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1 **An insight into the first stages of the Ferrar magmatism: ultramafic cumulates**
2 **from Harrow Peaks, northern Victoria Land, Antarctica.**

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17 Key words : ultramafic xenoliths ; high-Mg magmatic olivines ; orthopyroxenite ; Karoo-
18 Ferrar large igneous province .

19
20
21
22 **Abstract**

23
24 Ultramafic xenoliths hosted in Cenozoic hypabyssal rocks were collected from Harrow Peaks
25 (northern Victoria Land, Antarctica) and analysed for in situ (major and trace elements) geochemical
26 purpose. Samples are generally small (<10 cm) and consist of spinel bearing lherzolites,
27 harzburgites, composite xenoliths and one orthopyroxenite, the latter is a lithotype rarely observed
28 in mantle xenoliths of this region and elsewhere in Antarctica. Most of these rocks show
29 equigranular texture with the olivine's boundaries often converging at 120° triple points. The
30 composite xenoliths are formed by large clinopyroxene and/or amphibole veins and a dunitic matrix,
31 characterised by olivine with forsterite contents from 87.46 to 90.21.

32 Both lherzolites and harzburgites have olivine modal contents that relate to spinel (Cr#: 24.79-
33 63.28) and olivine (Fo: 84.24-88.39) compositions, which do not reflect any residual trend due to a
34 melting process. In turn, all the studied Harrow Peaks xenoliths (including the orthopyroxenite)
35 have olivine ascribed to a cumulus phase, and the presence of the orthopyroxenite suggests that the

36 inferred melt from which they stemmed, was close to silica saturation. In this view, the geochemical
37 characteristics of lherzolites/harzburgite, dunites and orthopyroxenite were modelled as cogenetic
38 products of a crystallising process occurring at the mantle-crust boundary depth.
39 The initial crystallisation of olivine from a high Mg-primitive melt (Mg# 72-picrites?) is the main
40 forming process of this group of xenoliths, and the potential temperature of crystallisation was
41 estimated at ~1300 °C, based on the olivine/melt equilibrium. Fe/Mg and REE mineral-melt
42 equilibrium models proved that the Harrow Peaks melts could be related to the initial stage of the
43 Jurassic Ferrar magmatism. These cumulitic fragments stationed in the shallow mantle and were
44 subsequently affected by a late stage alkaline metasomatism, sporadically detected in the northern
45 Victoria Land lithosphere. The metasomatic process formed the LREE-enriched alkaline-type
46 amphiboles, with a complete geochemical disequilibrium with the anhydrous (dunite and
47 orthopyroxenite) matrix.
48 During the Cenozoic, the xenoliths were brought to the surface by alkaline lavas belonging to the
49 McMurdo volcanism, connected to the West Antarctica Rift System opening.

50

51 **1. Introduction**

52

53 The West Antarctic Rift system (WARS) is one of the largest continental rift areas in the world and
54 is comparable to the East African Rift in scale (Martin et al., 2015; Le Masurier and Thomson 1990).
55 Since the Cenozoic it was affected by alkaline magmatism represented by the products of the Mc.
56 Murdo volcanic and Meander intrusive groups (Kyle 1980). The most primitive volcanic products
57 from Victoria Land contain mafic and ultramafic xenoliths in a large spectrum of lithologies that
58 testify for a complex mantle/crust evolution of the Subcontinental Lithospheric Mantle (SCLM).
59 Evidence of mantle partial melting and enrichment events (due to both alkaline metasomatism and
60 refertilisation by tholeiitic melts) as well as crystallization of alkaline melts at the Moho depth have
61 been largely documented (Gamble et al., 1988; Beccaluva et al., 1991; Cooper et al., 2007; Martin

62 et al., 2015; Pelorosso et al., 2016). Geochemical and isotopic data trace back the refertilisation
63 process of SCLM in the northern Victoria Land prior to the Cenozoic (Melchiorre et al., 2011;
64 Pelorosso et al., 2016), suggesting a possible connection with the most important magmatic activity
65 in Antarctica, which occurred before the Cenozoic: the Jurassic magmatism, part of the Karoo-
66 Ferrar Large Igneous Province. The Jurassic magmatism produced outcrops that stretch over 3500
67 km from the Theron Mountains of Antarctica to southeast Australia (Elliot & Fleming 2004),
68 including several volcanic bodies such as mafic sills, flood basalts (i.e. Kirkpatrick basalt, Vestfjella
69 ferropicrites), phreatomagmatic volcanic rocks, layered mafic intrusions (i.e. Ferrar dolerite sill)
70 and mafic dykes (Kyle et al., 1989; Fleming 1995; Elliot et al., 1999; Storey et al., 2013; Heinonen
71 and Luttinen, 2010; Bedard et al., 2007). In the Harrow Peaks (74.0 2785°S 164.47466°E, [Fig. 1](#)),
72 where the alkaline Cenozoic magmatism is dominant as well as in many other northern Victoria
73 Land magmatic localities, lavas brought to the surface ultramafic xenoliths. The samples available
74 for this study were collected during the XX Italian Expedition organised by PNRA (Programma
75 Nazionale Ricerche in Antartide) in the 2004/05 Austral summer. They are mainly (amphibole
76 bearing) spinel peridotites, but composite xenoliths are also abundant. The Harrow Peaks samples
77 include dunites and one orthopyroxenite, which are rarely observed or absent in the xenolith
78 populations sampled in the region and carried by the same magmatic system.

79 A few of these xenoliths were studied by Gentili et al. (2005) in order to evaluate the role of
80 amphibole formation in the modification of the original redox condition of this mantle domain. The
81 quoted authors revealed extremely high redox conditions ($\Delta\log QFM fO_2 = \sim +5, +7$) for the
82 amphibole formation, in clear disequilibrium with the peridotite matrix.

83 Here we present new petrological and geochemical data of this unique group of ultramafic xenoliths,
84 which provide evidence for a magmatic origin and show to what extent these rocks have been
85 affected by metasomatic event(s), and their relationships with Victoria Land SCLM.

86

87 **2. Sample description and Petrography**

89 The ultramafic xenoliths found in the Harrow Peaks lavas are rather small (2 to 5 cm) and sub-
90 rounded in shape; samples showing textural and geochemical evidence of host-basalt contamination
91 (Bonadiman et al., 2011) were included in the study for a comparative purpose.

92 Three out of the eight samples selected for this study are composite xenoliths. Hydrous phases
93 (mainly amphibole) occur in all samples, showing significantly different grain size and modal
94 contents from sample to sample.

95 The mineral modal proportion was determined by point counting, averaging two runs with more
96 than 2,000 points for each thin section (2.5 x 4.0 cm): the results were confirmed by thin section
97 colour code analyses (ImageJ© software) and reported in [Table 1](#). Based on the classification
98 diagram for mafic and ultramafic rocks ([Fig. 2](#); Streykeisen, 1974), the Harrow Peaks xenoliths
99 consist of spinel-bearing lherzolites (HP151, HP164 and HP166), one harzburgite (HP144) and one
100 orthopyroxenite (HP163).

101 The composite xenoliths combine various lithologies that make difficult to apply the standard
102 nomenclature for ultramafic rocks. HP124 and HP143 are made up of a dunite matrix crossed by
103 clinopyroxene (HP143, [Fig. 3a](#)) and clinopyroxene + amphibole + rare phlogopite (HP124, [Fig. 3b](#))
104 veins; HP121 is also composite with a harzburgite matrix (for containing up to 6% and 4% of modal
105 orthopyroxenes and clinopyroxenes, respectively), crosscut by a large amphibole vein ([Fig.3c](#)). The
106 textural similarities and the high olivine modal contents ([Fig. 2](#); [Table 1](#)) led us to equate HP121
107 sample with the rest of the composite dunites.

108 Regardless of the lithology, most of the samples exhibit tabular to equigranular textural types
109 (Mercier & Nicolas 1975) with equidimensional, polygonal in shape olivines (1 mm) delimited by
110 boundaries often converging at 120° (i.e. [Figs. 4a, b](#)). The modal content of the orthopyroxene is
111 higher in lherzolites (18 - 26 volume. %) with respect to the harzburgite (13 vol. %). The
112 clinopyroxene in lherzolites/harzburgite is small and interstitial ([Fig.4e](#)); small irregular spinel is
113 scattered within the olivine grains. Amphibole is present both as disseminated crystals (i.e. HP124;

114 Fig. 4b) or in veins (i.e. HP121; Fig.4d), and it is associated with rare phlogopite (HP124). The
115 “layered” textural effect evidenced in some samples (i.e. HP143 or HP151; Figs. 3a, e) is marked
116 by a trail of small irregular spinels within the olivine grains, and/or by the amphibole/clinopyroxene
117 veins in clear textural disequilibrium with the dunite matrix (HP121; HP124 and HP143).
118 The predominant equigranular texture exhibited by Harrow Peaks xenoliths is completely absent in
119 most of the xenolith suits from the same volcanic district (i.e. Baker Rocks and Greene Point), where
120 the protogranular texture is commonly observed. In addition, secondary textures (i.e. glassy patches
121 and sieved pyroxenes, spongy rims) are rarer with respect to other Antarctic xenolith occurrences
122 (Coltorti et al., 2004; Pelorosso et al., 2016). They consist of glassy thin menisc between minerals
123 (including disseminated amphiboles) or small pools (<100 microns) containing microlites of olivine,
124 spinel and clinopyroxene (HP124, HP143, HP144, and HP151).

125

126 3. Analytical Methods

127

128 Bulk major and trace element analyses were precluded due to the small size of the xenoliths.
129 Instead, thin and thick sections were prepared for in situ analytical protocols.
130 Major element compositions of minerals and glass were determined by a CAMECA SX100 electron
131 microprobe equipped with four wavelengths dispersive (WD) and one Energy Dispersive (ED)
132 spectrometer, at the Department of Lithospheric Research, University of Wien (Austria). The
133 operating conditions were 15 kV accelerating voltage, 20 nA beam current and 20 s counting time
134 on peak position.
135 In order to minimise Na and K loss, a 5µm defocused beam and 10 s counting time on peak position
136 were applied for glass analyses. Natural and synthetic standards were used for calibration and PAP
137 corrections were applied to the intensity data (Pouchou and Pichoir, 1991). The concentrations of
138 trace elements in pyroxene and glass were obtained by a Laser Ablation Microprobe-Inductively

139 Coupled Plasma Mass Spectrometry (LAM-ICP-MS) at Geosciences Montpellier Université de
140 Montpellier and at igg CNR, Pavia. Both laboratories applied the same analytical protocol and
141 processing data system.

142 Each analysis took 120s: 60s background acquisition (gas blank) and 60s acquisition for the sample.
143 The analyses were corrected with internal standards using CaO for clinopyroxene and glass, and
144 SiO₂ for orthopyroxene. The detection limit is a function of the ablation volume and counting time,
145 and is therefore calculated for each analysis. The ablation volume, in fact, greatly depends on the
146 instrument configuration, consequently, the detection limit decreases if spot size, beam power and
147 cell gas flow are reduced. A beam diameter of 40-100 μm and a scanning rate of 20 μm/s were used.
148 The theoretical limit of detection ranges between 10 and 20 ppb for REE, Ba, Th, U, and Zr and 2
149 ppm for Ti.

150 Data were processed using the Glitter[®] software (van Achterbergh et al., 2001), and element
151 concentrations were calibrated against the NIST612 certified reference material, using the values of
152 Pearce et al. (1997).

153

154 **4. Mineral chemistry**

155

156 Major element data collected by electron microprobe and reported as wt % oxides are provided for
157 olivine, orthopyroxene, clinopyroxene and spinel in [Tables S1, S2 and S3](#), as well as for amphibole
158 in [Table S4 of Supplementary Material](#).

159

160 *4.1 Mineral major element compositions*

161

162 Olivine is the ubiquitous phase of the Harrow Peaks peridotites; however, despite the variety of the
163 lithological types and the presence of composite xenoliths, it has a relatively narrow range of
164 composition. The forsterite (Fo) content, calculated as $[\text{Mg}/(\text{Mg}+\text{Fe}_{\text{tot}}) \times 100 \text{ atomic formula units}]$,

165 varies from 86.60 to 90.45, and the NiO content from 0.26 to 0.47 wt %, reported as sample average
166 (Fig. 5a; Table S1 of Supplementary Material); Fo values of olivine (~3% in volume) in
167 orthopyroxenite HP163 and lherzolite HP164 are not included in this compositional range; they are
168 in the range of 80.45-85.88, reflecting the common chemical zoning of cumulus olivines (Fig. 5a;
169 Table S1 of Supplementary Material).

170 Orthopyroxene varies in the range of $\text{En}_{66.62-83.89}\text{Fs}_{0.19-0.21}\text{Wo}_{0.038-0.17}$ with Mg# [Mg# = molar
171 $\text{Mg}/(\text{Mg}+\text{Fe}_{\text{tot}}) \times 100$] = 83.27-89.60 showing an apparent equilibrium with the coexisting olivine in
172 all the investigated samples, and remarkable differences from those of the residual peridotites.
173 Orthopyroxene presents Al_2O_3 contents mostly in the range of 1.24 to 3.32 wt %, and Cr_2O_3 and
174 TiO_2 (as sample compositional averages) always <0.50 and <0.20 wt %, respectively (Fig. 5b; Table
175 S2 of Supplementary Material).

176 Spinel is chemically heterogeneous, varying in composition from sample to sample without any
177 clear relationship with the lithological types. In composite xenoliths HP124 and HP143 it shows
178 high variable Cr# [Cr# = molar $\text{Cr}/(\text{Cr} + \text{Al}) \times 100$] (24.25-61.69) and Mg# (60.81-52.26) values,
179 with TiO_2 (0.5- 1.04 wt%) contents approaching the spinel composition of HP163. HP121 contains
180 relatively low-Ti spinel ($\text{TiO}_2=0.26-0.36$ wt %) with the lowest Fe_2O_3 (0.71 wt % as sample average)
181 recorded by the Harrow Peaks xenolith population (Fig.5c; Table S3 of Supplementary Material).

182 Spinel from harzburgite HP144 and lherzolite HP151 present the highest and the lowest Cr# values,
183 respectively, at similar low Fe_2O_3 (2.54-3.71 wt%) and (lowest) TiO_2 (<0.22 wt %) (Fig. 5c). The
184 spinel composition in terms of Cr# is also compared to the Fo content in the Arai (1994, Fig. 6) in
185 which Harrow Peaks samples plot outside the mantle array theoretical field.

186 Clinopyroxene occurs in this group of ultramafic xenoliths in evident textural disequilibrium with
187 the olivine + spinel + (orthopyroxene) assemblage. It is Cr-rich augite following the Morimoto
188 (1989) classification scheme, and heterogeneous in composition from one grain to another.
189 Irrespective of lithology, Mg# values span from 85.63 to 93.89 with $\text{TiO}_2 < 0.55$ wt % and $\text{Al}_2\text{O}_3 <$
190 4.13 wt % as sample averages (Figs., 7a, b; Table S4 of Supplementary Material). Only a few grains

191 of clinopyroxene concentrated in veins of both HP124 and HP143 dunites have TiO₂ and Al₂O₃
192 contents that reflect a host basalt low-pressure perturbation (Figs. 7 a, b). Among the possible
193 exchanging cations in M1 and M2 clinopyroxene sites, the only correlation observed is the
194 progressive Na enrichment at lower Mg#.

195 Amphibole is texturally related to the clinopyroxene and generally occurs as vein in dunites, while
196 it is small and disseminated in lherzolites and the orthopyroxenite. It is chemically classified as
197 kaersutite (HP121, HP143 and HP163), magnesio-hastingsite (HP124) and ferri-kaersutite (HP164),
198 following the Leake et al. (1997) scheme, with Mg# ranging from 82.86 to 90.39 and TiO₂ and
199 Al₂O₃ contents varying from 1.05 to 4.19 wt% and 11.43 to 14.86 %, respectively (Gentili et al.,
200 2015, Figs. 7c, d; Table S5 of Supplementary Material). The Harrow Peaks amphibole group had
201 already been investigated for a crystallochemical study that led to the discovery of a new amphibole
202 type: the ferri-kaersutite (Gentili et al. 2015). Based on the crystallochemical model of Oberti et al.
203 (2000), Harrow Peaks amphibole have lattice parameters primary ascribable to the magmatic-type
204 (Gentili et al. 2015), whereas amphiboles from the rest of Antarctic xenoliths have chemical and
205 structural characteristics that fall within the “mantle-type” (Oberti et al. 2000; Bonadiman et al.,
206 2014).

207 Phlogopite grains occurring in HP124 appear chemically zoned (Mg# ~ 91-92), coherent with the
208 observed variability of the other phases forming the vein. They have TiO₂ and Al₂O₃ contents
209 between 1.41-2.19 wt% and 13.88-15.86 wt%, respectively; alkalis are instead constant (Na₂O ~0.60
210 wt%, K₂O ~ 9.60 wt%, Table S5 of Supplementary Material).

211 Glass is extremely variable in composition (Table S6 of Supplementary Material) with SiO₂ contents
212 ranging from 56.21 to 68.35 wt %, in relation to the textural position and adjacent mineral type, as
213 well as the xenolith's lithology.

214

215 *4.2 Pyroxene trace elements*

216

217 Trace element concentrations of pyroxenes averaged *per* sample are reported in [Table 2](#), whereas
218 all the LA-ICP-MS data are in [Tables S7 –S10 of Supplementary Material](#).

219 In the chondrite-normalised REE+Sr+Zr+Hf+Ti+Y diagram, orthopyroxene displays a H-MREE
220 downward trend, with the slope increasing in the Yb and Lu region and with an ubiquitous positive
221 Ti anomaly ([Fig. 8](#)). HP121 and HP143 are composite xenoliths with orthopyroxene-bearing dunite
222 matrix, but only HP143 contains a few orthopyroxenes large enough for in situ trace element
223 analyses. In detail, the chondrite-normalised trace element distribution of HP143 orthopyroxene
224 mimics that of the HP164 lherzolite (clinopyroxene modal contents ~11 in volume %) and, to a
225 lesser extent, that of the HP151 lherzolite (clinopyroxene ~9 in volume %), with $Yb_N/Gd_N = 4.16$
226 and remarkable Ti ($Ti/Ti^* = 5.6$) positive and Sr ($Sr/Sr^* = 0.84$) negative anomalies. HP166 lherzolite
227 (clinopyroxene ~7 in volume %) displays orthopyroxene with high Zr, Hf (and Ti) and M-HREE
228 contents, depicting a M-HREE chondrite normalised profile ~7 times higher, but LREE and LILE
229 contents comparable to the orthopyroxene in lherzolites (and dunite). HP166 orthopyroxene shows
230 coupled strong Ti and Zr (Hf) positive anomalies. In turn, orthopyroxene in HP144 harzburgite
231 (clinopyroxene ~3 and orthopyroxene ~13 in volume %) contains the lowest M-HREE (plus Ti),
232 but the highest LREE and LILE concentrations of the entire xenolith population ([Table 2](#)),
233 exhibiting almost flat or slightly L-REE enriched patterns ($Ce_N/Nd_N \sim 2$), ([Fig. 8](#), [Table 2](#)).

234 Disseminated clinopyroxene is scarce in Harrow Peaks xenoliths. HP164 is the clinopyroxene-
235 richest sample with 11 volume % of modal content ([Table 1](#)). Clinopyroxene is extremely variable
236 in terms of trace element contents and, coherently with major elements behaviour, without any
237 correlation with the “fertile” character of the rock ([Fig. 9a](#); [Table 2](#)). In the chondrite normalised
238 incompatible multi-element variation diagram, it shows negative Sr and Ti anomalies. The large
239 compositional variability is even more evident in chondrite normalised REE diagrams ([Fig. 9b](#)) with
240 overall enriched (La_N 9.02-89.9 and Yb_N 3.30-20.0 times chondrite) patterns, ranging from almost
241 flat / slightly (i.e. HP143 $La_N/Yb_N = 2.36$) to highly (i.e. HP124: $La_N/Sm_N = 6.89$; $La_N/Yb_N = 12.53$)
242 LREE enriched. Strongly MREE enriched upward convex patterns (HP166:

243 $Gd_N/Yb_N=2.36; Gd_N/La_N=2.17$) can also be observed. The clinopyroxene of HP151 and HP166
244 lherzolites displays significant negative Eu anomalies (Fig. 9b).

245 Clinopyroxene (~7 volume%) in orthopyroxenite HP163 is compositionally variable; most of the
246 grains reproduce the convex upward pattern shown in HP166 lherzolite (at a lower degree of
247 enrichment, Fig. 9b), and a few small crystals evidence a progressive LREE enrichment ($La_N =$
248 15.4-179, Fig. 9b).

249 Clinopyroxene concentrated in veins in composite xenoliths (HP143, HP124) has generally low
250 HREE contents ($Yb_N=3.03$ for HP143 and 6.85 for HP124), with less pronounced or absent negative
251 Sr and Ti anomalies with respect to the disseminated clinopyroxene of lherzolites and harzburgite.
252 It shares (almost) flat chondrite-normalised REE distribution patterns, but different degrees of
253 LREE enrichments ($La_N/Yb_N=2.80-8.65$). As observed in lherzolites HP151 and HP166, a Eu
254 anomaly also characterises the small clinopyroxene in the amphibole-bearing vein of HP121
255 composite xenolith.

256

257 *4.3 Amphibole trace elements*

258

259 Amphibole is characterised by incompatible trace element chondrite-normalised patterns of similar
260 shape but at various degrees of trace element enrichment (i.e. $Ba_N=13.1-95.6; Nd_N=11.1-46.1$, Fig.
261 11a) and with systematic Nb and Ti positive anomalies (Fig. 10a). The trace element poorest
262 contents are recorded by the rare amphibole in the thick clinopyroxene vein of the HP143 composite
263 xenolith (Figs. 3a, 4c and Fig. 10b). They perfectly mimic the coexisting clinopyroxene REE pattern
264 (Fig. 10b). Conversely, the highest trace element concentrations are measured in veined amphibole
265 in HP121 composite xenolith that reflect a pure alkaline-like REE pattern (i.e. $\sim La_N 43.1, Nd_N 46.1,$
266 $Yb_N 11.3$). Veined amphibole in HP124 composite xenolith is also characterised by a gently convex
267 REE pattern (Fig. 10b), but depleted in LREEs with respect to the associated clinopyroxene. In
268 HP164 lherzolite, the disseminated amphibole roughly displays a pattern similar to the coexisting

269 clinopyroxene (Fig. 9b), but with higher LREE contents (i.e. amphibole: $L_{AN}= 32.2$; clinopyroxene
270 $L_{AN}=11.0$, Fig. 9b). Overall, the Harrow Peaks amphibole appears more heterogeneous and trace
271 element depleted with respect to that of the nearby Baker Rocks mantle xenolith population (Figs.
272 10a, b; Coltorti et al., 2004).

273

274 5. Discussion

275

276 Geochemical and textural features of Harrow Peaks ultramafic xenoliths highlighted remarkable
277 differences between this locality and others in Antarctica. Primarily, their modal compositions do
278 not depict any trend attributable to a common mantle residual path (Fig. 2; Bodinier and Godard,
279 2003; Workman and Hart, 2005) and secondly, the Fo content of olivine, in relation to the modal
280 content, is difficult to ascribe to i) residual (melting) trend defined by the mantle peridotites from
281 other localities in Antarctica (Fig.5a), or ii) alkaline cumulate rocks such as those sampled from Mt.
282 Overlord (~ 50 Km from Harrow Peaks, Perinelli et al., 2017) or in tholeiitic magmatic products
283 as those of the Basement Sills (Bedard et al., 2007, Fig. 5a). In addition, there is not an observed
284 correlation between spinel and olivine, as expected in a potential residual trend, with the sole
285 exception of the composite xenolith HP21, where spinel Fo and Cr# values of the dunitic matrix
286 intercept the mantle array curve in a region that reflect a moderate fertile lherzolite (melting degree
287 < 20%), inconsistent with the rock lithotype (Fig. 6).

288 Moreover, clinopyroxene, which is extremely low in modal content, tends to have Mg# values
289 comparable to those of the clinopyroxene from Antarctica mantle peridotites (Fig.7a, b), but without
290 any relationship with potential residual (i.e. decreasing of Al_2O_3 with the increasing of Mg#) or
291 metasomatic (i.e. enrichment in TiO_2 or Al_2O_3 at comparable Mg#) trends (Coltorti et al., 2004;
292 Perinelli et al., 2006; 2008; 2011; Armienti and Perinelli 2010; Pelorosso et al., 2016, 2017).

293 Thus, all the mentioned characteristics suggest that xenoliths from Harrow Peaks do not represent a
294 simple peridotite residuum after melt extraction. Consequently, we discuss the origin of these

295 xenoliths as potentially representing cumulates that crystallised from various silicate melts
296 migrating through the mantle and separated at mantle-crust boundary P-T conditions.

297

298 *5.1 Formation of Harrow Peaks xenoliths*

299

300 *T-P-fO₂* equilibration conditions of the Harrow Peaks xenolith suite have been previously explored
301 by Gentili et al. (2015), using a crystallochemical approach based on the amphibole
302 dehydrogenation. The study highlighted a strong discrepancy between the redox conditions recorded
303 by the amphibole dehydration equilibrium ($\Delta\text{QFM} = +5 - + 6.8$) and those of the coexisting
304 peridotite mineral assemblage ($\Delta\text{QFM} = - 2.78 - - 0.2$) finally equilibrated at $\sim 854 - 940^\circ\text{C}$
305 (calculated at $P = 1.5 \text{ GPa}$). Using the same approach, this decoupling was not detected for the
306 amphibole-bearing lherzolites and harzburgites of the nearby area of Baker Rocks where
307 amphiboles and peridotite matrix converge to $f\text{O}_2$ values between QFM and $\Delta\text{QFM} -1.78$
308 (Bonadiman et al., 2014).

309 We re-examined the Harrow Peaks samples in view of the suggested possible magmatic origin and
310 integrated previously determined thermobarometric results with new chemical data and new
311 samples.

312 Temperature values calculated using the combination of olivine-spinel Fe-Mg exchange
313 thermometers of Wood and Virgo (1989) and Ballhaus et al. (1991) in the lherzolites, harzburgite
314 and dunitic portions of composite xenoliths, suggest that these xenoliths finally equilibrated between
315 800 (orthopyroxenite HP163) and 1200°C (lherzolite HP151).

316 Presuming that the initial crystallisation of olivine is the main forming process of this group of
317 xenoliths, the potential temperature of crystallisation was also calculated based on olivine/melt
318 equilibrium, assuming that the Fe/Mg ratio of olivine (on sample average) was still close to the
319 initial crystallization conditions. This is reasonable since they were extracted from the forming
320 melts, and cumulated as monomineralic or bimineralic systems. Experimental determinations of Fe-

321 Mg diffusivity in such a system reveal that for an olivine composition of Fo₈₃₋₉₀ the closure
322 temperature is reached fast, and the estimated length scales for diffusion in a given time strongly
323 decreased (Chakraborty 1997; Costa and Dungan, 2005).

324 The FeO-MgO (and SiO₂) contents of the theoretical melt/s in equilibrium with dunitic portions of
325 composite xenoliths were inferred comparing various ^{Fe/Mg}Kd olivine/melt (Norman et al. 2002;
326 Putirka et al. 2008; Oeser et al. 2015) models. Using the equation [21] of Putirka (2008), and
327 assigning a maximum H₂O content of 1 wt %, the best fit obtained for the high Mg# HP121 is a melt
328 composition having Mg# 71.5. For HP124, HP143, the majority of lherzolites (Mg# 88) and the
329 harzburgite (Mg# 89), the melt in equilibrium (melt 2) is obtained by fractionation of ca. 4 % olivine
330 and 0.8 % of spinel from melt 1 (Tables 3, 4).

331 Furthermore, the olivine in the orthopyroxenite sample (Mg# 80.46) results to be in equilibrium
332 with a slightly evolved melt (melt 3) obtained by additional fractionation of ca. 5% of olivine, 4 %
333 of orthopyroxene, 1 % of clinopyroxene and 0.5 of plagioclase from melt 2 (Table 4).

334 Overall, the calculated melts show compositions that span between high-Mg tholeiites (picrites?)
335 and typical tholeiitic basalts, in a co-genetic evolution of the liquid line of descent. In this respect,
336 the HP121 dunitic portion represents the primitive crystallised product of the Kirkpatrick basalt
337 precursor (Kyle, 1980; Fleming 1995). The olivine potential crystallisation temperature, obtained
338 by iterative calculations using simultaneously melt compositions and Fo-contents (Putirka, 2008),
339 is in the narrow range of 1314-1202± 27 °C, with, coherently, the highest values recorded for the
340 olivine of the HP121.

341 The barometric conditions under which the Harrow Peak xenoliths were finally equilibrated are
342 firstly deduced by the presence of spinel as the sole aluminium phase (0.9 < P < 2.5 GPa, Wood,
343 1974). However, an attempt to evaluate the region where the crystallised phases olivine, olivine+
344 (orthopyroxene) and orthopyroxene + (olivine) separated from the hypothetical (calculated melt/s)
345 was carried out applying the geobarometers of Putirka (2008) and Nimis and Ulmer (1998), based
346 on orthopyroxene/melt and clinopyroxene/melt equilibrium, respectively. The obtained values

347 (Table 3) limit the stationing of Harrow Peaks cumulates in the P field of 1.3 (HP143) – 0.5 (HP163)
348 GPa. Considering that the estimated error (for the method) is ± 0.2 GPa, the obtained values are
349 coherent with those reported by Perinelli et al. (2011, 2017), which constrain the crust/mantle
350 boundary (Moho).

351

352 *5.2 Parental Melt modelling*

353

354 The predominant mineral assemblages (olivine+spinel) of the composite xenoliths, the occurrence
355 of orthopyroxenite and the calculated olivine (and orthopyroxene)/melt equilibrium, reveal that the
356 parental liquid had high MgO contents (Mg# 72) and was close to silica saturation. Such melts are
357 associated with an anomalous high temperature (excess T_{ex}) melting of the Phanerozoic mantle
358 (picrites) or the Archean cratons (komatiites). Picrites or komatiites either present cumulitic olivine
359 or are olivine-saturated even at high pressure, so that they can reach such high MgO and Mg# (Lee
360 et al., 2006; Class, 2008).

361 All the studied Harrow Peaks xenoliths (including orthopyroxenite HP163), have olivine as cumulus
362 phase, whose composition differs from most of the Antarctic ultramafic olivine cumulates,
363 commonly ascribed to the Cenozoic alkaline magmatism of the West Antarctic Ridge System
364 (Fig.5a; Gamble and Kyle, 1987; Gamble et al., 1988; Perinelli et al., 2017). Harrow Peaks' rocks
365 instead represent products crystallised at mantle depth, close to the Moho discontinuity, from the
366 precursor melt/s of the Ferrar Group tholeiitic magmatism (Kyle 1990, Fleming 1995), or from the
367 high Mg-tholeiites (low Ti –picrites), similar to those of Vestfalja (Heinonen and Luttinen, 2008;
368 2010).

369 The calculated primitive melt forming the dunite matrix of HP121 has Mg# (up to 72) higher than
370 any documented melt (Mg# 59-68), directly or indirectly (i.e. mantle metasomatic melt) related to
371 the Cenozoic magmatic system, which is responsible for xenoliths rising to the surface (Coltorti,
372 et. al. 2004; Perinelli et al. 2008; Nardini et al., 2009). In this respect, the HP121 dunite represents

373 the primitive crystallised product of the Kirkpatrick basalt (Ferrar Group) precursor (low Ti
374 picrites), whereas dunite portions of HP124 and HP143 composite xenoliths and harzburgite
375 /lherzolite samples are fractionated and accumulated by melt/s close in composition to the exposed
376 Ferrar tholeiitic magmas (i.e. Kirkpatrick basalt; dolerites; Kyle 1980; Fleming 1995) in a co-
377 genetic liquid line of descent (Table 4). The scarceness of olivine and the low Mg# of
378 orthopyroxene in HP163 orthopyroxenite, point towards a cumulate of the same evolved melt
379 (Table 4).

380 In such a frame, it is not easy to understand the role of clinopyroxene (and amphibole). The few
381 clinopyroxene grains of HP121 and those concentrated in amphibole-bearing veins in HP124 and
382 HP143, as well as the clinopyroxene occurring in harzburgite and lherzolites, generally show high
383 Mg# (90-91) and Fe/Mg clinopyroxene/olivine distribution coefficients ($^{Fe/Mg}Kd$ 0.59-0.97) out of
384 the range of equilibrium values ($^{Fe/Mg}Kd$ clinopyroxene/olivine = 0.83-0.86, Putirka 2008). Applying
385 the $^{Fe/Mg}Kd$ clinopyroxene/melt formulation of Putirka et al. (2008) and Mollo et al. (2017) to the
386 previous calculated Harrow Peaks parental melts, the ideal liquidus clinopyroxene has a $^{Fe/Mg}Kd$
387 clinopyroxene/melt of 0.24-0.27. The Harrow Peaks clinopyroxene instead, is characterised by
388 $^{Fe/Mg}Kd$ values (0.18-0.32) that cannot be equilibrated with most of the calculated melts (Table 3).
389 This is clear for the HP121 and HP124 composite xenoliths ($^{Fe/Mg}Kd$ clinopyroxene/melt = 0.18-
390 0.15), whereas the HP143 composite xenolith, where the clinopyroxene almost entirely forms the
391 monomineralic large vein (Figs. 2, 3), is the only sample that records a potential equilibrium of
392 clinopyroxene with its matrix olivine, and with a calculated melt ($^{Fe/Mg}Kd$ clinopyroxene/melt =
393 0.27) having Mg#=68 (Table 3). HP143 clinopyroxene is characterised by notably lower REE
394 contents with a slight LREE-enrichment and flat M-REE patterns ($La_N/Ce_N=1.25$). Moreover, it does
395 not present the high Eu negative anomaly, which instead characterises most of the Harrow Peaks
396 clinopyroxenes. Overall, the HP143 REE profile is coherent with the equilibrium with a tholeiite-
397 like forming melt and is clearly distinct from the alkaline-type clinopyroxene of HP121 and HP124
398 dunites (Fig.9b).

399 Furthermore, we use the REE of the HP143 clinopyroxene to evaluate the HP143 parental melt REE
400 contents. The ^{REE}Kd clinopyroxene/melt dataset used for the calculations is reported in [Table S11](#)
401 [of Supplementary Material](#). The resulting profile mimics those of the Ferrar tholeiites and,
402 coherently with the major element results, it suggests a potential equilibrium between the HP143
403 clinopyroxene and a parental tholeiitic melt ([Fig. 11](#)).

404 To sum up, HP121 and HP163 are cumulates from hypothetical parental melts that represent two
405 extremes: the most primitive and the most evolved terms of a hypothetical liquid line of descent,
406 respectively, suggesting that the parental melt of the Kirkpatrick basalt was a (low alkaline) picrite,
407 which stationed in the shallow mantle. The melt system, modified in time and space, produced the
408 rest of the Harrow Peaks ultramafic rocks, including the orthopyroxenite, at a shallower level. For
409 all, but HP143, clinopyroxene and amphibole were formed by a different magmatic or metasomatic
410 episode/s associated with more evolved melts that reasonably also crystallised plagioclase; this
411 phase, even if not present in these cumulus fragments, may have imprinted the scattered Eu negative
412 anomaly ([Fig.9](#)).

413

414 *5.3 Geodynamic implications*

415

416 The study of the Harrow Peaks xenoliths is clearly relevant as they represent the initial stages of the
417 Ferrar magmatism. The importance of the Ferrar magmatism in West Antarctica is widely attested
418 (Kyle 1980; Fleming 1992; Elliot 1999); the large volume of magmas generated by this event
419 permeated part of the lithospheric domains beneath northern Victoria Land, refertilising the residual
420 peridotitic upper mantle (Pelorosso et al., 2016).

421 The Ferrar large igneous system includes various magmatic bodies; among them, the doleritic
422 Basement Sill from the McMurdo Dry Valleys represents the plumbing system of the Ferrar flood
423 basalt eruptions (Bedard et al., 2007). The rock forming the sill displays a cumulate texture similar
424 to that observed in layered intrusions and is characterised by an “orthopyroxene enriched base”

425 (Bedard et al., 2007); in that view, the unique investigated orthopyroxenite HP163 could be
426 representative of the same basement.

427 Orthopyroxenites are quite rare among the Antarctica xenoliths suites. To the best of our knowledge
428 the only other one documented is associated with clinopyroxenites and peridotite xenoliths hosted
429 in basalts of the Mount Morning eruptive centre (Martin et al., 2015).

430 In northern Victoria Land, orthopyroxenites were also found as part of the layered sequence from
431 Niagara Icefall (northern Victoria Land, Tribuzio et al., 2008); they were interpreted as cumulates
432 of a boninitic melt related to the development of a back arc basin in an active continental margin.

433 Similarities between the Harrow Peaks ultramafic cumulates and the Niagara Icefall layered rocks
434 include: i) the occurrence of high Mg# dunites (Mg# values = 91-94) coexisting with
435 orthopyroxenites (Mg# values=84-88); ii) the matching of REE abundances and profiles of
436 clinopyroxenes from the Niagara Icefall gabbro with that of the HP143 clinopyroxene (Fig.

437 11). Therefore, considering the mentioned examples from different geodynamic settings, we cannot
438 exclude *a priori* that the Harrow Peaks cumulates represent fragments of cumulus “layers”
439 crystallised and separated from tholeiitic or boninitic melts. However, Harrow Peaks cumulates are
440 hosted in lavas sampled in the Wilson Terrane, where the Ferrar outcrops are represented by the
441 Ferrar Dolerite Dufek Intrusions and the Kirkpatrick Basalts (Hergt et al., 1989; Fleming et al.,
442 1995; Elliot et al., 1999). Thus the hypothesis that the investigated suite of samples could represent
443 cumulus rocks from the most primitive Ferrar melt seems the most plausible.

444 Tracing the origin of Harrow Peaks cumulates back to the Jurassic, it is reasonable that they
445 remained located within the shallow lithospheric mantle ($P \sim 1$ GPa) up to the Cenozoic (~ 50 Ma),
446 when the alkaline magmatism associated with the West Antarctic Rift brought them to the surface.

447 The Cenozoic (late stage) alkaline metasomatism pervaded large portions of mantle domains and
448 cumulates suites in Victoria Land (Coltorti et al., 2004; Perinelli et al., 2006, 2008, 2011; Martin et
449 al., 2015; Pelorosso et al., 2016, 2017), and affected the Harrow Peaks lithospheric portion, as

450 testified by the presence of alkaline-like amphiboles (phlogopite) and the LREE-enrichment in
451 clinopyroxene (HP121 and HP124).

452

453 **Conclusions**

454

- 455 1. The textural and geochemical characteristics of Harrow Peaks xenoliths evidence that these
456 ultramafic rocks represent deep crustal/mantle cumulates crystallised from a primitive
457 picritic melt with a potential temperature ~ 1300 °C;
- 458 2. The presence of orthopyroxenite in the system demonstrates that the inferred melt was
459 saturated in SiO₂;
- 460 3. The mineral/melt equilibrium allowed to identify the parental melt as corresponding to the
461 first stages of the Ferrar magmatism;
- 462 4. During Cenozoic, the lithospheric mantle beneath Victoria Land was variably affected by
463 an alkaline metasomatism related to the magmatic system, which was responsible for the
464 xenoliths sampling. This process also affected the Harrow Peaks district as testified by the
465 occurrence of hydrous phases in complete geochemical disequilibrium with the anhydrous
466 cumulus matrix.

467

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479

480

481

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698

699 Figure Captions

700

701 **Figure 1.** Sketch map of northern Victoria Land and location of Harrow Peaks outcrop (red star).
702 Greene Point (Melchiorre et al., 2011; Pelorosso et al., 2016), Baker Rocks (Coltorti et al., 2004;
703 Perinelli et al., 2006, 2008) and Handler Ridge (Pelorosso et al., 2017) are also reported (green
704 stars). The map of northern Victoria Land is modified from Estrada et al. (2016), the Antarctica
705 view is from Google Earth.

706

707 **Figure 2.** Harrow Peaks xenoliths in the olivine-orthopyroxene-clinopyroxene IUGS classification
708 diagram. (*) Dunitic matrix of composite xenoliths. The residual path (annotated by melting %)

709 from PM (Primordial Mantle, McDonough and Sun, 1995) to clinopyroxene consumption is also
710 calculated (dashed line) using theoretical and experimental results of Jensing and Holland (2015)
711 and Herzberg (2004).

712

713 **Figure 3.** Thin sections of Harrow Peaks mantle xenoliths, which are mainly characterised by
714 equigranular textural type. HP143, HP124, HP121 composite xenoliths (a, b, c) consist of large
715 clinopyroxene and amphibole veins cross-cutting a dunitic matrix. a) HP143 with evident spinel
716 trails. Important to note that the real modal content of spinel do not correspond to the black areas as
717 in the thin section. b) HP124 is characterised by dunitic matrix and the thinnest vein containing also
718 phlogopite. c) HP121 with the largest monomineralic vein that has been partly removed for
719 crystallochemical investigations (Gentili et al., 2015). Harzburgite HP144 (d) and high
720 clinopyroxene lherzolite (e). High orthopyroxene lherzolites (f, g) and orthopyroxenite (h).

721

722 **Figure 4.** Photomicrographs of representative microstructures from Harrow Peaks xenoliths.

723 Polygonal large olivine with boundaries-converging at 120° (a, b), detail of the large clinopyroxene
724 vein with embryonic amphibole (c). Tabular olivine zone and amphibole vein in sample HP121 (d).
725 Lherzolite HP166 characterised by equigranular olivine, orthopyroxene and small clinopyroxene
726 (e). Large orthopyroxene in the orthopyroxenite HP163 (f)

727

728 **Figure 5.** Chemical composition of (a) olivine (NiO vs. Fo), (b) orthopyroxene (Al_2O_3 vs. Mg#), (c)
729 spinel (Mg# vs. Cr#); (*) Dunitic matrix of composite xenoliths. Compositional fields defined by
730 Victoria Land peridotites (Coltorti et al., 2004; Martin et al., 2015; Pelorosso et al., 2016, 2017), Mt
731 Overlord cumulates (Perinelli et al., 2017) and Basement sill leucotroctolites (Bedard et al., 2007)
732 are reported for comparison.

733

734 **Figure 6.** Plot of spinel Cr# vs Fo content of coexisting olivine of Harrow Peaks ultramafic xenoliths.
735 (*) Dunitic matrix of composite xenoliths. The olivine–spinel mantle array is from Arai (1994).
736 Melting trend (annotated by melting %) is from Park et al. (2017); peridotite fields are modified
737 from Park et al. (2017).

738

739 **Figure 7.** Harrow Peaks clinopyroxene (a, b) and amphibole (c, d) compositions in term of Mg# vs.
740 TiO₂ and vs. Al₂O₃. Compositional fields reported for comparison in (a, b) are as in **figure 5**.

741

742 **Figure 8.** Chondrite-normalized trace element (REE+Sr+Zr+Hf+Ti+Y) distribution for Harrow
743 Peaks orthopyroxene (represented as sample average). Chondrite values from McDonough and Sun
744 (2005)

745 **Figure 9.** Chondrite-normalized trace-element patterns a) and Rare Earth Elements (REE) patterns
746 b) for Harrow Peaks clinopyroxene represented ad sample average, except for HP163
747 (orthopyroxenite). Due to the large trace element variability, all the analyses of the few HP163
748 clinopyroxenes are reported. Chondrite values from McDonough and Sun (2005)

749

750 **Figure 10.** Chondrite-normalized trace-element patterns (a) and REE patterns (b) of amphibole. For
751 comparison, the compositional field defined by the amphibole from northern Victoria Land
752 peridotites (Coltorti et al., 2004) is also reported. Chondrite values from McDonough and Sun
753 (2005)

754

755 **Figure 11.** Trace element composition modelling of melt 2, as calculated (major elements) in **Table**
756 **3**. The reconstructed melt 2, assumed in equilibrium with HP143 clinopyroxene, is obtained using
757 REE $K_d^{cpx/melt}$ data from GERM. The resulting profile converges on a tholeiitic melt that resembles
758 the Ferrar basalt (Kyle 1980; Fleming et al., 1995). In turn, HP143 clinopyroxene profile mimics

759 those occurring in the ultramafic sequence (cumulates) from Niagara Ice Falls (Tribuzio et al.,
760 2008). Chondrite-normalized values from McDonough and Sun (2005).