{Ce}/\text{Pb}$ ,  
480  $\text{Nb}/\text{Ce}$ ; Scott et al., 2014; McGee et al., 2015) or pyroxenite (e.g., elevated  $\text{Al}_2\text{O}_3$  coupled with  
481 radiogenic  $^{187}\text{Os}/^{188}\text{Os}$ ; Marchesi et al., 2014). This is in agreement with a recent study by  
482 McCoy-West et al. (2015) concluding that because the PGE budget is primarily controlled by  
483 residual sulphides, carbonatitic metasomatism beneath Zealandia (the wider micro-continent  
484 on which New Zealand sits) does not affect the PGE budget of the subcontinental peridotites.  
485 These results also support previous conclusions by Handler et al. (1997) who also showed that  
486 for wherlite and apatite-bearing peridotites, interaction with carbonatitic melt has no effect on  
487 the Os isotope composition or concentrations of Os or Re.

488 A number of studies (e.g., Kumar et al., 1996; Saal et al., 2001; Pearson and Nowell,  
489 2004; Luguet et al., 2008) present evidence that mantle pyroxenites and eclogites can contain  
490 secondary metasomatised sulphides with highly radiogenic  $^{187}\text{Os}/^{188}\text{Os}$  of  $>2$ , coupled with high  
491 concentrations of Os. If the AVF melts accumulated or interacted with such secondary sulphide  
492 of metasomatic origin, one would expect a correlation between high Os concentration and high  
493  $^{187}\text{Os}/^{188}\text{Os}$ , the contrary to what is generally observed. Only three of the AVF samples analysed  
494 (from Pupuke, and Wiri centres) do show slightly elevated  $^{187}\text{Os}/^{188}\text{Os}$  of 0.1374-0.1377 coupled  
495 with high Os concentrations ( $>200$  ppt; **Fig. 6.**), potentially indicative of a mantle-derived origin.  
496 Given that the latest lithospheric metasomatism event occurred 20-30 Ma ago, and based on the  
497 decay of  $^{187}\text{Re}$  to  $^{187}\text{Os}$ , the Os isotopic ratio of the present peridotitic lithospheric mantle would  
498 be ca.  $^{187}\text{Os}/^{188}\text{Os} = 0.1473\text{-}0.1524$  (assuming  $^{187}\text{Re}/^{188}\text{Os} = 50$  in recycled crust: Widom et al.,  
499 1999; and an original ambient mantle value of  $^{187}\text{Os}/^{188}\text{Os} = 0.1231\text{-}0.1283$ , from this study).  
500 This  $^{187}\text{Os}/^{188}\text{Os}$  ratio is however significantly higher than the values recorded in lithospheric  
501 mantle xenoliths from Zealandia (including dunite and wherlite samples from Ngatutura Point:  
502  $^{187}\text{Os}/^{188}\text{Os} = 0.1282 \pm 0.0012$  ca. 80 km south of Auckland; McCoy-West et al., 2013). Although  
503 the  $^{187}\text{Os}/^{188}\text{Os}$  and Os content is likely to be variable beneath North Island, New Zealand, the



504 generally lower peridotite mantle-like  $^{187}\text{Os}/^{188}\text{Os}$  values in Zealandia xenoliths suggests that  
505 the slightly elevated  $^{187}\text{Os}/^{188}\text{Os}$  at high Os contents observed in the Pupuke and Wiri Mt  
506 samples are derived from a different source, which is discussed below.

507

### 508 5.2.2. Crustal source

509 The trace element signature for the Waipapa terrane crustal rock is plotted on **Figure 3**  
510 (data from Price et al., 2015), highlighting the difference in concentrations between trace  
511 elements such as Nb and Ta. Due to the different geochemical signatures of crustal and mantle  
512 rocks, previous studies have suggested that certain trace element ratios can be indicative of  
513 crustal contamination, typically showing high La/Ta (>22) and La/Nb (>1.5) ratios (e.g., Abdel-  
514 Fattah et al., 2004) and low Nb/U (<37) ratios (e.g., Hofmann et al., 1986). All AVF samples fall  
515 outside of these ranges; exhibiting low La/Ta ratios (8.56 to 13.4) and low La/Nb ratios (0.58 to  
516 0.75), coupled with high Nb/U signatures (42.61 to 53.36), with the exception of Rangitoto  
517 sample AU59309, which shows a slightly lower Nb/U value (35.75) at low La/Ta and La/Nb  
518 values. This, coupled with the lack of correlation between the major and trace element  
519 signatures and the Os, Re concentrations and Os isotope signatures, indicates that the major and  
520 trace element geochemistry of the AVF samples are not affected by crustal contamination.

521 Radiogenic isotope systems can however be more sensitive to crustal input in  
522 comparison to the major and trace element compositions (e.g., Wilson, 1989). Sr-, Nd-, and Pb-  
523 isotope values for the Waipapa terrane (Price et al., 2015) are plotted on **Fig 4** in relation to the  
524 AVF samples. Binary mixing of the two signatures (Waipapa and EMII) into an AVF mantle-like  
525 source suggest that a  $\leq 10\%$  input from Waipapa terrane rocks and  $\leq 5\%$  input from an EMII-like  
526 source could be used to explain the spread in the Pb-isotope values (**Fig. 4 A&B**). However, this  
527 input would produce a much larger range in Sr- and Nd- isotope signatures, which is not  
528 observed within the AVF samples (**Fig. 4C&D**). This suggests that these isotope systems are also  
529 not controlled by crustal contamination and show decoupled signatures to the Os isotopic  
530 system (**Fig. 7**). Therefore the contamination highlighted by the Os signatures must be minimal  
531 in order to not show any impacts on the trace elements or Sr-, Nd- or Pb-isotopes.

532 For the Pupuke samples (as discussed previously) that show slightly elevated  
533  $^{187}\text{Os}/^{188}\text{Os}$  at high Os contents, an alternative source potentially causing the contamination  
534 with sulphide-bearing olivine is ultramafic rocks of the Dun Mountain Ophiolite Belt (DMOB)  
535 (primarily serpentinised dunite and harzburgite) (e.g., Coombs et al., 1976; Sivell and  
536 McCulloch, 2000; Eccles et al., 2005). Parts of the DMOB cross directly beneath Pupuke at  
537 shallow levels ( $\geq 1.5$  km depth, **Fig 1.C**.) making these rocks a likely melt contaminant. Although

538 no DMOB rocks were analysed for Os isotopes or Os concentrations, O'Driscoll et al. (2012)  
539 report Os contents and isotope ratios from the Shetland ophiolite complex. The latter, although  
540 older (429 Ma), has similarities to the DMOB in formation (obduction at an arc collision zone),  
541 and lithology (serpentinised dunite and harzburgite).  $^{187}\text{Os}/^{188}\text{Os}$  isotope values reported for  
542 rocks from the Shetland ophiolite complex range from 0.1204 to 0.1502 with Os concentrations  
543 of 300-8000 ppt (O'Driscoll et al., 2012). Assuming a mean value of  $^{187}\text{Os}/^{188}\text{Os} = 0.1353$  and Os  
544 = 4150 ppt, (within range of the Shetland rocks) simple binary mixing modelling (**Fig. 8.**)  
545 requires <10% input from the ultramafic rocks into our proposed ambient mantle values (of  
546  $^{187}\text{Os}/^{188}\text{Os} = 0.1231$ - $0.1283$  and Os = 50-200 ppt) to explain the  $^{187}\text{Os}/^{188}\text{Os}$  value and Os  
547 content observed in the Pupuke samples. However only ca. 5% olivine accumulation is required  
548 to explain the higher MgO, Ni and Cr contents of these samples (**Fig. 2.**), suggesting that either  
549 the Os concentration in the contaminant needs to be  $\geq 6000$  ppt (still within the range reported  
550 by O'Driscoll et al., 2012) or Os-bearing sulphides or metal alloys will dissolve more efficiently  
551 than silicates and therefore more readily contaminate the ascending melts. Regardless, we  
552 therefore attribute samples with elevated  $^{187}\text{Os}/^{188}\text{Os}$  and high Os contents to the accumulation  
553 of, or interaction with, the DMOB ultramafic rocks containing ancient subduction  
554 metasomatised domains.

555 The interaction between the DMOB and ascending melts cannot however explain the  
556 highly radiogenic Os isotope signatures ( $^{187}\text{Os}/^{188}\text{Os} = 0.1623$ - $0.5470$ ) and low Os  
557 concentrations (<50 ppt) in most other AVF samples. These highly radiogenic  $^{187}\text{Os}/^{188}\text{Os}$   
558 isotope ratios exceed all values reported from the lithospheric mantle beneath New Zealand  
559 (<0.133: cf. McCoy-West et al., 2015; Liu et al., 2015) and carbonatitic sources (<0.6: Widom et  
560 al., 1999) as previously discussed. Continental metasediments typically show highly radiogenic  
561  $^{187}\text{Os}/^{188}\text{Os}$  values (0.165-2.323: Saal et al., 1998) and low Os contents (20 to 100 ppt: Saal et al.,  
562 1998; Widom et al., 1999), suggesting that crustal derived rocks are most likely acting as a  
563 contaminating agent to explain the high  $^{187}\text{Os}/^{188}\text{Os}$  and low Os concentration in most AVF lavas.

564

### 565 **5.3. AFC modelling**

566 In order to quantify the amount of fractional crystallisation and crustal contamination  
567 acquired by the AVF samples, we used the combined assimilation fractional crystallisation  
568 model of DePaolo (1981) (**Fig. 8.**). Potential crustal contaminants have to be a known basement  
569 lithology, with a highly radiogenic Os isotope signature and low Os concentration. In addition,  
570 the eruptive products must have undergone low levels of fractional crystallisation (<6%; McGee  
571 et al., 2013), and the mixing proportions between contaminant and melt have to be low enough

572 in order to have little effect on trace element values or the Sr-Nd-Pb isotope systems (e.g., **Fig.4**,  
573 discussed in section 5.2.2.). Field studies mapping the basement terranes and crustal xenoliths  
574 in lavas indicate that, in addition to DMOB rocks, both Waipapa and Murihiku terrane  
575 greywacke metasediments are found beneath the AVF (**Fig. 1.B**: e.g., Kermode, 1992). Although  
576 no Os isotopic analyses exist for these terranes specifically, similar greywacke metasediments  
577 have been studied in Australia (Saal et al., 1998) and India (Wimpenny et al., 2007), and  
578 generally have highly radiogenic  $^{187}\text{Os}/^{188}\text{Os}$  of 1.2832 to 5.1968 and low Os concentrations of  
579 40-100 ppt (Wimpenny et al., 2007). Os values similar to those exhibited by Wiri Mt AU43931 is  
580 chosen as representative of the mantle melt as it plots (along with Rangitoto sample AU59309)  
581 within the OIB field for Os concentrations, Os isotopic and Pb isotopic values (**Fig. 6. and 7.**),  
582 but does not have the analytical error that is associated with the Rangitoto sample (**Table 1**).

583 Modelling suggests that  $\leq 1\%$  bulk assimilation of greywacke metasediments with Os =  
584 44.5 ppt and  $^{187}\text{Os}/^{188}\text{Os} = 1.2832$  (Saal et al., 1998) into a mantle melt ( $^{187}\text{Os}/^{188}\text{Os} = 0.1283$ ; Os  
585 = 194 ppt; Wiri Mt value from this study), coupled with  $\leq 5\%$  fractional crystallisation, can  
586 reproduce the range of observed radiogenic Os isotope signatures (and low Os contents) in the  
587 AVF samples (**Fig 8.**). Variations in mantle source signature between  $^{187}\text{Os}/^{188}\text{Os} = 0.1$  and 0.15,  
588 and Os content 50 to 200 ppt (to simulate heterogeneity in the mantle) have little impact on the  
589 percentage contribution from the contaminant, which remains at  $\leq 1\%$  for these ranges (details  
590 can be found in the supplementary material). Such low percentages of crustal contamination  
591 have little to no effect on the Sr-, Nd-, or Pb-isotope signatures as modelled in Figure 4.

592

#### 593 **5.4. Implications for magma origin, generation and ascent**

594 The AVF  $^{187}\text{Os}/^{188}\text{Os}$  and Os contents fall into the range of  $^{187}\text{Os}/^{188}\text{Os}$  and Os contents  
595 from other intraplate volcanic fields (e.g., Newer Volcanic Province; Central European Volcanic  
596 Province) and oceanic islands (e.g., Canary Islands; Cook Austral Islands, Comores; Hawaiian  
597 Islands, **Fig. 6.**). Although the formation of large oceanic island volcanoes, such as the Canary or  
598 Hawaiian Islands has mainly been attributed to the presence of relatively stationary thermal  
599 anomalies in the mantle (e.g., Morgan, 1971; Bennett et al., 1996; Montelli et al., 2006), the  
600 origin of small-scale intraplate volcanism is much less clear. Models to explain the origin of  
601 small continental intraplate volcanoes or volcanic fields include local lithospheric extension  
602 (e.g., Weaver and Smith, 1989; McCoy-West et al., 2013), lithospheric delamination (e.g., Jull and  
603 Kelemen, 2001; Elkins-Tanton, 2007), or edge driven convection (e.g., King, 2006). The  
604 processes driving melt generation beneath the Auckland and other Quaternary Volcanic Fields  
605 on North Island, New Zealand is a matter of current debate, but there is a general consensus that

606 extension-related magmatism is involved (Weaver and Smith, 1989; Smith et al., 1993; Huang et  
607 al., 1997; Cook et al., 2005; Needham et al., 2011).

608 As a consequence of long-lasting exposure of the lithospheric mantle to subduction  
609 metasomatism (through the influx of fluids during the Mesozoic), and plume-related  
610 magmatism during the Cretaceous, Huang et al. (1997) and Cook et al. (2005) argue that both  
611 the 'young' HIMU and EMII-type signatures observed in the volcanic field geochemistry  
612 originate in a partially melted zone in the lithospheric mantle. Conversely, McGee et al. (2013,  
613 2015) suggest that the 'young' HIMU-type signature is linked to carbonated domains within the  
614 upper asthenosphere. Between 70-90 km depth beneath the Auckland region seismic  
615 experiments revealed a low velocity zone, which is interpreted to represent regions of partial  
616 melt within the mantle caused by crustal extension following rollback of Pacific Plate  
617 subducting beneath the North Island of New Zealand (Horspool et al., 2006). Beneath South  
618 Island New Zealand, the base of the lithosphere has been estimated to lie at ca. 100 km (e.g.,  
619 Molnar et al., 1999), which would place this proposed partially melted zone within the  
620 lithospheric mantle. However, because this lithospheric thickness of ca. 100 km has been  
621 determined on the Pacific Plate (Molnar et al., 1999), there is some uncertainty as to how  
622 representative this thickness is of the Auckland area, which sits on the Indo-Australian Plate  
623 (e.g., Stratford and Stern, 2004). Regardless, regional lithospheric extension events may  
624 sufficiently thin the lower lithosphere to facilitate regional influx of hot MORB-like  
625 asthenospheric mantle and subsequent partial melting of the uppermost asthenospheric and  
626 lithospheric mantle via decompression and heating.

627 As previously outlined, the geochemical composition of lavas from Rangitoto is  
628 consistent with their derivation from the metasomatised lithosphere (e.g., Needham et al., 2011;  
629 McGee et al., 2015). Accordingly the lowest  $^{187}\text{Os}/^{188}\text{Os}$  of 0.123 in lavas from Rangitoto could be  
630 representative of the  $^{187}\text{Os}/^{188}\text{Os}$  of the lithospheric mantle beneath the Auckland Volcanic  
631 Field. This would be consistent with the unradiogenic  $^{187}\text{Os}/^{188}\text{Os}$  values of peridotite xenoliths  
632 from Zealandia (ca. 0.12-0.13; McCoy-West et al., 2013; Liu et al., 2015), however the error  
633 associated with this measurement mean constraining an accurate value is problematic.  
634 Alternatively, the low-silica lavas from Wiri Mt. show  $^{187}\text{Os}/^{188}\text{Os}$  of 0.128 similar to the Os  
635 isotopic composition of peridotite xenoliths in the  $1.68 \pm 0.15$  Ma intraplate Ngatutura Volcanics  
636 (e.g., Briggs et al., 1994). Based on the major and trace element and Sr-, Nd-, and Pb-isotopic  
637 composition in lavas from Wiri, McGee et al. (2013) proposed a predominately garnet-bearing  
638 peridotitic source for these lavas ( $\pm$  carbonate; McGee et al., 2015). However, because Wiri is  
639 located above the DMOB, we cannot be sure that the  $^{187}\text{Os}/^{188}\text{Os}$  in these lavas are

640 representative of the present day mantle beneath the AVF, or reflect minor contamination from  
641 the DMOB.

642         Regardless of mantle melt generation complexities beneath the AVF most of our new Os  
643 isotope data are more radiogenic than typical mantle values, suggesting that the ascending  
644 melts assimilated radiogenic  $^{187}\text{Os}/^{188}\text{Os}$  isotope signatures from the metasedimentary  
645 Murihiku or Waipapa terranes beneath Auckland. The exceptions to this are the olivine-bearing  
646 samples from Pupuke (and potentially Wiri Mt.), indicating the significant assimilation of  
647 sulphide-bearing olivine from the subsurface DMOB (Eccles et al., 2005). Of note is that both  
648 Pupuke and Wiri Mt. are located near or on faults separating the DMOB from adjacent rocks of  
649 the Murihiku Terrane (c.f. **Fig. 1.C**; Eccles et al., 2005), which suggests olivine accumulation  
650 beneath the AVF could occur while magma ascends along faults cutting or bounding the  
651 ultramafic rocks of the DMOB (Eccles et al., 2005). Generally terrane boundaries represent  
652 crustal (and potentially lithospheric) weak-zones, which are commonly associated with shear  
653 zones and intense faulting (e.g., Smith and Mosley, 1993). Because all three main Quaternary  
654 intraplate volcanic field in the North Island of New Zealand are roughly spatially correlated to  
655 the position of the DMOB, and the crustal-scale Taranaki fault (e.g., Giba et al., 2010; **Fig. 1.C**),  
656 we suggest that feeder melts may exploit the faults and shear-zone along this major terrane  
657 boundary to ascend to the surface (**Fig. 9**) rapidly rather than forming large crustal magma-  
658 reservoirs. Even though ascent may be rapid the high solubility of sulphides in basaltic silicate  
659 melt provides an effective way of contaminating rising melts with crustal sulphides.

660

## 661 **6. Conclusions**

662         In summary, we show that most of the AVF samples studied show Os isotopic  
663 compositions higher than typical mantle values ( $\geq 0.13$ ) and Os contents lower than 50 ppt.  
664 Conversely, a minority of samples (from Pupuke, Wiri and Rangitoto) show unradiogenic Os  
665 isotopic ratios coupled with high ( $>150$  ppt) Os concentrations. Neither of these signatures  
666 shows any obvious correlation to the tracers of mantle sources, implying that they are not  
667 primarily caused by source heterogeneity, nor do they show co-variance with elements  
668 dominant in olivine or pyroxene minerals, suggesting that they are not primarily caused by  
669 fractional crystallisation. Contrary to previous interpretations our results suggest that AVF  
670 melts do interact with continental crust during ascent leading to contamination signatures.  
671 From these two signatures two differing sources of crustal contamination are identified; 1)  
672 metasediments (e.g., Waipapa and Murihiku Terranes) which contain highly radiogenic Os  
673 coupled with minimal Os concentrations and 2) xenocrystic olivine hosted sulphides (from the

674 DMOB) with unradiogenic mantle-like  $^{187}\text{Os}/^{188}\text{Os}$  and high Os contents. Less than 1%  
675 contamination from the crust coupled with  $\leq 5\%$  fractional crystallisation, and  $\leq 10\%$   
676 contamination from xenocrystic olivine-bearing sulphide is sufficient to reproduce the Os  
677 contents and isotopic compositions observed in the AVF rocks.

678 The presence of mantle-derived sulphide hosted in olivines in eruptive products from  
679 centres located above the DMOB strongly argues for their derivation from crustal levels. We  
680 therefore propose that, although there is no evidence for long residence time of melts within the  
681 crust, crustal contamination does occur beneath most of the AVF. However, the low amount of  
682 crustal contamination is insufficient to affect the primitive geochemical and Sr-, Nd- and Pb-  
683 isotopic signatures of the AVF products.

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696

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1095 **Figure Captions**

1096 **Figure 1.** Schematic map of the Auckland Volcanic Field (AVF). (A) Location of the AVF within the North  
1097 Island, New Zealand, inset shows the location of New Zealand in relation to the east coast of Australia. (B)  
1098 DEM map and location of the volcanic centres in the AVF, highlighted are those sampled for this study,  
1099 shaded zones show the proposed terranes at 1.5 km depth (from Eccles et al., 2005). The green dashed line  
1100 (A&B) indicates the Dun Mountain Ophiolite Belt identified by the Junction Magnetic Anomaly (JMA) (Eccles  
1101 et al., 2005). (C) Schematic diagram adapted from Kermode (1992) and Eccles et al. (2005) outlining the  
1102 basement geology beneath the AVF: cross section A-A' is shown in diagram A.

1103  
1104 **Figure 2.** Diagram panels showing major elements vs. MgO (wt.%) for studied samples. Grey symbols  
1105 represent previously published data from the SAVF, with squares denoting Group A and crosses denoting  
1106 Group B samples, as assigned by Cook et al. (2005); and circles showing data from Briggs et al. (1994), which  
1107 were ungrouped. Also shown in the SiO<sub>2</sub> vs. MgO panel is the impact of mixing olivine into the samples,  
1108 showing that the addition of 5% Dun Mountain derived olivine (values from Sano and Kimura, 2007) could  
1109 produce the Pupuke sample signatures.

1110  
1111 **Figure 3.** Diagram showing representative primitive mantle-normalised multi-element compositions to  
1112 present the AVF trace element data range (McGee et al., 2013). The end members are exemplified by Purchas  
1113 Hill and Rangitoto 2. Wiri Mt. samples show intermediate trace element ratios. The grey field marks the  
1114 range from the AVF, black line shows a typical OIB-type signature (OIB values from Sun and McDonough,  
1115 1989, and normalisation values from McDonough and Sun, 1995), and blue dashed line show the  
1116 composition of the Waipapa terrane metasediments (from Price et al., 2015).

1117  
1118 **Figure 4.** <sup>206</sup>Pb/<sup>204</sup>Pb vs. (A) <sup>207</sup>Pb/<sup>204</sup>Pb, (B) <sup>208</sup>Pb/<sup>204</sup>Pb, (C) <sup>87</sup>Sr/<sup>86</sup>Sr, and (D) <sup>143</sup>Nd/<sup>144</sup>Nd for all samples  
1119 from the Auckland Volcanic Field (this study and McGee et al., 2013), and South Auckland Volcanic Field  
1120 (Cook et al., 2005). The inset shows the mantle reservoir end members from Zindler and Hart (1986):  
1121 Enriched Mantle (EMI and EMII), Pacific MORB Mantle, and HIMU. Values for Northland from Huang et al.  
1122 (2000), for Otago from Timm et al. (2010), for Banks Peninsular from Timm et al. (2009), and for Lookout  
1123 Volcanics from McCoy-West et al. (2010), for Waipapa metasediments from Price et al. (2015). Modelling  
1124 lines show binary mixing of Waipapa metasediments (blue) and lithospheric mantle (EMII; green) into AVF  
1125 mantle-like source, with the shaded area showing the potential signatures caused by these inputs.

1126  
1127  
1128 **Figure 5.** Os and Re concentrations versus MgO (in wt.%), Ni, Cu and Zr (all in ppm) contents for samples  
1129 analysed in this study. Symbols are as in **Figure 2**. Note that only 5 samples were analysed by this study for  
1130 major and trace elements. Previous studies (McGee et al., 2013) did not present Cu data for these samples,  
1131 therefore a reduced number of samples are plotted in Cu vs. Os and Re plots.

1132  
1133 **Figure 6.** <sup>187</sup>Os/<sup>188</sup>Os vs Os (ppt) for all samples from the AVF. Also shown in grey symbols are global OIB  
1134 values for Pacific regions: Austral-Cook (Hauri and Hart, 1993; Reisberg et al., 1993; Hanyu et al., 2011);  
1135 Samoa ( Jackson and Shirey, 2011); Newer Volcanic Province, Australia (NVP; McBride et al., 2001);  
1136 Louisville Seamount Chain (Tejada et al., 2015), and for the Atlantic regions: the Canary Islands  
1137 (Marcantonio et al., 1995; Widom et al., 1999; Day et al., 2009); St Helena and Comores (both from Reisberg  
1138 et al., 1993); Cape Verde Islands (Escrig et al., 2005); Azores (Widom and Shirey, 1996; Larrea et al., 2014);  
1139 Central European Volcanic Province (CEVP; Jung et al., 2011). Error bars are shown for one Rangitoto  
1140 sample, for all other analyses the errors are smaller than the symbol size.

1141  
1142 **Figure 7.** Diagram after Day (2013), to show mixing relationships for various potential contaminants in  
1143 (A) <sup>206</sup>Pb/<sup>204</sup>Pb, (B) <sup>143</sup>Nd/<sup>144</sup>Nd, and (C) <sup>87</sup>Sr/<sup>86</sup>Sr vs. <sup>187</sup>Os/<sup>188</sup>Os isotope space. Data values for MORB and  
1144 OIB from Widom et al. (1999), Hofmann (1997), and Day et al. (2010); pelagic sediments from Roy-Barman

1145 and Allègre (1995) and Eisele et al. (2002); carbonatite from Widom et al. (1999) and Escrig et al. (2005);  
1146 young HIMU from Day et al. (2009); and continental crust from Saal et al. (1998) and Widom et al. (1999).

1147  
1148 **Figure 8.** Whole-rock  $^{187}\text{Os}/^{188}\text{Os}$  vs Os concentration (ppt) for AVF samples with proposed methods of  
1149 signature formation. Sulphide assimilation is modelled (green) using binary mixing from AVF mantle values  
1150 with values measured for sulphides found in dunite within Shetland Ophiolite complex ( $^{187}\text{Os}/^{188}\text{Os} = 0.1353$ ,  
1151  $\text{Os} = 4150$  ppt; O'Driscoll et al., 2012). Fractional crystallisation is modelled (blue) using the Rayleigh  
1152 equation for samples from 50 to 7 ppt, showing that it requires  $\leq 10\%$  fractional crystallisation to reduce the  
1153 Os concentrations. Crustal assimilation is modelled (orange) through binary mixing of post 10% fractional  
1154 crystallisation of the mantle-derived melt ( $^{187}\text{Os}/^{188}\text{Os} = 0.1283$ ,  $\text{Os} = 7$  ppt) with  $< 10\%$  crust, with values  
1155 reflective of greywacke ( $^{187}\text{Os}/^{188}\text{Os} = 1.283$ ,  $\text{Os} = 44.5$  ppt; Saal et al., 1998). Assimilation fractional  
1156 crystallisation (AFC) is modelled (after DePaolo, 1981) with  $D_{\text{Os}} = 20$  (Widom et al., 1999),  $r$  (rate of  
1157 fractional crystallisation) = 0.95, for mantle values ( $^{187}\text{Os}/^{188}\text{Os} = 0.1283$ ,  $\text{Os} = 194$  ppt (Wiri sample) with  
1158 crustal values of greywacke (black line,  $^{187}\text{Os}/^{188}\text{Os} = 1.283$ ,  $\text{Os} = 44.5$  ppt; Saal et al., 1998), and carbonatite  
1159 (grey line,  $^{187}\text{Os}/^{188}\text{Os} = 0.6$ ,  $\text{Os} = 15$  ppt; from Widom et al., 1999 and Escrig et al., 2005).

1160  
1161 **Figure 9.** Schematic model to illustrate the proposed magma ascent pathways for the AVF eruptions. Melts  
1162 are derived from a heterogeneous source, including ambient peridotite mantle containing HIMU-like  
1163 carbonated peridotite veins (McGee et al., 2015) at a depth of  $> 80$  km, and a subduction metasomatised  
1164 lithosphere at  $< 80$  km depth, all of which give the ascending magma its Sr-, Nd-, Pb-isotopic and major and  
1165 trace element signatures (McGee et al., 2013). Minimal fractionation occurs on ascent, efficiently reducing  
1166 the Os concentration from mantle values, followed by preferential assimilation of olivines from the Dun  
1167 Mountain Ophiolite Belt (DMOB) as xenocrystic material causing increase in Os concentration, and finally  
1168 minor crustal contamination of magmas causing the radiogenic isotope signatures. A crustal depth of 20-30  
1169 km and the positions of the low velocity melt zone region are from tomography by Horspool et al. (2006),  
1170 and upper cross section is adapted from Seebeck et al. (2014).

1171

## 1172 Table Caption

1173 **Table 1.** Selected geochemical data for AVF centres, \* denotes data from lavas previously analysed by McGee  
1174 et al. (2013). Suffixes 'i, ii, iii' etc denote duplicate analyses from the same sample but with a different  
1175 digestion, (2) denotes duplicate analysis of the same sample and same digestion.

1176

**Supplementary Material**

**SM1** – All major and trace element, and isotopic values for samples analysed used in this study.

**SM2** – All standard data for analysis of major, trace and isotopes.

**SM2.1.** Major elements by XRF.

**SM2.2.** Trace elements by solution ICP-MS.

**SM2.3.** Pb isotope analysis by multi-collector ICP-MS.

**SM2.4.** Os isotope analysis by TIMS.

**SM2.5.** Re isotope analysis by sector field ICP-MS.

**SM3** – Modelling parameters for the contamination effects on other isotope systems

**SM3.1.** Values used for modelling

**SM3.2.** Calculations for  $^{206}\text{Pb}/^{204}\text{Pb}$  vs  $^{87}\text{Sr}/^{86}\text{Sr}$

**SM3.3.** Calculations for  $^{206}\text{Pb}/^{204}\text{Pb}$  vs  $^{143}\text{Nd}/^{144}\text{Nd}$

**SM3.4.** Calculations for  $^{206}\text{Pb}/^{204}\text{Pb}$  vs  $^{208}\text{Pb}/^{204}\text{Pb}$

**SM3.5.** Calculations for  $^{206}\text{Pb}/^{204}\text{Pb}$  vs  $^{207}\text{Pb}/^{204}\text{Pb}$

**SM4** – Modelling parameters for AFC calculations

**SM4.1.** Optimum values used for plotting figure 8.

**SM4.2.** Variations in mantle source signature.

**SM4.3.** Variations in r value.