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1 **Flow dynamics in mid-Jurassic dikes and sills of the Ferrar large igneous province and**
2 **implications for long-distance magma transport**

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25 **Abstract**

26 Magma flow paths in sill-fed dikes of south Victoria Land, Ferrar large
27 igneous province (LIP), contrast with those predicted by classic models of dike transport in
28 LIPs and magmatic rift settings. We examine anisotropy of magnetic susceptibility (AMS)
29 flow paths in dike networks at Terra Cotta Mountain and Mt Gran, which intruded at
30 paleodepths of ~2.5 and ~1.5 km. These intrusions (up to 30 m thick) exhibit irregular,
31 interconnected dike-sill geometries and adjoin larger sills (~200-300 m thick) at different
32 stratigraphic levels. Both shallowly dipping and sub-vertical magma flow components are
33 interpreted from AMS measurements across individual intrusions, and often match
34 macroscopic flow indicators and variations in dike attitudes. Flow paths suggest that intrusive
35 patterns and magma flow directions depended on varying stress concentrations and rotations
36 during dike and sill propagation, whereas a regional extensional tectonic control was
37 negligible or absent. Unlike giant dike swarms in LIPs elsewhere (e.g., 1270 Ma MacKenzie
38 LIP), dikes of the Ferrar LIP show no regionally consistent vertical or lateral flow patterns,
39 suggesting these intrusion were not responsible for long-distance transport in the province.
40 In the absence of regionally significant, colinear dike swarms, or observed intrusions at
41 crustal depths ≥ 4 km, we suggest that long distance magma transport occurred in sills within
42 Beacon Supergroup sedimentary rocks. This interpretation is consistent with existing
43 geochemical data and thermal constraints, which support lateral magma flow for ~3,500 km
44 across the Gondwana supercontinent before freezing.

45

46 **Keywords**

47 Antarctica; anisotropy of magnetic susceptibility; thermo-mechanical model; sill-fed dikes;
48 Terra Cotta Mountain; Mount Gran

49

50 **1. Introduction**

51 Investigating how magma is transported and accommodated in the crust can
52 yield key insights into the processes governing the growth and breakup of continental
53 lithosphere (Buck, 2004; Ebinger et al., 2013), and the dynamics of magmatic systems that
54 feed volcanic eruptions (Tibaldi, 2015). In large igneous provinces (LIPs), the intrusive
55 components controlling both the lateral and vertical migration of magma transport are often
56 depicted as colinear swarms of giant dikes (Ernst et al., 1995; Ernst et al., 2001). The primary
57 direction of magma flow documented for these dike systems changes from vertical near the
58 plume head (300-500 km from plume center) to lateral away from the source. Examples
59 include the 1270 Ma MacKenzie and ~180 Ma Okavango dike swarms (Ernst and Baragar,
60 1992; Aubourg et al., 2008). However, the shallow plumbing systems (<10 km depth) of a
61 few LIPs, such as the 250 Ma Siberian LIP, form interconnected sill networks capable of
62 feeding voluminous outpourings of lavas (Naldrett et al., 1995; Cartwright and Hansen, 2006;
63 Muirhead et al., 2014). The geometries of dikes within these sill-dominated provinces differ
64 from classic depictions of LIP dike systems. These intrusions, termed by Muirhead et al.
65 (2014) as sill-fed dikes (but also referred to previously as inclined “sheets”; Airoidi et al.,
66 2011), exhibit short lengths (<5 km), variable dips (10-90°), form at sill peripheries, and link
67 sills at different stratigraphic levels (Johnson and Pollard, 1973; Czamanske et al., 1995;
68 Muirhead et al., 2012).

69 Although sill-fed dikes form a key component of the shallow plumbing
70 systems of sill-dominated LIPs, magma flow dynamics within these intrusions remain largely
71 unknown. Many studies focus on magma transport through the outer sheets and internal sills
72 of saucer-shaped intrusions (Ferré et al., 2002; Thomson and Hutton, 2004; Hansen and
73 Cartwright, 2006a; Maes et al., 2008; Polteau et al., 2008b; Galland et al., 2009). In the
74 Karoo LIP arrangement of intrusive segment ‘lobes’ and flow kinematics from anisotropy of

75 magnetic susceptibility (AMS) suggest an up-dip flow component in the outer inclined sheets
76 that connect to sill peripheries (Polteau et al., 2008a; Schofield et al., 2010; Galerne et al.,
77 2011). Airoidi et al. (2012), however, revealed complex lateral and vertical magma flow
78 patterns in shallowly dipping, sill-fed dikes in the Allan Hills region of Ferrar LIP,
79 Antarctica. These data were interpreted to record intermittent phases of 'passive' magma
80 injection into fracture networks forming in response to the forceful injection of underlying
81 sills. However, it is currently unknown whether this model of dike growth is regionally
82 consistent throughout the Ferrar LIP. The role that dikes played in controlling the regional
83 distribution of Ferrar magmas is therefore poorly constrained.

84 We analyze magma transport dynamics at various depths in the magmatic
85 plumbing system of the Ferrar LIP. Emplaced ~10 million years prior to the breakup of East
86 from West Gondwana, this widespread (c.a. 4,100 km long) magmatic province forms part of
87 the ~183 Ma Karoo-Ferrar LIP (Encarnaciòn et al., 1996), and provides important insights
88 into the tectono-magmatic conditions across Antarctica during this continental breakup event.
89 AMS is applied to Ferrar intrusions at Terra Cotta Mountain and at Mt Gran, south Victoria
90 Land (Fig. 1), to constrain a model for magma transport dynamics throughout the province.
91 These analyses are used to infer (1) controls on intrusion propagation at different levels of the
92 plumbing system, (2) characteristic flow modes within dikes and sills, and (3) the intrusive
93 structures responsible for broad-scale magma transport throughout the Ferrar LIP.

94

95 **2. The Ferrar large igneous province**

96 Ferrar LIP rocks are exposed for 3,500 km along the Transantarctic Mountains
97 of East Antarctica. These intrusive and extrusive rocks, with a total estimated volume around
98 300,000 km³ (Ross et al., 2005), represent the most laterally extensive LIP system on Earth.
99 Studies addressing the broad-scale emplacement of the province support a lateral transport

100 model, with Ferrar magmas travelling >3,000 km from the Weddell Sea across the east
101 Antarctic margin into south-eastern Australasia (Elliot et al., 1999; Elliot and Fleming, 2000;
102 Leat, 2008). Emplacement of the province occurred over 349 ± 49 kyr (Burgess et al., 2015),
103 during (or just prior to) the earliest stages of Gondwanaland breakup, and coincided with the
104 emplacement of the earliest Karoo lavas and sills (U-Pb ages on zircon and baddeleyite
105 between 183.6 ± 1.0 and 182.8 Ma, cf. Encarnaciòn et al., 1996 and Burgess et al., 2015, and
106 references therein). The cross-continental distribution of Ferrar LIP rocks has led authors to
107 suggest that Ferrar magmas intruded and erupted in a continental rift system driven by
108 regional extension, or trans-tension, in a back-arc setting (Wilson, 1993; Storey, 1995; Elliot,
109 2013). However, intrusion and fracture systems trends consistent with regional extension in
110 the Jurassic are lacking (Muirhead et al., 2012). Instead, regional dike patterns are consistent
111 with magma emplacement under a far-field neutral stress regime (Muirhead et al., 2014).

112 Ferrar intrusions are observed dissecting the flat-lying, ~2.5 km-thick Beacon
113 Supergroup sedimentary sequence and the upper ~0.5 km of underlying basement granitoid,
114 amphibolite and metasedimentary rocks (Elliot and Fleming, 2008). Sills are significantly
115 more voluminous than dikes (Muirhead et al., 2014). In south Victoria Land, sills reach
116 ~5,000 km² in area and up to 450 m in thickness (Gunn and Warren, 1962). Some sills are
117 observed ascending the stratigraphy in a ‘step-wise’ fashion (Elliot and Fleming, 2004;
118 Airoldi et al., 2011) and, within the upper Permian and lower Triassic members of the Beacon
119 sequence, sills become progressively thinner (0-100 m) in places and laterally less continuous
120 (Elliot and Fleming, 2004, 2008).

121 Swarms of shallow to moderately dipping dike intrusions are reported from
122 various localities in the central Transantarctic Mountains (e.g. Hornig, 1993; Leat, 2008 and
123 references therein) and south Victoria Land (Skinner and Ricker, 1968; Wilson, 1993;
124 Morrison and Reay, 1995; Muirhead et al., 2014). Regional field and remote sensing studies

125 reveal that these intrusions connect sills at different stratigraphic levels, and are inferred to
126 assist in the vertical transport of magma in the upper 4 km of the plumbing system to the
127 surface (Muirhead et al., 2012; Muirhead et al., 2014). Magma flow dynamics within these
128 sill-fed dikes are, however, poorly constrained.

129

130 **3. Field sites**

131 **3.1. Terra Cotta Mountain**

132

133 Dike intrusions in the Terra Cotta Mountain area are exposed along NE- and
134 SW-facing cliffs (Fig. 2 in Morrison and Reay, 1995). These cliffs reveal a swarm of
135 moderately dipping (mean dip 51° : Muirhead et al., 2012) intrusions dissecting Beacon
136 Supergroup rocks and connecting to the lower contact of a sill capping the mountain
137 (Muirhead et al., 2012). Mt Kuipers lies immediately east of Terra Cotta Mountain, where a
138 ~200 m-thick sill and two dike intrusions can be seen on the western flanks. Basement
139 granitoids exposed at the northern foothill of the nunatak underlie a sequence of quartz-rich
140 sandstones, siltstones and minor mudstones and conglomerates. These sedimentary sequences
141 belong to units from the Windy Gully Sandstone to the Beacon Heights Orthoquartzite of the
142 ~1.5 km thick Taylor Group rocks (Harrington, 1958; Gunn and Warren, 1962; Ross et al.,
143 2008). A ~1.0 km-thick sequence of relatively undeformed, flat-lying, sedimentary rocks of
144 the Victoria Group lie unconformably on Taylor Group rocks. At Terra Cotta Mountain,
145 Ferrar intrusions are observed dissecting rocks of the Windy Gully Sandstone up to the Arena
146 Sandstone, which suggests emplacement at paleodepths of ~1.5 – 2.5 km (cf. Fig. 1)
147 (Morrison, 1989; Muirhead et al., 2012).

148 Dike attitudes at Terra Cotta Mountain are irregular, with orientations varying
149 along strike to form zig-zag patterns. Intrusion dips also vary up-section, where some

150 intrusions change into sills that connect offset dike segments, resulting in a transgressive dike
151 geometry (Airoldi et al., 2011). Some intrusions bifurcate locally into smaller dikes. For
152 example, dikes up to 10 m thick can be seen connecting to smaller (2-6 m thick), ‘offshoot’
153 dikes, some of which in turn feed into thinner (<2 m thick) ones (Fig. 2a). Offshoot dikes are
154 characterized by irregular geometries, with different segments exhibiting left- and right-steps,
155 both along-strike and up-section, and curved tips (Fig. 2b).

156 The relative timing of diking events is ambiguous on the SE slopes of Terra
157 Cotta Mountain. Although dikes do cross one another in places, no chilled margins are
158 observed along intrusive contacts that would allow interpretation of the relative timing of
159 intrusion events. However, within individual dikes, chilled contacts are observed trending
160 sub-parallel to the plane of the intrusion (Fig. 3a-c). Chilled zones within intrusions exhibit
161 either sharp or diffuse contacts. Thin zones comprising a mixture of un-melted, baked and
162 thermo-mechanically deformed host rock material, host rock fragments, calcite veins, and
163 chilled dolerite fragments are observed within some dikes, and trend sub-parallel to the
164 nearest intrusion margin. Similar contact relationships also appear along dike selvages (Fig.
165 3b-c).

166 ‘Baked’ zones are common in the host rock alongside intrusion margins and
167 are typically 1-2 cm wide. Evidence of thermo-mechanical deformation affecting both
168 country rock and dolerite is observed at several locations. For example, where sharp selvages
169 are present, country rock at the margins of intrusions exhibits deformed surfaces (Fig. 4).
170 Striations and, locally, discontinuous veins 1-10 cm wide, are also observed on country rock
171 walls, along dike margins. These locally exhibit mineral striations with near-vertical
172 lineations (Fig. 4a). No fault planes dissecting the dikes were observed. Variably shaped
173 cusps-and-grooves and drag folds are preserved along country rock margins (Fig. 4b and c).
174 There are peperite zones in sedimentary rocks near some intrusion margins. These zones

175 include chaotic arrays of angular fragments and/or rounded pods of chilled dolerite,
176 intermingled with lithified medium-to-fine sand material (Fig. 3d).

177

178 **3.1.1. Target intrusions**

179

180 The principal intrusions investigated in this study are shown in Fig. 5. At the
181 base of Terra Cotta Mountain is a >20 m-thick intrusion (d#3). The lower contact of the
182 intrusion is visible, striking 340° and dipping east at 75°. However, the upper contact cannot
183 be seen anywhere in the field area, and the intrusion does not appear to have significant
184 lateral continuity. Near the summit of the mountain, multiple intrusions are observed
185 branching out from d#3 upwards, into the thick (>100 m) sill that caps the mountain (s#4).
186 On the northern and eastern slopes of the mountain variably dipping (2-76°) intrusions
187 exhibit alternating dike-sill geometries (e.g., t#1a-b, and t#2). On the western slopes of Mt
188 Kuipers is a ~200 m-thick sill, a ~20 m-thick, ~090° striking dike (d#5), and a 10 m-thick,
189 ~160° trending dike (d#6). The latter dike (d#6) tapers down and ends to the north-west
190 before reaching d#5, and it was not observed on Terra Cotta Mountain's southeastern cliff.
191 Further south-east on Mt Kuipers, d#6 truncates the sill.

192

193 **3.2. Mount Gran**

194

195 Mt Gran is located ~30 km south-east of Allan-Coombs Hills. Here, a steep,
196 ~750 m high, southeast-facing cliff exposes a complex network of intrusions (White et al.,
197 2009) (Fig. 5e). These Ferrar dikes and sills intrude upper Taylor Group and lower Victoria
198 Group rocks. Based on thickness estimates for the Taylor Group (Harrington, 1958; Gunn
199 and Warren, 1962), we infer that Ferrar intrusions at Mt Gran were emplaced at a paleodepth

200 of 1-1.5 km (White et al., 2009; Fig. 1). The most prominent intrusion at Mt Gran is a ~30 m-
201 thick, sub-vertical dike (d#7), which truncates a >40 m-thick sill (s#8). A network of
202 interconnected sills and transgressive dikes, all less than 20 m thick (s#9), are exposed on the
203 northeastern side of the cliff. These shallowly dipping (typically <30°) intrusions transgress
204 the stratigraphy up-section to the southwest, before merging into d#7 (Fig. 5e). Other inclined
205 intrusions also extend outward from the western margin of d#7. Two dikes (d#10 and d#11)
206 are exposed in a valley a few hundred metres northwest of the cliff, and exhibit 086° and
207 056° strikes.

208 **4. Methods: anisotropy of magnetic susceptibility**

209 Paleo-magma flow directions can be determined by analyzing the preferred
210 alignments and orientations of Fe-bearing minerals using a method known as anisotropy of
211 susceptibility (AMS) (Tarling and Hrouda, 1993). This non-destructive approach has become
212 common in the last few decades due to its time- and cost-effectiveness, and is commonly
213 used to constrain interpretations of dike and sill emplacement dynamics within volcanic
214 plumbing systems (see Baer and Reches, 1987; Walker et al., 1999; Ferré et al., 2002; Liss et
215 al., 2002; Poland et al., 2004; Delcamp et al., 2014 for a few examples) from the 'primary',
216 dominant magmatic flow indicated by AMS fabrics (e.g., Hrouda, 1982; Dragoni et al., 1997;
217 Stevenson et al., 2007; Magee et al., 2012b). This method has also been used to understand
218 the broad scale emplacement of LIPs, such as the MacKenzie LIP (Ernst and Baragar, 1992),
219 the volcanic margin of east Greenland (Callot and Geoffroy, 2004), the Karoo LIP (Aubourg
220 et al., 2008; Polteau et al., 2008b), the British and Irish Paleogene igneous province (Magee
221 et al., 2012a), and the Siberian LIP (Callot et al., 2004).

222

223 **4.1. Sampling and analyses**

224

225 Data presented in this study come from 97 oriented block samples collected
226 close to the walls of Ferrar dikes and sills at Terra Cotta Mountain (81 samples) and Mt Gran
227 (16 samples). The orientations of dolerite block samples were determined in the field with
228 both solar and magnetic compasses and a clinometer. Samples sizes were approximately
229 10×10×15 cm, to provide sufficient material to perform petrographic and magnetic analyses.
230 Depending on intrusion size and outcrop accessibility, sampling was also performed across
231 intrusion interiors in order to detect any significant compositional and/or textural variations.
232 Where possible, both walls of intrusions were sampled to investigate imbricated magnetic
233 foliations (Knight and Walker, 1988).

234 Samples were prepared for petrographic and magnetic analysis at the
235 University of Otago Geology Department and Otago Paleomagnetic Research Facility
236 (OPRF), New Zealand. At least one sample per intrusion was petrographically analyzed.
237 Between 3 and 15 core specimens (diameter = 25 mm, length = 22 mm) were taken from
238 every block sample for magnetic analyses. Magnetic susceptibility and AMS measurements
239 on over 500 core specimens from Terra Cotta Mountain were analyzed at the inter-university
240 research centre Alpine Laboratory of Paleomagnetism (ALP - Peveragno, Italy), using an
241 AGICO KLY-3 Kappabridge. Susceptibility versus temperature analyses were run for one
242 selected specimen per intrusion using a CS-3 furnace at OPRF. Isothermal remanent
243 magnetization (IRM) acquisition, thermal demagnetization and backfield curves were
244 obtained either using a JR-6 spinner magnetometer (Lowrie, 1990) at ALP, or with a
245 Princeton Instruments Vibrating Sample Magnetometer at OPRF. Magnetic carriers in
246 igneous rocks from Terra Cotta Mountain were determined through the analysis of rock
247 magnetic properties. This included magnetic susceptibility, defined by the ratio between the
248 induced magnetization of the material and the inducing magnetic field, IRM, remanence

249 coercivity (B_{CR}), temperature-dependant susceptibility (K_T vs T), blocking (T_B), and Curie
250 temperatures (T_C) at which magnetic minerals lost their magnetic properties.

251 AMS of Mt Gran intrusions was measured using a KLY-4s Kappabridge
252 apparatus at the University of Southern California. Natural remanent magnetization (NRM)
253 and stepwise alternating field (AF) demagnetization measurements were performed with a 2-
254 G cryogenic magnetometer with inline AF demagnetizer (up to 200 mT). T_B spectra and an
255 estimate of Curie temperatures were determined on a ASC thermal demagnetizer. The
256 magnetic mineralogy of Mt Gran samples was determined by combining information such as
257 AMS and NRM, remanence coercivity and Curie and blocking temperatures.

258

259 **4.2. Interpretation of magnetic fabrics**

260

261 Flow textures in intrusive rocks are the result of the hydrodynamic alignment
262 of elongate crystals during magma flow. Fe-Ti oxides such as (titano-)magnetite mimic this
263 alignment because they form within and/or along the edges of earlier crystallized, non-
264 ferromagnetic crystals (e.g. feldspar) after magma flow has ceased. As a consequence, flow
265 directions are commonly inferred from the arrangement and orientations of all magnetic
266 components within the rock fabric and overall intrusive body.

267 The anisotropy of magnetic susceptibility, or AMS, is modelled as an ellipsoid
268 with mutually orthogonal axes $k_1 \geq k_2 \geq k_3$ (respectively, maximum, intermediate and
269 minimum susceptibility axes). These axes can be graphically plotted as lineations on equal
270 area stereographic projections (Fig. 6). The anisotropy parameters defined for any magnetic
271 fabric ellipsoid are the mean magnetic susceptibility (K_m) and anisotropy degree (P or P_j ,
272 corrected anisotropy degree) defining the absolute anisotropy of a rock specimen, and
273 magnetic lineation (L), foliation (F), and shape parameter (T) (see Tarling and Hrouda, 1993,

274 table 1.1, p. 18, for their mathematical expression). Together, L, F and T define the geometry
275 of the AMS ellipsoid. Prolate fabric ellipsoids are elongate ($L > F$) and characterized by –
276 $1 \leq T < 0$, whereas oblate ellipsoids are flattened ($F > L$) and characterized by $0 \leq T \leq 1$. In
277 the directional analysis of AMS fabrics, magnetic lineation and foliation correspond
278 respectively to the maximum susceptibility axis direction k_1 , and to the plane perpendicular to
279 k_3 and defined by k_1 and k_2 i.e. the magnetic foliation plane (FPL).

280 Susceptibility and its parameters also depend upon the magnetocrystalline
281 properties and/or distribution of each magnetic mineral species. Ferromagnetic multi-domain
282 (titano-)magnetite grains are the common magnetic carriers in mafic igneous rocks and
283 typically produce a prolate AMS fabric, whose magnetic lineation and foliation are aligned
284 with the plane of the intrusion (or imbricated up to 30° , cf. Dragoni et al., 1997) and indicate
285 the flow direction during magma emplacement (Tarling and Hrouda, 1993). For this type of
286 fabric, also termed a *normal fabric* (N, in Fig. 6), the minimum susceptibility axis is sub-
287 perpendicular to the intrusion plane (IPL).

288 Magnetic fabrics in intrusive rocks are, however, also known to exhibit
289 deviations from the normal fabric described above. Imbrication angles of 30° to 45° between
290 the FPL and intrusion plane, as well as the exchange of the intermediate and minimum axes
291 of the fabric ellipsoid, are commonly related to composite magnetic mineralogy of AMS
292 sources with different properties (e.g. Ferré, 2002; Aubourg et al., 2008). These are known as
293 *intermediate fabrics* ($I_1 - I_3$ in Fig. 6), and are classified after Airoidi et al. (2012, and
294 references therein) as 3 types:

- 295 • I_1 AMS fabrics are prolate, with the magnetic lineation lying within 45° from the
296 intrusion plane and k_2 and k_3 dispersed on a girdle.
- 297 • I_2 fabrics are either prolate or oblate, have both k_1 and k_3 aligned with the intrusion
298 plane and FPL orthogonal to it.

299 • I_3 is a 'nearly normal' planar fabric, with intrusion and magnetic foliation planes sub-
300 parallel to one another, and imbrication or intersection angle $>30^\circ$; the intermediate
301 susceptibility k_2 , rather than the magnetic lineation k_1 , lies closest to the intrusion
302 plane. For fabrics of this type, either k_2 , or the intersection between intrusion and
303 magnetic foliation planes, can be used as proxy of the magma flow direction (e.g.
304 Geoffroy et al., 2002).

305 The last fabric type presented in this study, termed *inverse* (R, in Fig. 6), is
306 related to the presence of single-domain magnetic grains within the rock (Rochette et al.,
307 1999; e.g. Airoidi et al., 2012). This fabric type is characterized by minimum susceptibility
308 axes aligned within the intrusion plane, and the magnetic foliation perpendicular to the
309 intrusion.

310

311 **4.3. Corroboration of AMS data**

312

313 AMS fabrics were also compared with macroscopic indicators of magma flow
314 observed in the field. In these instances, the shape, trend and plunge of preserved
315 macroscopic features both within dikes and along intrusion selvages were used to corroborate
316 AMS data. Some studies use the orientation of the long axis of a broken bridge or step
317 structure between dike and sill segments (Airoidi et al., 2012 and references therein) or
318 microscopic alignments of minerals and vesicles as direct indicators of magma flow (e.g.
319 Geshi, 2008; Soriano et al., 2008). Cusps-and-grooves and plumose structures along dike
320 selvages may give information on both the local magma flow lineation and sense of shear
321 along an intrusion (Varga et al., 1998; Correa-Gomes et al., 2001; Baer et al., 2006; Urbani et
322 al., 2015). Striations on dike walls could represent both magma flow, and shear between
323 magma and encasing rocks related to dike opening (e.g. Correa-Gomes et al., 2001; Baer et

324 al., 2006) and/or early-stage shear fracturing ahead of a propagating dike tip (Wilson et al.,
325 2016). In the current study, AMS results were compared with flow directions indicated by the
326 presence of cusps-and-grooves and drag folds observed along 5 intrusions out of 7 at Terra
327 Cotta Mountain (all but d#3 and t#1b). No flow indicators were recorded at Mt Gran.

328

329 **5. Source of Ferrar Dolerite magnetism**

330 The interpretation of magnetic fabric properties in rocks requires the
331 identification of magnetic carriers. The general petrographic characteristics observed in Terra
332 Cotta Mountain and Mt Gran dolerites are described below, followed by a description of the
333 results of magnetic mineralogy tests and their significance.

334

335 **5.1. Petrographic characteristics**

336

337 Ferrar dolerites from the two study locations are compositionally and
338 texturally similar. They exhibit a narrow range of crystal sizes (commonly 100 μm to
339 500 μm) and compositions. All samples contain a combination of orthopyroxenes,
340 clinopyroxenes and plagioclase, with variable amounts of opaques and secondary/alteration
341 minerals. Larger pyroxene crystals occasionally enclose tabular plagioclase. Rutile and
342 magnetite either included within or between grains are the main opaque phases observed at
343 Terra Cotta Mountain, whereas small amounts of magnetite and hematite with variable Fe-Ti
344 content were determined from reflected light microscopy in Mt Gran intrusions, near or
345 within pyroxene crystals.

346 At Terra Cotta Mountain, d#3 and d#5 are characterized by the above mineral
347 assemblage, with orthopyroxene enstatite and clinopyroxene augite and pigeonite crystals
348 around 50-60%, and commonly $\leq 40\%$ plagioclase crystals. Within t#1a-b and t#2,

349 orthopyroxene becomes less common, and clinopyroxene and plagioclase increase in
350 abundance up-section. S#4 and the intrusions at Mt Gran are petrographically similar. In
351 these intrusions, plagioclase is the most abundant mineral phase (40-60%), with 25-40%
352 clinopyroxene (Aug±Pig), and <20% orthopyroxene.

353 Terra Cotta Mountain dolerite textures are commonly microlitic porphyritic to
354 glomeroporphyritic, with no visible microscopic or macroscopic flow textures (Fig. 7a and
355 b). Glass is uncommon, with the exception of a few chilled margins. Iron oxides and
356 alteration products regularly replace the microlitic groundmass, and are especially common in
357 s#4 and d#6 (Fig. 7b).

358 At Mt Gran, variations in crystal size, shape, and texture occur as function of
359 proximity to chilled margins. Fine-grained textures with intersertal regions of glassy
360 mesostasis, pervasive opaques and alteration products are commonly observed near intrusion
361 margins (e.g., samples Mg 18-1a and 1-1b). Glomeroporphyritic and/or microlithic
362 porphyritic textures characterize the internal portions of dolerite intrusions (e.g., the center of
363 d#7), where opaques and alteration products also become less abundant (e.g., Mg 14-1a to
364 Mg 17-1a, see supplementary Table S1).

365

366 **5.2. Magnetic mineralogy properties**

367

368 Magnetic properties of Terra Cotta Mountain and Mt Gran intrusions are
369 rather uniform. K_m values range from 724 to 62183 μSI and 423 to 40800 μSI , respectively,
370 with ~90% of the data on the orders of 10^{-3} and 10^{-2} SI. The degree of anisotropy is normally
371 1-5%, with maximum values of 1.047. The least anisotropic samples were collected on d#5.

372 Thermomagnetic curves obtained from K_T vs T tests on Terra Cotta Mountain
373 dolerites present either a stable (two specimens) or, commonly, an irregular behavior, where

374 the curves display upward inflexions of the bulk susceptibility and decay around 400 °C (Fig.
375 8a). Significant final alteration of the specimens at high temperature is uncommon. For
376 example, there is no sharp variation in susceptibility at the end of the progressive thermal
377 demagnetization (PTD); instead, a gradual removal of the total rock magnetization occurs
378 between 550 and 600 °C. Plots of magnetic intensity upon PTD from Mt Gran samples also
379 show drops in the 550 - 600 °C thermal interval.

380 Similarly, steep IRM decay occurs as temperatures approach 400 °C during
381 PTD of different B_{CR} fractions (Fig. 8b). This IRM decay is not accompanied by irregular K_m
382 vs T paths (Fig. 8d) which, if present, would indicate mineralogical alteration. Soft
383 remanence coercivity ($B_{CR} < 500$ mT) magnetic components isolated with the Lowrie test are
384 normally over 58%, and the total contributions from the medium ($500 < B_{CR} < 1000$ mT) and
385 hard magnetic fractions ($B_{CR} \geq 1000$ mT) are below 37% and 7%, respectively (Fig. 8b).

386 Saturation of Ferrar specimens is reached with applied field values (B_S) of
387 300 mT, indicating a dominant low-coercivity magnetic phase. Additional irregular steps
388 observed in the IRM decay curves are likely due to demagnetization of soft and medium B_{CR}
389 fractions (B_{CR} ranging between 30 and 60-70 mT) during application of the back field (Fig.
390 8c).

391

392 **5.3. Interpretation of magnetic carriers**

393

394 Magnetic saturation, remanence coercivity and Curie temperature values
395 determined for Terra Cotta Mountain dolerites, with remanence coercivity overlap in IRM
396 plots, indicate the presence of both soft and medium remanence coercivity magnetite and/or
397 maghemite (cf. Borradaile and Jackson, 2004 and references therein). Magnetite is the
398 common magnetic carrier in basalts (see Tarling and Hrouda, 1993). However, selective

399 oxidation of magnetite can lead to formation of maghemite, particularly in hydrothermal
400 environments (see O'Reilly, 1983; de Boer and Dekkers, 1996). It is possible that magnetite
401 present in Ferrar dolerites altered to maghemite during, for example, a post-Ferrar
402 hydrothermal event in south Victoria Land (e.g. Craw et al., 1992; Ballance and Watters,
403 2002). In fact, the predominance of K_m values $>10^{-3}$ SI indicates contributions from both
404 ferromagnetic (e.g. (titano-) magnetite and maghemite) and paramagnetic (e.g. pyroxenes and
405 micas) minerals to the magnetic properties of the samples (Owens, 1974; Rochette, 1987;
406 Hrouda, 2002 and references therein).

407 Uniformity of magnetic properties, with blocking temperatures around 550 °C
408 and low coercivities, suggests magnetite with variable Ti-content is the dominant magnetic
409 carrier in Mt Gran rocks.

410 We infer a magnetic mineralogy derived from contributions by different
411 magnetic carriers (for instance, accessory magnetic minerals such as pyrrhotite and
412 titanohematite) from petrographic observation of a diffuse oxidation patina in a few samples,
413 occasionally associated with un-differentiated opaque minerals, and magnetic properties. As
414 discussed in Section 5.2, bulk susceptibility inflexion and decay, and steepening of IRM
415 curves around 400 °C during progressive thermal demagnetization of Terra Cotta Mountain
416 samples occur in the absence of any observable mineralogical alteration in the samples. These
417 variations in IRM are, however, consistent with breakdown of pyrrhotite around 300-400°,
418 and may represent reorganization and/or recrystallization of heated magnetic grains in both
419 the single and multi-domain state (Thompson and Oldfield, 1986; Hopkinson, 1989).

420

421 **6. Anisotropy of magnetic susceptibility of Ferrar intrusions**

422 The presence of magnetite and/or maghemite as main magnetic carrier(s) in
423 Ferrar samples is demonstrated by the magnetic properties, and validates the interpretation of

424 'normal' and 'intermediate' magnetic fabrics on the basis of the magnetic lineation direction
425 k_1 and magnetic foliation plane (see also Section 4.2).

426

427 **6.1. Magnetic fabric distribution**

428

429 Terra Cotta Mountain samples are characterized by both prolate (55%) and
430 oblate (45%) magnetic susceptibility ellipsoids. AMS fabric types include normal (23%), I-
431 type (59%) and inverse (4%), and 14% of samples exhibit (near-) isotropic magnetic fabrics
432 ($P < 1.005$, $F=L$, low values in tests of anisotropy). Sample-by-sample AMS parameters
433 defined for Terra Cotta Mountain dataset are presented in supplementary Table S2. Similarly
434 at Mt Gran, samples exhibit both prolate (56%) and oblate (44%) magnetic susceptibility
435 ellipsoids. Normal and intermediate AMS fabrics comprise ~20% and ~45% of the total data,
436 respectively. 40% of Mt Gran dataset (s#8, two d#11 sites and three of the eastern
437 sills/shallowly dipping sheets sites) is characterized by anomalous oblate reverse fabrics.
438 Samples producing either isotropic or inverse fabrics (30% of Terra Cotta Mountain samples
439 and ten sites from Mt Gran) were discarded from directional AMS analysis.

440

441 **6.2. AMS flow directions**

442

443 Samples from Terra Cotta Mountain and Mt Gran were grouped into 35 and
444 16 sub-sections, respectively (Tables 1 and 2). Each sub-section contains data analyzed from
445 1 to 4 sample sites on individual intrusions. Samples within each sub-section produced
446 consistent AMS fabrics and directions.

447 Maximum and/or intermediate susceptibility axes commonly lie within 20° of
448 the intrusion plane (Tables 1 and 2). k_1 (or k_2 , in I_3 fabrics) is a reliable flow proxy in 50% of

449 all sub-sections. In instances where the intersection between IPL and FPL was used (e.g., t#2
450 sites, and see d#5-6b in Fig. 6), the flow direction is $<30^\circ$ from the k_1 or k_2 axes. Except for
451 minor local misfits, AMS data are in good agreement with intrusion geometries (i.e., flow
452 directions sub-parallel to intrusion walls) and macro-scale kinematic indicators. 71% of AMS
453 flow directions trend within 35° of the macroscopic indicators. A similar fit was observed
454 between macroscopic and magnetic flow fabrics at Allan Hills (70% of AMS fabrics are
455 within 35° : Airoidi et al., 2012) and intrusive swarms elsewhere (Ardnamurchan, Scotland:
456 Magee et al., 2013), suggesting that magnetic lineations presented in this study correlate to
457 magma flow axes.

458

459 **6.2.1. General magma flow characteristics**

460

461 Flow components recorded along the margins of analyzed dike intrusions are
462 variable. Magnetic lineation plunges of dikes at Terra Cotta Mountain and Mt Gran range
463 from 7 to 79° (Fig. 9). Of the 35 dike sub-sections analyzed, 17% of magnetic flow directions
464 plunge $\leq 20^\circ$, 37% plunge $21-45^\circ$, and 46% plunge $>45^\circ$. Similarly, the trends of magnetic
465 lineations are variable, and almost any orientation is represented (Fig. 9). These multiple flow
466 directions are also reflected in the orientations of cusps-and-grooves along the walls of
467 intrusions (Figs. 4 and 10).

468

469 **6.2.2. Magma flow at Terra Cotta Mountain**

470

471 AMS directions in the Terra Cotta Mountain region follow the geometrical
472 variations of the dike intrusions. For example, within transgressive dikes (e.g., t#1), shallowly
473 plunging ($<25^\circ$) flow paths occur along shallow dipping segments (t#1a-1, t#1a-3, t#1a-4),

474 whereas steeper flow paths ($>25^\circ$) are observed only in the steeper dike segments (intrusion
475 dips $>50^\circ$) (t#1b). Dikes are characterized by variable magma flow paths (Fig. 9 and 11).
476 This is particularly evident in the thickest dikes (d#3 and d#5), where no specific lateral or
477 vertical flow-modes characterize dike selvages and/or the intrusion interiors. For example,
478 analysis of 12 sub-sections along d#5 reveal magma flow plunges ranging $7\text{-}79^\circ$ (Table 1 and
479 Fig. 11), with both shallowly plunging and sub-vertical flow lineations aligning with the
480 plane of the intrusion. AMS flow trends for d#3 range from $267\text{-}336^\circ$, with shallow-to-
481 moderate plunges ($13\text{-}37^\circ$) along the south-western margin of the intrusion (d#3-1 and d#3-
482 4). Steeper flow plunges ($46^\circ\text{-}68^\circ$) correspond to the innermost sampling sites (d#3-2a and
483 d#3-3, Table 1 and Fig. 11). Multiple flow directions can also be inferred for all individual
484 intrusions from cusps-and-grooves observed on the walls of dikes. Evidence of composite
485 flow-modes is, however, not always observed in thinner dikes (width <10 m, e.g., d#6),
486 although this may in part be the result of smaller AMS sample sets across some of these
487 intrusions.

488 AMS flow lineations from the Terra Cotta Mountain summit sill (s#4) exhibit
489 westward trends ($268\text{-}310^\circ$), with sub-horizontal plunges ($<10^\circ$). These magnetic lineations
490 are sub-parallel to lineations of the macroscopic flow indicators.

491

492 **6.2.3. Magma flow at Mt Gran**

493

494 The magnetic fabric at Mt Gran exhibits a general consistency with the overall
495 geometry of the sampled intrusions (i.e., flow sub-parallel to the dike walls). Magma flow in
496 dikes is generally sub-vertical, with 75% of sampled sub-sections exhibiting flow plunges
497 $>55^\circ$. AMS flow lineations constrained for the 30 m-thick central dike (d#7) define an overall
498 north-trending, sub-vertical ($63^\circ\text{-}78^\circ$ plunges) flow, with two shallow ($19\text{-}34^\circ$ plunges)

499 anomalous AMS directions at the center and eastern margin of the intrusion (Fig. 12). Due to
500 either isotropic or reverse magnetic fabric, no directional information could be constrained
501 for the large sill at the base of the cliff (s#8), the shallowly dipping sills and transgressive
502 dikes (s#9) on the northeast end of the cliff face, and one dike (d#10).

503

504 **6.3. Summary**

505 Terra Cotta Mountain and Mt Gran samples are characterized by both prolate
506 and oblate magnetic susceptibility ellipsoids, with normal AMS fabrics adding up to about
507 20%, and intermediate ones to ~45% of the total data respectively. Isotropic or inverse fabrics
508 were discarded from directional AMS analysis.

509 Magma Flow directions were inferred from AMS fabrics by applying a
510 geometric approach based on the orientation of the magnetic lineation and/or the magnetic
511 foliation plane relative to each intrusion's plane to the magma flow. Over 70% of the
512 magnetic flow indicators and macroscopic kinematic indicators trend within 35° of each
513 other, and are consistent with intrusion geometries. Multiple magma flow paths are common
514 along individual intrusions, with flow plunges as low as 7° and as steep as 79° (19°-78° at Mt
515 Gran) along the dikes, and flow trends of almost any orientation. Magma flow paths defined
516 for the Terra Cotta Mountain summit sill are consistently sub-horizontal, with westward
517 trends.

518

519 **7. Discussion**

520 Long-distance magma transport in LIPs is often depicted to occur through the
521 emplacement of giant dikes, 100s of km long and 10s of m thick (Ernst et al., 1995). These
522 dikes are shown to have transported magma >1000 km laterally away from an inferred plume
523 source (e.g., MacKenzie dike swarm, Ernst and Baragar, 1992). The development of sill-

524 dominated magmatic systems within LIPs, however, has been increasingly recognized over
525 the past decade (e.g., Thomson and Hutton, 2004; Cartwright and Hansen, 2006; Magee et
526 al., 2014; Magee et al., 2016). These sill complexes comprise a stacked series of mafic
527 intrusions (e.g., the Golden Valley Sill Complex, Karoo LIP, and sill complexes in the North
528 Atlantic igneous province, see Magee et al., 2016 for a review), contrasting with magma
529 systems conventionally depicted for many extensional rift systems (Wright et al., 2012; e.g.,
530 magmatic rift segments of Iceland and East Africa: Muirhead et al., 2015; Urbani et al.,
531 2015). AMS studies addressing magma flow within the intrusive systems of sill-dominated
532 LIPs are rare compared to studies investigating sub-parallel swarms of dikes (e.g., Delcamp
533 et al., 2014; Eriksson et al., 2014 and references therein). Below we discuss magma transport
534 dynamics within dikes and sills of the Ferrar LIP.

535

536 **7.1. Magma transport at Terra Cotta Mountain**

537

538 Structural and kinematic observations at Terra Cotta Mountain suggest a sill
539 source underlies the exposed dike network (Muirhead et al., 2012). Although many dikes
540 exhibit a lateral flow component, 34% of sampled sub-sections exhibit sub-vertical magma
541 flow paths ($>45^\circ$), suggesting that the dike swarm probably transported magma upward from
542 this underlying sill. Many of the dikes of this swarm were locally fed upward from large
543 (>10 m thick) “parent” intrusions. For example, a complex network of dikes is observed
544 branching outward from the top of d#3. Magma flow paths along a 280 m-wide region of d#3
545 are sub-vertical, suggesting that magma travelled upward into the overlying dikes adjoining
546 the upper contact of the intrusion. The replacement of ortho- and clino-pyroxene by
547 plagioclase moving up-dip, determined petrographically, supports a model of vertical flow
548 through the central region of this intrusion. The intrusions overlying d#3 can be seen merging

549 into the large sill (s#4) that caps Terra Cotta Mountain, and probably fed magma vertically
550 into the base of the intrusion.

551 We interpret the 42° range in distribution of AMS flow paths defined at s#4b
552 sub-sections as a consequence of multiple injection points at the base of the intrusion. Indeed,
553 Muirhead et al. (2012) document at least forty dikes ascending the stratigraphy, many of
554 which connect to the base of s#4, and our AMS results suggest these dikes fed magma
555 upward into this sill intrusion. From these feeder intrusions, we infer that magma flowed
556 outward along radial paths to produce the observed complex magma flow trajectories.

557

558 **7.2. Magma transport at Mt Gran**

559

560 At Mt Gran, AMS flow directions in dikes define a dominantly sub-vertical
561 flow (67% of data). In the 30 m-thick d#7, 60% of AMS flow directions are >60°, despite
562 coarse glomeroporphyritic textures away from dike margins, suggesting extensive late-stage
563 crystal growth under slow cooling rates. Anomalous, shallowly plunging magma flow,
564 constrained from intermediate AMS fabrics (Mg 14-1 and Mg 18-1), reflects a mix of oblate
565 and prolate contributions by magnetic particles to the rock's overall AMS fabric (e.g. Ferré,
566 2002; Aubourg et al., 2008), as well as local variations (vertically and laterally) of magma
567 flow, perhaps owing to pulsation in magma supply across the intrusion.

568

569 **7.3. Controls on magma emplacement and fracture dynamics**

570

571 The number and geometric complexity of the localized intrusive networks
572 dispersed throughout south Victoria Land (e.g., Allan Hills, Coombs Hills, Mt Gran, Terra
573 Cotta Mountain) point to the key role of local magmatic stresses in driving dike formation by

574 host-rock fracturing during the forceful intrusion of sills (Muirhead et al., 2012; Muirhead et
575 al., 2014). Our AMS data suggest that dikes ascended from these larger sill intrusions,
576 diverging along several trajectories, intruding both along the walls of pre-existing intrusions,
577 newly formed fractures, and bedding horizons (Fig. 13). Magma deflection along bedding
578 planes represents the primary control on intrusion propagation by pre-existing structures (Fig.
579 13; see also Airoidi et al., 2011). Up-dip and along-strike variations in dike attitude in other
580 parts of south Victoria Land (e.g., Allan-Coombs Hills: White et al., 2009; Muirhead et al.,
581 2012) represent the response of intrusions to local deviations in the principal stress directions
582 in an otherwise homogeneous and isotropic stress field (Airoidi et al., 2011; Muirhead et al.,
583 2014). Such stress rotations are shown in previous studies to be provided by rigidity contrasts
584 in the layered propagation medium (Gudmundsson and Brenner, 2004; Kavanagh et al.,
585 2006), stress concentrations and rotations related to sill inflation (Johnson and Pollard, 1973;
586 Malthe-Sørenssen et al., 2004; White et al., 2005), and intermittent magma propagation
587 resulting in fluctuating stress concentrations ahead of crack tips, upon both dike and sill
588 inception (Kavanagh et al., 2015), and later cooling (Chanceaux and Menand, 2014).

589 Variations in magma flow paths along individual intrusions suggest that
590 magma transport cannot be explained purely through a simple vertical flow model. The variety
591 of dike orientations and flow directions, coupled with evidence of multiple injections,
592 suggests that magma propagated intermittently. Variations between shallowly dipping to
593 vertical flow may represent distinct modes of magma flow occurred through time across a
594 single intrusion. For example, as dikes widened (in some instances to >10 m), variations in
595 magma crystallinity and viscosity between intrusion margins and interiors may have
596 coincided with the development of distinct flow paths and velocities. Alternatively, dominant
597 vertical flow might have changed with time to a lateral one, or vice versa. Temporal
598 variations in magma flow directions could result, for example, from changes in magma

599 buoyancy from crystallization and/or magma degassing, or changes in magnitude or direction
600 of driving pressure resulting from opening of new interconnected dikes/sills. Variable magma
601 flow in Ferrar dikes at Allan Hills were interpreted by Airoidi et al. (2012) as the result of
602 “passive” injection of magma into zones of intense fracturing above inflating sills. In this
603 model, magma pressures generated at the dike-tip are not the primary force driving dike-
604 fracture growth and propagation through the host. Instead, opening of country rock fractures
605 formed during sill-related deformation (e.g., Johnson and Pollard, 1973) creates pressure
606 gradients that draw magma into these highly strained zones. Field relationships throughout
607 the region imply that dike intrusions at Terra Cotta Mountain are underlain by a >200 m-thick
608 sill (Morrison and Reay, 1995; Marsh, 2004) and dike-emplacment orientations were
609 probably controlled by local stress conditions related to the inflation of a large underlying
610 sill, rather than by the far-field tectonic stress state (Muirhead et al., 2012; Muirhead et al.,
611 2014). Consequently, variations in dike attitude and magma flow direction recorded at Terra
612 Cotta Mountain are consistent with the sill-driven model of fracture growth and magma
613 propagation of Airoidi et al. (2012).

614

615 **7.4. Emplacement of the Ferrar LIP during Gondwana breakup**

616 **7.4.1 Magmatic-tectonic environment of the Ferrar LIP**

617

618 Ferrar magmas were originally proposed to have been emplaced in extensional
619 basins in a back-arc rift setting (e.g., Elliot and Larsen, 1993; Storey, 1995), but structural
620 evidence consistent with a rifting environment is absent across the Transantarctic Mountains.
621 For example, no significant Jurassic-age normal faults or long, colinear dike swarms, like
622 those in Iceland and East Africa (Wright et al., 2012; Muirhead et al., 2015), are observed.
623 The thickness (~2,500 m) and age (Devonian to Jurassic) of the Beacon Supergroup are

624 consistent with subsidence rates of only $0.011\text{-}0.014\text{ mm yr}^{-1}$, which is 1-2 orders of
625 magnitude lower than in active continental rift settings ($10^{-1}\text{-}10^0\text{ mm yr}^{-1}$), even those
626 exhibiting extension rates of only a few mm yr^{-1} (e.g., the Kenya Rift Valley, Birt et al.,
627 1997). Rare observations ($n=2$) of monoclines by Elliot and Larsen (1993) in Ferrar basalt
628 and tuff layers, originally interpreted as fault-related folds (i.e., Grant and Kattenhorn, 2004),
629 are more likely the result of folding at the termination of sills (Hansen and Cartwright, 2006a;
630 Magee et al., 2014), like that demonstrated at Allan Hills, Mt Fleming and Shapeless
631 Mountain (Grapes et al., 1974; Korsch et al., 1984; Pyne, 1984; Airoidi et al., 2011). Elliot
632 (2013) suggested that the substantial thickness of the Kirkpatrick basalts (up to 230 m)
633 necessitates confining topography resulting from rift-related subsidence. However, sill-driven
634 uplift is also shown to produce significant surface topography, resulting in the formation of
635 basins, 100s of meters deep and 10s of kilometers long, observed above sill complexes in
636 seismic reflection imaging (Trude et al., 2003; Hansen and Cartwright, 2006b). At Shapeless
637 Mountain for instance, the intrusion of Ferrar sills produced differential uplift that, in places,
638 would have produced $>200\text{ m}$ -high topography at the surface (Korsch et al., 1984). A distinct
639 absence of Jurassic-age normal faulting, rift basin subsidence, and long sub-parallel dike
640 swarms thus provide compelling evidence that Ferrar LIP emplacement was not accompanied
641 by extensional tectonics and the formation of continental rift basins. This conclusion is
642 further supported by the orientations of >600 Ferrar dikes in south Victoria Land (Muirhead
643 et al., 2014). These dikes show no preferred alignments, consistent with emplacement in an
644 isotropic stress regime. Dike orientations were instead controlled by local magmatic stresses
645 related to the emplacement of sills.

646 These observations suggest that the East Antarctic Margin was not subjected
647 to significant regional tectonic stresses prior to and during Ferrar LIP emplacement, which
648 may explain why continental breakup did not initiate throughout Antarctica during and after

649 Ferrar-Karoo magmatism. Instead, breakup of the Gondwana supercontinent focused in what
650 is now the Weddell Sea region.

651

652 **7.4.2 Broad-scale emplacement of the Ferrar LIP**

653

654 Igneous rocks of the Ferrar LIP crop out in present day Antarctica, SE
655 Australia, and New Zealand. Although the Ferrar magmatic province exhibits a broad
656 geographic distribution, the remarkably homogenous compositions exhibited by Ferrar rocks
657 suggest a single source (Elliot et al., 1999). Furthermore, decreasing Mg# and MgO contents
658 away from the inferred source area (Weddell Sea) along the length of the province are
659 consistent with fractional crystallization during lateral magma flow (Elliot and Fleming,
660 2000; Leat, 2008). Widespread Ordovician dikes of the Vanda dike swarm have been mapped
661 in basement granitoids and meta-sedimentary rocks (Allibone et al., 1993), yet no Ferrar
662 dikes, or sills, are observed at basement depths, below the Basement Sill, ~500 m from the
663 lower contact of the Beacon Supergroup (Marsh, 2004; Leat, 2008). As suggested by Leat
664 (2008), these observations imply that long-distance, lateral magma transport in the Ferrar LIP
665 occurred almost exclusively in the Beacon Supergroup (Fig. 14).

666 AMS flow directions from dike swarms presented in this study, as well as
667 those at Allan Hills (Airoidi et al., 2012), provide insights into regional trends in magma flow
668 in dikes across arguably the best-exposed portion of the province in south Victoria Land (Fig.
669 14). These results show no consistent lateral or vertical flow components that typically
670 characterize the dike feeder systems of LIPs (Ernst and Baragar, 1992). It is therefore likely
671 that the observed Ferrar dikes were not responsible for the regional transport of magma
672 laterally along the East Antarctic margin during Ferrar magmatism. Instead, magma flow in
673 dikes was of local importance, creating pathways through which magma could ascend the

674 stratigraphy through interconnected sills and, eventually, erupt at the surface (Muirhead et al.,
675 2014).

676 As Ferrar dikes were not responsible for long distance magma transport along
677 the province, we highlight the Ferrar sills as the probable conduits throughout which magma
678 was transported laterally ~3,500 km within Beacon Supergroup rocks across the Gondwana
679 supercontinent from its source (see conceptual model in Fig. 14). Lateral magma transport in
680 sills for ~3,500 km would require both sufficient magma input and appropriate thermal
681 conditions to avoid solidification and arrest (Annen and Sparks, 2002; Chanceaux and
682 Menand, 2014). One of the main limiting factors to long-distance transport relates to the
683 original magma temperature and the temperature at which magma freezes and stalls
684 (solidification temperature). However, magma intrusions and lavas flows in LIPs are shown
685 to travel >1,000 km with assistance of shear heating (i.e., viscous dissipation) (Fialko and
686 Rubin, 1999) and the insulating effects of a fine-grained chilled intrusive margin (e.g.,
687 Delaney and Pollard, 1982; Marsh, 2002), or an external crust (Keszthelyi and Self, 1998). In
688 many ways, sills are similar to lava flows, as they form through the progressive lateral
689 propagation, inflation and linkage of magma fingers or lobes (Pollard et al., 1975; Thomson
690 and Hutton, 2004; Schofield et al., 2010; Schofield et al., 2012), some of which reach
691 thicknesses of 10 to 100s of meters (Hansen and Cartwright, 2006a; Schofield et al., 2015).
692 Although magma propagation speeds for upper crustal basaltic intrusions (e.g., 10^{-1} to 10^{-2} m
693 s^{-1} ; Wright et al., 2012) are lower than for basalt lavas (10^1 to 10^{-1} m s^{-1} ; e.g. Keszthelyi and
694 Self, 1998; Self et al., 1998; Self et al., 2008), intrusions are comparatively more insulated
695 because they are surrounded by warmer crustal rocks rather than cool atmosphere.

696 The shallow plumbing system of Antarctica comprises a series of stacked,
697 interconnected sills. The four major sills in south Victoria Land average ~300 m in thickness
698 (Marsh, 2004; Elliot and Fleming, 2008). By applying a heat balance model similar to that

699 developed by Keszthelyi and Self (1998) for lava flows with an insulating crust, the thermal
700 efficiency of sill flow is determined by viscous heating, conductive heat loss from the upper
701 and lower margins of the intrusion, and thermo-physical parameters. Assuming conservative
702 magma propagation velocities of 0.03, 0.05 and 0.1 m s⁻¹, negating the effects of viscous
703 heating, and applying typical thermo-physical properties for mafic magmas (injection
704 temperature of 1,250 °C, density 2,700 kg m⁻³, thermal conductivity of 2.1 J s⁻¹ m⁻¹ °C,
705 specific heat of 1,200 J kg⁻¹ °C; Barker et al., 1998; Wohletz et al., 1999; Wang et al., 2010),
706 we estimate that the maximum heat loss for a 300 m-thick Ferrar sill emplaced at 2.5 km
707 depth would be between 0.04 and 0.01 °C km⁻¹ (refer to Section 2 of the supplementary
708 material). These thermal constraints would have affected transport of magma in 300 m-thick
709 Ferrar sills for 3,500 to 12,000 km along the East Antarctic margin before solidifying.

710 The thermal constraints on long-distance magma transport in Ferrar LIP sills
711 may be tested further by estimating the minimum sill thickness required for sustained magma
712 flow (cf. Holness and Humphreys, 2003, and refer to Section 2 of the supplementary text).
713 Adopting the thermo-mechanical intrusion parameters of Holness and Humphreys (2003),
714 and assuming a constant and conservative overpressure at the magma source equal to the
715 tensile strength of rock (3 MPa; Schultz, 1995), we estimate that lateral magma transport for
716 3,500 km across the East Antarctic Margin would require a minimum sill thickness of 110 m.
717 Long-distance magma flow would be further assisted by the channeling of Ferrar sills within
718 the Beacon Supergroup sedimentary basin (Leat, 2008), which is laterally continuous along
719 the full length of the East Antarctic margin (Barrett, 1981). These estimates are similar to
720 those obtained for some of the longest identified Deccan lava flows (Rajahmundry lavas: Self
721 et al., 2008), which advanced >1,000 km in a cooler subaerial environment, and are
722 consistent with field studies and thermo-mechanical models of long-distance transport in

723 giant dikes, which in some instances extend for 1000s of km laterally from their source (e.g.,
724 the MacKenzie dike swarm: Ernst and Baragar, 1992; Fialko and Rubin, 1999).

725

726 **8. Conclusions**

727 Magma flow in dike swarms of the Ferrar LIP investigated in this study
728 contrast with flow paths predicted by classic models of dike transport in LIPs and magmatic
729 rift settings (Ernst and Baragar, 1992; Wright et al., 2012). Our data suggest that dikes
730 transported magma vertically between sills rather than controlling long-distance lateral
731 transport in sub-parallel swarms throughout SVL.

732 AMS data provide new insights into the growth of sill-fed dike swarms in
733 LIPs. The heterogeneity of magma flow and variability in dike attitudes at various depths and
734 scales (from a few meters to kilometers) suggest that tectonic stresses had little influence on
735 the growth of the intrusive networks. A complex flow model, with both shallowly dipping
736 and sub-vertical flow components, can be defined for most intrusions. Variable magma flow
737 within individual intrusions may have developed along strike and up-dip either during a
738 single intrusive event, and/or as a result of multiple injections, and locally represent early- vs
739 late-stage magma propagation.

740 The Ferrar LIP formed during the earliest stages of Gondwanaland breakup
741 and was originally interpreted to be emplaced in a back-arc extensional setting. However,
742 fracture systems trends, dike patterns, and magma flow patterns in SVL are consistent with
743 magma emplacement in an isotropic stress regime, with bedding anisotropy providing the
744 dominant structural control on intrusion geometries. Flow patterns observed regionally in
745 dike swarms across south Victoria Land show no consistent lateral or vertical flow directions.
746 As Ferrar dikes were not responsible for the long-distance transport of magma laterally across
747 Antarctica, the Ferrar sills remain the most likely candidate for long-distance transport.