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This is the author's manuscript

Original Citation:

Availability:

This version is available http://hdl.handle.net/2318/1584577

since 2017-05-24T10:58:49Z

Published version:

DOI:http://dx.doi.org/10.1016/j.tecto.2016.06.029

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(Article begins on next page)

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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) flow paths in dike networks at Terra Cotta Mountain and Mt Gran, which intruded at 29 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 negligible or absent. Unlike giant dike swarms in LIPs elsewhere (e.g., 1270 Ma MacKenzie 37 38 LIP), dikes of the Ferrar LIP show no regionally consistent vertical or lateral flow patterns, 39 suggesting these intrusion where 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 \geq 4 km, we suggest that long distance magma transport occurred in sills within 42 Beacon Supergroup sedimentary rocks. This interpretation is consistent with existing geochemical data and thermal constraints, which support lateral magma flow for ~3,500 km 43 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

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 depicted as colinear swarms of giant dikes (Ernst et al., 1995; Ernst et al., 2001). The primary 56 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"; Airoldi et al., 2011), exhibit short lengths (<5 km), variable dips (10-90°), form at sill peripheries, and link 66 67 sills at different stratigraphic levels (Johnson and Pollard, 1973; Czamanske et al., 1995; 68 Muirhead et al., 2012).

Although sill-fed dikes form a key component of the shallow plumbing systems of sill-dominated LIPs, magma flow dynamics within these intrusions remain largely unknown. Many studies focus on magma transport through the outer sheets and internal sills of saucer-shaped intrusions (Ferré et al., 2002; Thomson and Hutton, 2004; Hansen and Cartwright, 2006a; Maes et al., 2008; Polteau et al., 2008b; Galland et al., 2009). In the 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., 2011). Airoldi et al. (2012), however, revealed complex lateral and vertical magma flow 77 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 (Encarnación 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.

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2. The Ferrar large igneous province

Ferrar LIP rocks are exposed for 3,500 km along the Transantarctic Mountains of East Antarctica. These intrusive and extrusive rocks, with a total estimated volume around 300,000 km³ (Ross et al., 2005), represent the most laterally extensive LIP system on Earth. 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. Encarnación 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).

Ferrar intrusions are observed dissecting the flat-lying, ~2.5 km-thick Beacon 112 113 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).

Swarms of shallow to moderately dipping dike intrusions are reported from various localities in the central Transantarctic Mountains (e.g. Hornig, 1993; Leat, 2008 and references therein) and south Victoria Land (Skinner and Ricker, 1968; Wilson, 1993; Morrison and Reay, 1995; Muirhead et al., 2014). Regional field and remote sensing studies reveal that these intrusions connect sills at different stratigraphic levels, and are inferred to assist in the vertical transport of magma in the upper 4 km of the plumbing system to the surface (Muirhead et al., 2012; Muirhead et al., 2014). Magma flow dynamics within these 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 (Muirhead et al., 2012). Mt Kuipers lies immediately east of Terra Cotta Mountain, where a 137 ~200 m-thick sill and two dike intrusions can be seen on the western flanks. Basement 138 139 granitoids exposed at the northern foothill of the nunatak underlie a sequence of guartz-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., 2008). A ~1.0 km-thick sequence of relatively undeformed, flat-lying, sedimentary rocks of 143 144 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 Sandstone, which suggests emplacement at paleodepths of $\sim 1.5 - 2.5$ km (cf. Fig. 1) 146 147 (Morrison, 1989; Muirhead et al., 2012).

148Dike attitudes at Terra Cotta Mountain are irregular, with orientations varying149along strike to form zig-zag patterns. Intrusion dips also vary up-section, where some

intrusions change into sills that connect offset dike segments, resulting in a transgressive dike
geometry (Airoldi et al., 2011). Some intrusions bifurcate locally into smaller dikes. For
example, dikes up to 10 m thick can be seen connecting to smaller (2-6 m thick), 'offshoot'
dikes, some of which in turn feed into thinner (<2 m thick) ones (Fig. 2a). Offshoot dikes are
characterized by irregular geometries, with different segments exhibiting left- and right-steps,
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 chilled dolerite fragments are observed within some dikes, and trend sub-parallel to the 163 164 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 include chaotic arrays of angular fragments and/or rounded pods of chilled dolerite,intermingled with lithified medium-to-fine sand material (Fig. 3d).

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178 **3.1.1. Target intrusions**

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180 The principal intrusions investigated in this study are shown in Fig. 5. At the base of Terra Cotta Mountain is a >20 m-thick intrusion (d#3). The lower contact of the 181 intrusion is visible, striking 340° and dipping east at 75°. However, the upper contact cannot 182 be seen anywhere in the field area, and the intrusion does not appear to have significant 183 184 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 exhibit alternating dike-sill geometries (e.g., t#1a-b, and t#2). On the western slopes of Mt 187 Kuipers is a ~ 200 m-thick sill, a ~ 20 m-thick, $\sim 090^{\circ}$ striking dike (d#5), and a 10 m-thick, 188 $\sim 160^{\circ}$ trending dike (d#6). The latter dike (d#6) tapers down and ends to the north-west 189 190 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.

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3.2. Mount Gran

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

of 1-1.5 km (White et al., 2009; Fig. 1). The most prominent intrusion at Mt Gran is a ~30 m-200 201 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 $<30^{\circ}$) 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 common in the last few decades due to its time- and cost-effectiveness, and is commonly 212 213 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

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 magnetization (IRM) acquisition, thermal demagnetization and backfield curves were 243 244 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 coercivity (B_{CR}), temperature-dependant susceptibility ($K_T vs T$), blocking (T_B), and Curie temperatures (T_C) at which magnetic minerals lost their magnetic properties.

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

258

259 4.2. Interpretation of magnetic fabrics

260

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

The anisotropy of magnetic susceptibility, or AMS, is modelled as an ellipsoid with mutually orthogonal axes $k_1 \ge k_2 \ge k_3$ (respectively, maximum, intermediate and minimum susceptibility axes). These axes can be graphically plotted as lineations on equal area stereographic projections (Fig. 6). The anisotropy parameters defined for any magnetic fabric ellipsoid are the mean magnetic susceptibility (K_m) and anisotropy degree (P or P_J, corrected anisotropy degree) defining the absolute anisotropy of a rock specimen, and magnetic lineation (L), foliation (F), and shape parameter (T) (see Tarling and Hrouda, 1993, table 1.1, p. 18, for their mathematical expression). Together, L, F and T define the geometry of the AMS ellipsoid. Prolate fabric ellipsoids are elongate (L > F) and characterized by – $1 \le T < 0$, whereas oblate ellipsoids are flattened (F > L) and characterized by $0 \le T \le 1$. In the directional analysis of AMS fabrics, magnetic lineation and foliation correspond respectively to the maximum susceptibility axis direction k₁, and to the plane perpendicular to k₃ and defined by k₁ and k₂ 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).

Magnetic fabrics in intrusive rocks are, however, also known to exhibit deviations from the normal fabric described above. Imbrication angles of 30° to 45° between the FPL and intrusion plane, as well as the exchange of the intermediate and minimum axes of the fabric ellipsoid, are commonly related to composite magnetic mineralogy of AMS sources with different properties (e.g. Ferré, 2002; Aubourg et al., 2008). These are known as *intermediate fabrics* (I₁ – I₃ in Fig. 6), and are classified after Airoldi et al. (2012, and references therein) as 3 types:

I₁ AMS fabrics are prolate, with the magnetic lineation lying within 45° from the
 intrusion plane and k₂ and k₃ dispersed on a girdle.

I₂ fabrics are either prolate or oblate, have both k₁ and k₃ aligned with the intrusion
 plane and FPL orthogonal to it.

• I₃ is a 'nearly normal' planar fabric, with intrusion and magnetic foliation planes sub-300 parallel to one another, and imbrication or intersection angle >30°; 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).

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

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- 311 4.3. Corroboration of AMS data
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313 AMS fabrics were also compared with macroscopic indicators of magma flow observed in the field. In these instances, the shape, trend and plunge of preserved 314 315 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 structure between dike and sill segments (Airoldi et al., 2012 and references therein) or 317 318 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 al., 2015). Striations on dike walls could represent both magma flow, and shear between 322 magma and encasing rocks related to dike opening (e.g. Correa-Gomes et al., 2001; Baer et 323

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

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5. Source of Ferrar Dolerite magnetism

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

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335 **5.1. Petrographic characteristics**

336

Ferrar dolerites from the two study locations are compositionally and 337 338 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.

At Terra Cotta Mountain, d#3 and d#5 are characterized by the above mineral assemblage, with orthopyroxene enstatite and clinopyroxene augite and pigeonite crystals around 50-60%, and commonly \leq 40% plagioclase crystals. Within t#1a-b and t#2, orthopyroxene becomes less common, and clinopyroxene and plagioclase increase in abundance up-section. S#4 and the intrusions at Mt Gran are petrographically similar. In these intrusions, plagioclase is the most abundant mineral phase (40-60%), with 25-40% clinopyroxene (Aug±Pig), and <20% orthopyroxene.

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

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

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- 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 μ SI and 423 to 40800 μ SI, 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

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

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

Saturation of Ferrar specimens is reached with applied field values (B_S) of 300 mT, indicating a dominant low-coercivity magnetic phase. Additional irregular steps observed in the IRM decay curves are likely due to demagnetization of soft and medium B_{CR} fractions (B_{CR} ranging between 30 and 60-70 mT) during application of the back field (Fig. 80) 8c).

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392 5.3. Interpretation of magnetic carriers

393

Magnetic saturation, remanence coercivity and Curie temperature values determined for Terra Cotta Mountain dolerites, with remanence coercivity overlap in IRM plots, indicate the presence of both soft and medium remanence coercivity magnetite and/or maghemite (cf. Borradaile and Jackson, 2004 and references therein). Magnetite is the 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 hydrothermal event in south Victoria Land (e.g. Craw et al., 1992; Ballance and Watters, 402 2002). In fact, the predominance of K_m values >10⁻³ SI indicates contributions from both 403 404 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 magnetic carriers (for instance, accessory magnetic minerals such as pyrrhotite and 411 412 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 variations in IRM are, however, consistent with breakdown of pyrrhotite around 300-400°, 417 418 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 k₁ 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, respectively. 40% of Mt Gran dataset (s#8, two d#11 sites and three of the eastern 436 sills/shallowly dipping sheets sites) is characterized by anomalous oblate reverse fabrics. 437 438 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

Samples from Terra Cotta Mountain and Mt Gran were grouped into 35 and 16 sub-sections, respectively (Tables 1 and 2). Each sub-section contains data analyzed from 1 to 4 sample sites on individual intrusions. Samples within each sub-section produced 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 I₃ 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 sites, and see d#5-6b in Fig. 6), the flow direction is $<30^{\circ}$ from the k₁ or k₂ axes. Except for 450 451 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 flow directions trend within 35° of the macroscopic indicators. A similar fit was observed 453 454 between macroscopic and magnetic flow fabrics at Allan Hills (70% of AMS fabrics are within 35°: Airoldi et al., 2012) and intrusive swarms elsewhere (Ardnamurchan, Scotland: 455 456 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

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

468

469 6.2.2. Magma flow at Terra Cotta Mountain

470

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

whereas steeper flow paths (>25°) are observed only in the steeper dike segments (intrusion 474 dips >50°) (t#1b). Dikes are characterized by variable magma flow paths (Fig. 9 and 11). 475 476 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, analysis of 12 sub-sections along d#5 reveal magma flow plunges ranging 7-79° (Table 1 and 478 479 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 267-336°, with shallow-tomoderate plunges (13-37°) along the south-western margin of the intrusion (d#3-1 and d#3-481 482 4). Steeper flow plunges (46°-68°) 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 flow-modes is, however, not always observed in thinner dikes (width <10 m, e.g., d#6), 485 486 although this may in part be the result of smaller AMS sample sets across some of these 487 intrusions.

AMS flow lineations from the Terra Cotta Mountain summit sill (s#4) exhibit westward trends (268-310°), with sub-horizontal plunges (<10°). These magnetic lineations are sub-parallel to lineations of the macroscopic flow indicators.

491

- 492 6.2.3. Magma flow at Mt Gran
- 493

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

Terra Cotta Mountain and Mt Gran samples are characterized by both prolate and oblate magnetic susceptibility ellipsoids, with normal AMS fabrics adding up to about 20%, and intermediate ones to ~45% of the total data respectively. Isotropic or inverse fabrics 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 along individual intrusions, with flow plunges as low as 7° and as steep as 79° (19°-78° at Mt 514 515 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

Long-distance magma transport in LIPs is often depicted to occur through the emplacement of giant dikes, 100s of km long and 10s of m thick (Ernst et al., 1995). These dikes are shown to have transported magma >1000 km laterally away from an inferred plume source (e.g., MacKenzie dike swarm, Ernst and Baragar, 1992). The development of sill524 dominated magmatic systems within LIPs, however, has been increasingly recognized over the past decade (e.g., Thomson and Hutton, 2004; Cartwright and Hansen, 2006; Magee et 525 526 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 Atlantic igneous province, see Magee et al., 2016 for a review), contrasting with magma 528 529 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°), suggesting that the dike swarm probably transported magma upward from this underlying sill. Many of the dikes of this swarm were locally fed upward from large 542 (>10 m thick) "parent" intrusions. For example, a complex network of dikes is observed 543 544 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 into the large sill (s#4) that caps Terra Cotta Mountain, and probably fed magma verticallyinto the base of the intrusion.

We interpret the 42° range in distribution of AMS flow paths defined at s#4b sub-sections as a consequence of multiple injection points at the base of the intrusion. Indeed, Muirhead et al. (2012) document at least forty dikes ascending the stratigraphy, many of which connect to the base of s#4, and our AMS results suggest these dikes fed magma upward into this sill intrusion. From these feeder intrusions, we infer that magma flowed 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 crystal growth under slow cooling rates. Anomalous, shallowly plunging magma flow, 563 564 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 Airoldi 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 (Airoldi 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 though 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 Airoldi 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-emplacement 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 Airoldi 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

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

consistent with subsidence rates of only 0.011-0.014 mm yr⁻¹, which is 1-2 orders of 624 magnitude lower than in active continental rift settings $(10^{-1}-10^{0} \text{ mm yr}^{-1})$, even those 625 exhibiting extension rates of only a few mm yr⁻¹ (e.g., the Kenya Rift Valley, Birt et al., 626 627 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; Airoldi 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 seismic reflection imaging (Trude et al., 2003; Hansen and Cartwright, 2006b). At Shapeless 636 637 Mountain for instance, the intrusion of Ferrar sills produced differential uplift that, in places, 638 would have produced >200 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 Ferrar-Karoo magmatism. Instead, breakup of the Gondwana supercontinent focused in whatis 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 (2008), these observations imply that long-distance, lateral magma transport in the Ferrar LIP 664 665 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 (Airoldi 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 stratigraphy through interconnected sills and, eventually, erupt at the surface (Muirhead et al.,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 was transported laterally ~3,500 km within Beacon Supergroup rocks across the Gondwana 678 679 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). Although magma propagation speeds for upper crustal basaltic intrusions (e.g., 10^{-1} to 10^{-2} m 692 s^{-1} ; Wright et al., 2012) are lower than for basalt lavas (101 to 10^{-1} m s^{-1} ; e.g. Keszthelyi and 693 694 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.

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

developed by Keszthelyi and Self (1998) for lava flows with an insulating crust, the thermal 699 700 efficiency of sill flow is determined by viscous heating, conductive heat loss from the upper and lower margins of the intrusion, and thermo-physical parameters. Assuming conservative 701 magma propagation velocities of 0.03, 0.05 and 0.1 m s⁻¹, negating the effects of viscous 702 703 heating, and applying typical thermo-physical properties for mafic magmas (injection temperature of 1,250 °C, density 2,700 kg m⁻³, thermal conductivity of 2.1 J s⁻¹ m⁻¹ °C, 704 specific heat of 1,200 J kg⁻¹ °C; Barker et al., 1998; Wohletz et al., 1999; Wang et al., 2010), 705 706 we estimate that the maximum heat loss for a 300 m-thick Ferrar sill emplaced at 2.5 km depth would be between 0.04 and 0.01 °C km⁻¹ (refer to Section 2 of the supplementary 707 708 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 for giant dikes, which in some instances extend for 1000s of km laterally from their source (e.g.,

the MacKenzie dike swarm: Ernst and Baragar, 1992; Fialko and Rubin, 1999).

725

726 8. Conclusions

Magma flow in dike swarms of the Ferrar LIP investigated in this study contrast with flow paths predicted by classic models of dike transport in LIPs and magmatic rift settings (Ernst and Baragar, 1992; Wright et al., 2012). Our data suggest that dikes transported magma vertically between sills rather than controlling long-distance lateral 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 the growth of the intrusive networks. A complex flow model, with both shallowly dipping 735 736 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 dominant structural control on intrusion geometries. Flow patterns observed regionally in 744 745 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 Antarctica, the Ferrar sills remain the most likely candidate for long-distance transport. 747