

This manuscript is a **preprint** uploaded to **EarthArXiv** and has been submitted for publication in **EARTH AND PLANETARY SCIENCE LETTERS**. The manuscript will be updated throughout the peer-review process and the content may change until the manuscript is accepted for publication. We therefore encourage downloading the latest version from EarthArXiv before usage.

1 **The Ediacaran Grenville dykes (SE Canada) reveal the weakest**
2 **sustained palaeomagnetic field on record.**

3 Daniele Thallner^{1*}, Andy Biggin¹, Henry Halls²

4 ¹Dept. of Earth, Ocean and Ecological Sciences, University of Liverpool, L69 3BX, UK

5 ²Department of Earth Sciences, University of Toronto, 22 Russel Street, Toronto, Canada M5S
6 3B1

7 *Corresponding author:

8 Daniele Thallner

9 Email: daniele.thallner@liverpool.ac.uk

10 Address: Dept. Earth, Ocean and Ecological Sciences, University of Liverpool, L69 3BX, UK

11 **Keywords:** Paleointensity; Ediacaran; Laurentia; Grenville Dykes

12

13 **Abstract**

14 Long-term variations of the geomagnetic field, observed in the palaeomagnetic record, have
15 the potential to shed much light on the evolution of Earth's deep interior. With a geomagnetic
16 field characterised by anomalous directions and ultra-low intensities, the Ediacaran period
17 (635-541 Ma) is a time of special interest. Steep and shallow directions, leading to virtual
18 geomagnetic poles (VGPs), separated by angles of up to 90° and very close in age could have
19 recorded a geomagnetic field switching between axial and equatorial dipole-dominated
20 states. Alternatively, the field may simply have been highly nondipolar and subject to rapid
21 reversals. Palaeointensity determinations of units that record the anomalous directions could
22 potentially help to discriminate between morphologies but the spatial and temporal

23 distribution of palaeomagnetic data require improvement. Here we present new
24 palaeointensities from 11 sites from the western end of the Grenville Dyke swarm that
25 recorded directionally anomalous geomagnetic fields around ~585 Ma. Palaeointensities,
26 obtained through microwave Thellier, Shaw and pseudo-Thellier methods, show field
27 strength values of $2.9 \pm 2.2 \mu\text{T}$ and corresponding virtual dipole moments of $0.3\text{-}1.7 \times 10^{22} \text{ Am}^2$.
28 These field strengths are an order of magnitude weaker than the present-day field. The most
29 extreme palaeointensity values of $1.4\text{-}2.1 \mu\text{T}$ are half as strong as seen in previous studies of
30 the Ediacaran field and as low as Mars' recently measured crustal field intensity, giving a new
31 lower bound for the Earth. VGPs grouping in two distinct clusters with almost identical angular
32 dispersions of VGPs ($S_B = 18.5^\circ$ and 18.9°) may argue for the presence of an equatorial dipole.
33 In contrast, the palaeointensities associated with the steep and shallow components are
34 indistinguishable. This observation, together with the overall very large VGP dispersion may
35 rather support that the Grenville Dykes have recorded enhanced secular variation linked to a
36 highly unstable, multipolar and reversing field.

37

38 **1. Introduction**

39 Extreme climatic changes in the late Neoproterozoic (Evans and Raub, 2011) and tectonic
40 activity between the final breakup of Rodinia and the assembly of Gondwana have made the
41 Ediacaran period (542-635 Ma) a prime target for palaeomagnetic and geodynamic research.
42 Especially the individual drifting histories of Laurentia and Baltica remain mysterious (Li et al.,
43 2013) although both their positions are relatively well constrained around the beginning and
44 the end of the Ediacaran at ~615 Ma and from ~550 Ma (e.g McCausland et al., 2007; Meert,
45 2014). Several sets of palaeomagnetic data are available to constrain the palaeogeography

46 for Laurentia (e.g. Bono and Tarduno, 2014; McCausland et al., 2007; McCausland and
47 Hodych, 1998) and Baltica (e.g. Bazhenov et al., 2016; Meert, 2014), but the resulting
48 palaeopoles lead to ambiguous apparent polar wander paths (AWPW). Datasets from
49 Laurentia and Baltica show both steep and shallow primary directional components that are
50 extremely close in age. For Laurentia, both components were found in studies of the Sept-Îles
51 intrusion (Bono and Tarduno, 2014; Tanczyk et al., 1987), the Grenville Dykes (Murthy, 1971,
52 Halls et al., 2015) and the Catoctin Basalts (Meert et al., 1994). These components usually
53 show dual-polarities and are often supported by field tests. The limitations imposed by the
54 nature of a geocentric axial dipole field lead to discussions about which of the two
55 components of each study could be less reliable and would not be used for the calculation of
56 the APWP (e.g. Bono and Tarduno, 2014; McCausland et al., 2007; Robert et al., 2017).
57 Depending on which of the seemingly primary components are rejected, high plate motion
58 velocities are required to allow the latitudinal changes in the resulting APWPs which lead to
59 different theories explaining the geodynamic processes in the Ediacaran, including inertial
60 interchange true polar wander (Kirschvink et al., 1997). In more recent studies, the puzzling
61 APWPs have been attributed to anomalous directional behaviour of the geomagnetic field
62 itself. Specifically, the steep and shallow directions have been interpreted as being caused by
63 the geomagnetic field flipping between an axial and an equatorial dipole field configuration
64 (Abrajevitch and Van der Voo, 2010) or as showing an intermediary state of an axial dipole
65 field during continuous reversals (Halls et al., 2015). Numerical geodynamo simulations have
66 suggested that the occurrence of a protracted equatorial dipole field may be possible during
67 the reversal of an axial dipole field (Aubert and Wicht, 2004; Gissinger et al., 2012) with
68 significant differences between the intensities of the axial and equatorial field states. Studies
69 of magnetostratigraphic data from the late Ediacaran and the early-mid Cambrian (Bazhenov

70 et al., 2016), resulting in reversal frequencies of more than 20 reversals/ Ma, claim that the
71 field might have been in a hyperactive state at that time. Due to the correlation between
72 reversal frequency and dipole field strength (Kulakov et al., 2019; Tauxe et al., 2013) the
73 dipole field strength in the Ediacaran is expected to be low. First palaeointensity studies of
74 Ediacaran rocks from Canada (Bono et al., 2019) and Ukraine (Shcherbakova et al., 2020) show
75 ultra-low virtual dipole moments (VDM) that are an order of magnitude weaker than dipole
76 moments in the Phanerozoic. To date, similarly low dipole moments have only been found in
77 the Devonian (Hawkins et al., 2019; Shcherbakova et al., 2017) and the Jurassic (Kulakov et
78 al., 2019; Tauxe et al., 2013). Information about reversals and palaeointensities are critical to
79 delineate the anomalous field behaviour in the Ediacaran. However, no estimates for the
80 strength of the geomagnetic field exist for the early-mid Ediacaran before 580 Ma. Here we
81 report multi-method palaeointensity measurements performed on mid-Ediacaran age units
82 showing both steep and shallow directions to look for differences that could help explain the
83 directional observations of that time.

84 **2. Materials and methods**

85 **2.1. Sample material:**

86 The samples from dykes within the Grenville province used in this study were those collected
87 for the study of Halls et al. (2015) and were taken preferentially from chilled margins. Several
88 dykes were sampled at multiple sites over distances of up to 150 km. Analysis of geochemical
89 composition helped with longitudinal correlation of the dykes. Samples were selected for
90 intensity determination if sister-samples from the same site showed a stable component of
91 ChRM in the previous study, which was the case for 99 samples from 15 sites within nine
92 dykes (Figure 1). Following the naming convention used in Halls et al. (2015), the five dykes

93 analysed in this study were: Coniston dyke (sites GD02, GD33), French River dyke (site GD23),
94 Key River dyke (sites GD10, GD15, GD16, GD19, GD37) and Sand Bay dyke (site GD29).
95 Additionally, four sites from other dykes in the area associated with the dyke swarm (sites
96 GD07, GD25, GD26, GD30) were also analysed. U-Pb ages exist for Sand Bay dyke (585.2 ± 0.8
97 Ma), Augusta Lake dyke (584 ± 0.6 Ma), Key River dyke (587.3 ± 0.7 Ma) and French River dyke
98 (598.0 ± 1.4 Ma).

99

100

101 Distances between sites of the same dyke were generally larger than 2km and the sites were
102 therefore regarded as separate spot readings of the field strength for the purpose of this
103 study. Differences in chemical composition between the individual dykes suggest single
104 intrusion events at different development stages of the magma chamber (Halls et al., 2015).
105 The directional results in Halls et al. (2015) show four components resulting in antipodal steep
106 and shallow directions close to directions of older studies of the Grenville dykes (Hyodo and
107 Dunlop, 1993; Murthy, 1971). Positive contact (Halls et al., 2015) and reversal tests (Hyodo
108 and Dunlop, 1993) support the primary nature of the steep direction. This comprises
109 components B and E, that can be seen in sites of the Key River dyke (Halls et al., 2015) which
110 are close to the directions recorded by the Mattawa dyke (585.9 Ma, Hyodo and Dunlop,
111 1993). The shallow component from the almost antipolar components C and D can also be
112 observed continuously in sites of the same dyke and component D differs significantly from
113 the regional overprint recorded in the close-by Whitestone anorthosite (Ueno et al., 1975).
114 Because of an inconclusive contact test, it cannot be excluded that this component was
115 remagnetised (Halls et al., 2015; Robert et al., 2017). In this study, we nevertheless analysed
116 samples from sites that recorded both steep and shallow components.

117 **2.2 Methods:**

118 All measurements were carried out at University of Liverpool's Geomagnetism Laboratory. At
119 least one specimen per site was selected for rock magnetic measurements. Measurements of
120 IRM, hysteresis parameters and thermomagnetic curves were done on a Variable Field
121 Translation Balance (VFTB) using both crushed specimens and cylinder specimens with a
122 diameter of 5 mm. Additionally, temperature dependent susceptibilities were measured on
123 an AGICO MFK-1A Kappabridge using crushed specimens heated in air to 700°C. To monitor
124 thermochemical alterations in more detail, the same measurements were also performed in
125 incremental heating and cooling cycles in temperature steps of 100°C between 150°C and
126 650°C. Scanning electron microscope (SEM) images were taken with a Hitachi TM3000
127 tabletop microscope and analysed with Quantax70.

128 The magnetic viscosity of 59 specimens was determined by measuring their remanent
129 magnetisation after being stored for three weeks in different orientations without shielding
130 from the ambient field (Prévot, 1981). For this experiment, the specimens were stored on a
131 rack in the laboratory with their z axis parallel to the field lines of the geomagnetic field for
132 the first duration and antiparallel to the field lines for the second duration. The remanent
133 magnetisations were measured on an AGICO JR6 spinner magnetometer.

134 Four different methods were used to acquire palaeointensity data:

135 Microwave Thellier type palaeointensity experiments were performed on the high frequency
136 (14GHz) microwave system and SQUID magnetometer (Hill and Shaw, 1999). This instrument
137 uses cylinder specimens with a 5mm diameter. These were drilled from the available 1"
138 cylinders, which allowed for 5 to 15 sub-specimens to be drilled from one standard cylinder
139 specimen. Specimens are subjected to microwave treatment with incrementally increasing

140 power and/or duration in a range between 20W.s and 500 W.s. The succession of zero-field
141 (Z) and in-field (I) steps for the experiments followed the IZZI protocol (Tauxe and Staudigel,
142 2004). A pTRM-check was done after every other I-Z-pair, resulting in 4-5 pTRM-checks per
143 specimen. To detect multidomain effects, the laboratory field was applied in angles of 45°-
144 90° to the ChRM of the specimen (Yu and Tauxe, 2005). The laboratory field itself was varied
145 between 3 and 30 μT between experiments to detect possible field dependencies of the
146 results.

147 Thermal Thellier experiments were undertaken using the IZZI protocol with pTRM checks in
148 vacuum, using an MMTD80 oven with a bias field of 10 μT in the temperature range of 100°C-
149 550°C. Magnetisations of the specimens were measured on an AGICO JR6 spinner
150 magnetometer after being 'cleaned' with a 2 mT AF step before every measurement using an
151 AGICO LDA5 AF demagnetiser.

152 The double-heating Shaw method of (Tsunakawa and Shaw, 1994) used AF demagnetisations
153 and ARM acquisitions carried out on a 2G RAPID superconducting rock magnetometer system
154 in incremental steps of 2-10 mT up to a peak AF-field of 100 mT. ARMs were given at peak AF-
155 field in bias fields of 57.9 μT or 81.2 μT . The specimens were given a full TRM twice by heating
156 them to 610° or 650°C in bias fields of 5 μT or 10 μT to be able to apply an ARM correction
157 (Rolph and Shaw, 1985).

158 The pseudo-Thellier method (Paterson et al., 2016; Tauxe et al., 1995) was performed on the
159 automatic 2G RAPID system with the same AF levels as used in the Shaw method. This was
160 done to have results that avoided heating the specimen at all during palaeointensity
161 determination but its results were only used to confirm the results from the other methods.
162 For the purpose of this experiment, bias fields of 11.4 μT and 81.2 μT were used for the ARM

163 acquisition and, in some cases, Shaw and pseudo-Thellier experiments were run as a
164 combined experiment, where the ARM acquisition steps were performed between the NRM0
165 and ARM0 steps of the Shaw experiments.

166 Different sets of selection criteria have been used to assess the quality of the palaeointensity
167 experiments using the different methods.

168 For all Thellier-type experiments, the selection criteria followed the Standard Palaeointensity
169 Definition (SPD, (Paterson et al., 2014) and were modified from the SELCRIT2 criteria (Biggin
170 et al., 2007). The criteria of $N=4$, $FRAC \geq 0.35$, $\beta \leq 0.1$, $q \geq 1$, $MAD \leq 15^\circ$, $\alpha \leq 15^\circ$, $DRAT \leq 10\%$
171 and $CDRAT \leq 10\%$ were used together with a curvature factor if the best-fit line (Paterson,
172 2011) of $|k'| \leq 0.48$.

173 The selection criteria used for the Shaw method were similar to those set out by Yamamoto
174 et al., (2010),: number of consecutive data points $N \geq 5$, correlation coefficients of the linear
175 parts of the NRM-TRM1* diagram and the TRM1-TRM2* diagram of $r^2_N \geq 0.995$ and $r^2_T \geq$
176 0.995 , the fraction of used NRM $f \geq 0.2$ and the slope of the linear part of the TRM1-TRM2*
177 diagram $0.95 \leq slope_T \leq 1.05$. In addition, the selected part of the NRM must appear
178 convergent on the orthographic plot of the NRM demagnetisation with α and MAD values \leq
179 10° .

180 The SPD selection criteria used for pseudo-Thellier experiments were slightly relaxed from
181 Paterson et al., (2016) and applied to a convergent part of NRM in the orthographic plot: $N \geq$
182 6 , $f \geq 0.3$, $\beta \leq 0.1$, $r^2_{corr} \geq 0.990$, $f_{resid} \leq 0.15$, $\alpha \leq 10^\circ$, $MAD \leq 10^\circ$, $|k'| \leq 0.27$ and $0.85 \leq |b_{AA}|$
183 ≤ 1.15 .

184 Results that passed all criteria were classified as ‘class A’ result. We found that by relaxing a
185 single criterion to a less strict value, we could dramatically increase success rates without
186 severely affecting the overall result quality. Results that required a single criterion to be
187 relaxed were classified as ‘class B’ results. Limits for relaxed selection criteria were: $FRAC \geq$
188 0.25 , $DRAT \leq 15\%$ or $CDRAT \leq 15\%$ for Thellier-type experiments and $r^2_N \geq 0.990$, $r^2_T \geq 0.990$
189 or $0.90 \leq slope_T \leq 1.10$ for Shaw experiments. If more than one criterion would have to be
190 relaxed for a result to pass, the estimate was rejected.

191 **3 Results**

192 Thermomagnetic curves and susceptibility versus temperature curves measured on samples
193 from different dykes showed quite diverse behaviour (supplementary figure 1). Most
194 specimens showed one or two Curie temperatures between 550°C and 580°C. Specimens
195 from the French River and Key River dykes as well as specimens from sites GD25 and GD26
196 also had noticeable Hopkinson peaks, indicating fine magnetite. Exceptions were
197 measurements on specimens from site GD01 with a Curie temperature of 470°C and from
198 sites GD07 and GD30, where susceptibilities were about two orders of magnitude weaker.
199 With the exception of these specimens and those from Coniston dyke (GD01) and site GD33
200 of Augusta Lake dyke, thermomagnetic curves were mostly reversible. The susceptibility
201 curves of specimens from Coniston (GD01, 14), French River (GD23) and Sand Bay (GD29)
202 dykes showed a small ‘toe’ with another Curie temperature up to 610-620°C, possibly
203 suggesting the presence of a small fraction of maghemite or titanohematite. Backscattered
204 electron imaging (figure 2a-c) shows mostly relatively fresh coarse (~30-200 μm) (Ti-
205)magnetite grains with sharp edges as well as small dendritic (< ~40 μm) TM grains (e.g. figure
206 2a,b), indicating rapid cooling of the rocks. EDS elemental composition mapping of larger

207 grains shows structures of full and partial exsolutions with lamellae of titanium rich ilmenite
208 and titanium poor magnetite (figure 2c, supplementary figure 2). The preservation of high
209 temperature textures and lack of low-temperature oxidation structures suggest that the
210 grains are carriers of a primary TRM. For some samples it cannot be ruled out that the oxy-
211 exsolution continued to temperatures lower than the Curie temperatures, but this has shown
212 to only have small effects on the resulting palaeointensities (Shcherbakov et al., 2019).

213
214 Hysteresis parameters, shown in Figure 2e) and f), fall in two clusters on the Day et al., (1977)
215 plot with one large cluster in the “PSD” area and another cluster with higher Bcr/Bc ratios.
216 Most specimens plot close to the TM0 line (Wang and Van der Voo, 2004). This includes
217 results from the French River dyke, which showed a substantially higher TiO₂ content in the
218 geochemical analysis of the sites in Halls et al. (2015). In contrast, specimens from the low-Ti-
219 type Coniston dyke (Halls et al., 2015) plot further away from the TM0 and closer to the TM60
220 line.

221 Fractions of viscous remanent magnetisations (VRM) of specimens from the different sites,
222 acquired after 3 weeks in the ambient field, did not show any differences at the dyke level.
223 The average viscosity index (v) for all sites (figure 2 d) was 14.4% (median = 9.6%, standard
224 deviation = 14.9%). This mean viscosity index includes v values of up to 80% from sites GD07
225 and GD30. Specimens from those sites were basically non-magnetic and the viscosity index
226 was mostly calculated from noise. Specimens with v values of >30% were not used for
227 intensity determination. We note that low palaeointensities will naturally produce weak
228 NRMs which will then be more radically affected by a VRM acquired in the strong present-day
229 field.

230 Results of palaeointensity experiments are summarised in table 1. The total number of Shaw
231 experiments only accounts for specimens that were retained for the full routine. The
232 experiment was stopped after the first demagnetisation series (NRM0) for 16 specimens
233 whose ChRM could not be isolated. Microwave demagnetisation experiments were
234 performed on 29 sister specimens to determine their response to microwave treatments.
235 Specimens from the same core as specimens that did not respond well to the microwave
236 treatment during demagnetisation, or were too weak to be demagnetised at all, were not
237 used for intensity experiments with the microwave Thellier method. In total, palaeointensity
238 measurement was attempted on 189 specimens of the Grenville dykes. Of these, 128 (68%)
239 used microwave IZZI, 20 (11%) used thermal IZZI, and 41 (21%) used Shaw and/or pseudo-
240 Thellier experiments (Table 1).

241 Specimens universally showed non-ideal behaviour during thermal Thellier experiments.
242 These mostly resulted in chaotic Arai and orthographic plots (figure 3e) and the experiment
243 was stopped at 550°C. Strong ‘zig-zagging’ and high β values suggest strong MD behaviour
244 over the full temperature range. In contrast, pTRM checks passed the criteria up to the highest
245 temperature steps.

246 The majority of Arai plots from all Thellier experiments (75%) and of passing Thellier results
247 (94%) showed two slopes as illustrated in figure 3. Similar to the results of other studies
248 yielding extremely low palaeointensities (Hawkins et al., 2019; Shcherbakova et al., 2020), the
249 sharp bend between the high and low temperature/power slopes usually coincided with the
250 junction between ChRM and secondary components in the associated Zijdeveld plots
251 (figure3 a,c), but this is was not clearly visible for all results (figure 3 b). Of the accepted
252 microwave Thellier type results, 10 passed as class A and 15 as class B result. For all class B

253 results, either DRAT or CDRAT (but not both) had to be relaxed. About 10% of the results that
254 were not accepted failed only one of the selection criteria. All other non-accepted results
255 failed more than one selection criterion (e.g. figure 3e). The most commonly failed criteria
256 were β (57% of results), FRAC (56%) and DRAT/CDRAT (48/52%). The high failure rate due to
257 β and DRAT/CDRAT shows that MD effects and mineral alterations during heatings were both
258 primary issues in the results of thermal and microwave experiments. Of the low
259 temperature/power slopes, 6 results (5%) would pass the selection criteria, but it is clearly
260 visible in the orthographic plot that the selected fraction is not the ChRM (figure 3d). One
261 outlier intensity from a high-power slope of 26.7 μT was rejected as unreliable even though
262 the result passed the selection criteria (GD19-43B1). Here, sister specimens from the same 1"
263 core gave inconsistent intensities ranging from 3.6 to 140 μT .

264

265 Of the 12 accepted Shaw results, 7 passed as class A and 5 as class B results. For all 7 class B
266 results, the correlation coefficient of the best fit line in the NRM-TRM1* plot r^2_{N} was relaxed
267 to $r^2_{\text{N}} \geq 0.990$ (figure 4a). Results not passing r^2_{N} and/or slope $_{\text{T}}$ was the main reason for
268 unsuccessful experiments. Specimens showed a wide range of coercivities with a mean MDF
269 = 19.2 \pm 9.6 mT. Secondary magnetisations that accounted for NRM fractions of up to 80%
270 were usually removed by AF fields of less than 40 mT. Similar to the AF demagnetisation
271 experiments in Halls et al. (2015), that successfully generated reliable directions and passed
272 field tests, the remaining fraction of primary NRM was large enough in the Shaw experiments
273 as well so that only two specimens failed the FRAC criterion.

274 From the 7 pseudo-Thellier and the 21 Shaw/pseudo-Thellier experiments 5 pseudo-Thellier
275 results passed selection criteria and were accepted. Due to the high uncertainty of pseudo-

276

277 Thellier intensities associated with the used calibration factor (Paterson et al., 2016), the
278 resulting palaeointensities were not considered in the calculation of the mean
279 palaeointensities. However, since thermal alteration of the specimens during experiments
280 was one of the main issues in the other methods, the pseudo-Thellier results can be useful to
281 confirm that the results from the other methods are not biased by such alterations. Results
282 from successful pseudo-Thellier experiments showed good agreement with the site mean
283 values calculated from microwave and Shaw results (Table 1) with the mean values of sites
284 GD33, GD29 and GD26 being within the 25% uncertainty range of the associated pseudo-
285 Thellier results. For sites GD26 and GD37 that show a larger difference ($>1 \mu\text{T}$) between
286 intensities from pseudo-Thellier experiments and intensities from other methods, the
287 palaeointensities from pseudo-Thellier experiments are lower than the palaeointensities
288 from the other methods. Most of the unsuccessful pseudo-Thellier experiments failed due to
289 a high curvature of the selected line segment in the demag-demag plot (figure 4b), suggesting
290 that NRM and ARM (de)magnetisations behave differently in these specimens and a
291 palaeointensity calculated from these would be unreliable.

292 Combining class A and class B results, 25 results of the microwave Thellier experiments (20%
293 success rate) and 12 results of the Shaw experiments (35%) passed the used selection criteria.
294 No result of the thermal Thellier experiments passed selection criteria, leading to an overall
295 success rate of 20% for Shaw and Thellier-type experiments. Palaeointensity estimates were
296 averaged for each site as separate spot readings of the field strength (Table 1). The resulting
297 intensity values range from $1.4 \mu\text{T}$ to $7.6 \mu\text{T}$, and yielded virtual dipole moments (VDM) from
298 0.3 to $1.7 \cdot 10^{22} \text{ Am}^2$. Five pseudo-Thellier results passed selection criteria as well and gave

299 intensity values between 1.1 μT and 4.1 μT . All intensity results with critical values are listed
300 in supplementary tables 1-3 and measurement data of all intensity experiments are available
301 on the MagIC database (www2.earthref.org/MagIC).

302 **4 Discussion**

303 The geomagnetic field in the Ediacaran is characterised by ambiguous field directions that
304 impede the construction of reliable APW paths. Recently, it has been stated that the time
305 averaged dipole field strength is uniquely low in the Ediacaran (Bono et al., 2019) which is
306 supported by the only other palaeointensity study of Ediacaran rocks to date (Shcherbakova
307 et al., 2020). We expand the intensity data from these studies with the whole rock
308 palaeointensities of the Grenville dykes with virtual dipole moments between $0.3 \cdot 10^{22}\text{Am}^2$
309 and $1.7 \cdot 10^{22}\text{Am}^2$.

310 Due to the low number of available samples and non-ideal behaviour during thermal Thellier
311 experiments, microwave experiments were preferred as the small sample size allowed for a
312 high number of experiments with sub-specimens. To account for a possible method-bias of
313 the MW results, a number of thermal Thellier, Shaw and pseudo Thellier experiments were
314 conducted as well. All methods suffered from low success rates that made it hard to compare
315 the different methods on the site level. Where such a comparison was possible, the results
316 from different methods differed by less than 1 μT , similar to the overall intensity variability
317 on the dyke level. From this, a method related systematic bias seems unlikely. Furthermore,
318 results from Thellier and Shaw experiments are also supported by the results of the pseudo-
319 Thellier experiments.

320 To confirm that the intensity results are not biased by the choice of selection criteria, other
321 commonly used sets of selection criteria have been used to re-analyse all Thellier-type results.

322 Accepted results from the different sets of criteria have been compared (supplementary table
323 4). All sets show only small differences in the range of $\sim 1 \mu\text{T}$ and agree well with the results
324 of the Shaw experiments. Supplementary table 4 also shows that results from combining class
325 A and class B results only differ from pure class A results by $0.2 \mu\text{T}$ and are well inside the
326 standard deviation of class A results, while the number of accepted results almost doubled.

327 The two-slope behaviour, seen in most Arai plots of Thellier experiments in this study, has
328 been recognised in other studies of basalts with low palaeointensities as well (Hawkins et al.,
329 2019; Kodama et al., 2019). Similar behaviour is often connected to MD behaviour (Riisager
330 and Riisager, 2001; Smirnov et al., 2017) and/or alteration of magnetic minerals during
331 heating steps of the experiments (Kissel and Laj, 2004) as well as to instabilities of aged
332 thermoremanence (Shaar and Tauxe, 2015). Thermochemical alteration during the heatings
333 was very common in the experiments of this study and results that were affected by
334 alteration, as seen from pTRM checks, were not accepted. Curved or zig-zagging Arai plots,
335 caused by MD behaviour and/or instability of TRM have been excluded as well if the curvature
336 or β criteria were not met. Hawkins et al. (2019) argued that the two-slope behaviour without
337 a corresponding directional change in their study was attributed to strong thermal or
338 thermoviscous overprints of similar direction. This is also the case in our study, but here a
339 directional change is often observable at the bend between the two slopes in the Arai plots
340 as well. Therefore, the slope of the high unblocking temperature ranges, that also carried the
341 ChRM in Halls et al. (2015), have been selected in the Arai plots to calculate palaeointensities.
342 The overall reliability of palaeointensity results was quantified by the application of Q_{PI} criteria
343 (Biggin and Paterson, 2014) to each of the sites (see supplementary table 5).

344 The AGE criterion was met by all sites in the dykes dated in Halls et al. (2015). Sites GD33 from
345 Augusta Lake dyke and site GD29 from Sand bay dyke were the only sites to meet the STAT
346 criterion. The site GD23 from French River dyke has more than the minimum of 5 individual
347 intensity estimates, but the estimate had a dispersion (standard deviation/mean) $\leq 25\%$
348 (Paterson et al., 2010), which is often the case for ultra-low palaeointensity results (e.g.
349 Hawkins et al., 2019; Shcherbakova et al., 2020). All other sites yielded less than the 5
350 successful palaeointensity results needed to meet the STAT criterion.

351 Mostly fresh looking titanomagnetite grains with high temperature oxy-exsolution structures
352 and rapidly cooled dendritic magnetite grains without signs of low-temperature oxidations as
353 seen from SEM suggest that the magnetisation is a primary TRM. However, concerns can be
354 raised about the validity of the magnetisation itself due to the anomalous nature of the
355 palaeodirections (Halls et al., 2015). This is especially the case for sites associated with the
356 shallow directional component where the possibility of a remagnetisation event cannot be
357 completely dismissed due to inconsistent results of a baked contact test (Robert et al., 2017).
358 Halls et al. (2015) argued that these directions are also primary because they are almost
359 antipodal and can be followed continuously along the dyke. Palaeointensity estimates,
360 irrespective of the associated directions were similar and extremely low and the angular
361 dispersions of VGPs show almost identical behaviour of steep and shallow components. On
362 the basis of the combined microscopic and palaeomagnetic evidence, we chose to award the
363 TRM criterion to all sites.

364 The use of pTRM checks and the application of the IZZI protocol with β and k' criteria as well
365 as the double heating checks of the Shaw experiments enabled the exclusion of all estimates
366 that could be significantly biased by thermochemical alteration or MD behaviour during the

367 experiments. Therefore, all sites passed both the ALT and MD criteria. Following the standard
368 palaeointensity definitions 1.1 (Paterson et al., 2014), the angle between the laboratory field
369 and the last pTRM check of the Thellier experiments was calculated as $\gamma = 3.2^{\circ} \pm 1.2^{\circ}$, showing
370 that the results were not majorly influenced by anisotropy effects. Systematic bias of
371 estimates due to non-linear TRM behaviour was avoided by using different laboratory TRM
372 fields between 3 μT and 30 μT and ARM fields between 10 μT and 81.2 μT . The similarity of
373 results between the Shaw method with cooling times of $\sim 1\text{h}$ and the microwave Thellier
374 method with cooling times in the order of a fraction of a minute exclude a cooling rate bias.
375 Therefore, the ACN criterion was given to all sites.

376 Four different methods were used to determine palaeointensities. To pass the TECH criterion,
377 the palaeointensity estimate of a unit has to comprise results from at least two different
378 methods and this was the case for 4 sites (supplementary table 5). The criterion LITH was only
379 awarded to site GD37 since it combines results from both the dyke and baked host rocks. It
380 was not met by any of the other sites as all intensity estimates came from the same lithology.
381 Results of all intensity experiments with critical values are available in supplementary tables
382 1-3. In addition, all intensity measurement data is available on the MagIC database, which
383 awards the MAG criterion to all units.

384 Summing up the QPI criteria results in scores of 6-8, showing that the palaeointensity
385 estimates in this study are of high quality. This allows the palaeointensities associated with
386 the steep B and shallow C+D directions found in Halls et al. (2015) to be compared. Averaging
387 site mean palaeointensities of all sites with shallow directions results in $3.7 \pm 2.3 \mu\text{T}$ whereas
388 the average palaeointensity of sites with steep directions gives $5.0 \pm 0.5 \mu\text{T}$. Assuming a high-
389 latitude Laurentia around $\sim 590 \text{ Ma}$, the palaeointensities of the shallow component – if

390 recording an equatorial dipole field state – might be expected to be much lower than the
391 intensities of the steep component. The average intensities of sites with the shallow
392 component are indeed weaker, but with the higher values being within one standard
393 deviation of the weaker values, the difference of $< 2 \mu\text{T}$ is negligible. The similarity of the
394 ultra-low palaeointensities of the two components suggests that the Grenville Dykes have
395 recorded a highly unstable field as proposed in Halls et al. (2015). In contrast, the high and
396 almost identical values for angular dispersions (supplementary figure 3) of two distinct groups
397 of VGPs around the mean VGPs of the steep ($S_B=18.5^\circ$) and the shallow component ($S_B=18.9^\circ$)
398 look consistent with the existence of an equatorial dipole. However, some caution is advised
399 when taking the dispersions at face value. Previous studies of VGP scatter required a
400 minimum of $N = 9$ sites (Dobrovine et al., 2019; Veikkolainen and Pesonen, 2014). Due to
401 the exclusion of lower quality directions from sites with $n \leq 4$ or $k \leq 30$ in the calculations, this
402 requirement was only met by the shallow component ($N = 11$), but not by the steep ($N = 7$).
403 The dispersion values seem reasonable for the time period with comparable values for
404 Laurentia showing a wide range between $S_B = 13.5^\circ$ (Skinner Cove volcanics, McCausland and
405 Hodych, 1998; Veikkolainen and Pesonen, 2014) and $S = \sim 26^\circ$ (Sept-Îles intrusion, Bono et al.,
406 2019). If the two groups of VGPs were interpreted as one group, showing a transitional field,
407 then the resulting VGP dispersion would be $S_B = 32.4^\circ$ (variable cutoff = 63.4°) at low latitude.
408 This would be an extremely high value but the current lack of constraints on the Ediacaran
409 field means that it is not implausible.

410 With the exception of sites GD14 and GD25, where single high (7.6 and 12.6 μT) microwave
411 results lead to high VDMs of $1.7 \cdot 10^{22}\text{Am}^2$, the site mean VDMs of $0.3 - 0.9 \cdot 10^{22}\text{Am}^2$ are
412 comparable to, but lower than, the results of the Sept Iles (Bono et al., 2019) and the Volyn
413 Traps (Shcherbakova et al., 2020) with corresponding VDMs of 0.5 and $1.0 \cdot 10^{22}\text{Am}^2$. They

414 therefore define a new lower boundary for the field strength in the Ediacaran. The
415 palaeointensities, coming from dykes with ages that span ~15 Ma, suggest a sustained
416 geomagnetic field with these extremely low intensities. These results are an order of
417 magnitude weaker than the strength of the present-day field and the average
418 palaeointensities of site GD26 and sites from the French River dyke and the Augusta Lake dyke
419 with values between 1.4 and 2.1 μT are even as low as recent measurements of Mars' crustal
420 field made by the InSight lander of ~2 μT (Johnson et al., 2020). Palaeointensities this low
421 have been reported for Earth before, but were generally not attributed to a sustained field.
422 The PINT database (Biggin et al., 2015) contains 6 site mean estimates with $H_{\text{pal}} \leq 5 \mu\text{T}$ or VDM
423 $\leq 0.5 * 10^{22} \text{Am}^2$, $N > 1$ and reliable experiment types (excluding single-heating Shaw and total
424 TRM experiments) that can be roughly divided into two groups. The first group comprises
425 entries with mid-Miocene or younger ages that show the ultra-low intensity values in single
426 basalt flows from Iceland (Lawley, 1970) and the Canary Islands (Brown et al., 2009; Leonhardt
427 and Soffel, 2002) that are all connected to the short-term drop of dipole moments during
428 polarity transitions interrupting a much stronger sustained field. The palaeointensities with
429 Mesoproterozoic to Archaean ages from the second group are either extremely low due to
430 fractions of CRM (Yoshihara and Hamano, 2004), would not satisfy any modern sets of
431 selection criteria (Ueno, 1995), or are only seen as spot reading of the field in a single dyke
432 (Smirnov and Tarduno, 2005) in a dyke swarm showing a weak, but overall stronger field (Halls
433 et al., 2004). Non-Ediacaran weak sustained fields as in the Jurassic (e.g. ~2.8 $* 10^{22} \text{Am}^2$, Tauxe
434 et al., 2013) or the Devonian (e.g. ~1.1 $* 10^{22} \text{Am}^2$, Hawkins et al., 2019) are all stronger.
435 However, a field strength behaviour, similar to the one in the Ediacaran, can be observed in
436 the Upper Devonian around ~370 Ma, where site mean palaeointensities as low as 2.4 μT (0.4
437 $* 10^{22} \text{Am}^2$) have been reported in the weak time averaged field as well (Hawkins et al., 2019).

438 The ultra-low Ediacaran intensities are consistent with the predicted weak field state of the
439 geodynamo before the onset of inner core growth (Bono et al., 2019; Driscoll, 2016; Landeau
440 et al., 2017). Under this scenario, the dynamo was operating marginally, powered by thermal
441 convection due to heat loss at the core-mantle boundary alone and the field was diminished.
442 Subsequently, the inner core nucleated providing additional convective power from the
443 release of light elements and latent heat of crystallisation. We cannot, however, rule out an
444 entirely different cause of a massively diminished dipole moment in the Ediacaran, perhaps
445 related to a reconfigured convective pattern in the core perhaps related to an unusual core-
446 mantle heat flow pattern. That the measured palaeointensities are similar in magnitude to
447 ground measurements of the crustal field of Mars, a planet which is suspected to have been
448 without a core dynamo for 4.1 billion years, raises profound questions concerning
449 measurement limits and the history of the geodynamo. Is it even possible to measure
450 geomagnetic-field-derived palaeointensities that are any lower than the values presented
451 here, given the possibility of contamination by a (palaeo-) crustal field? Can we therefore be
452 certain that the geodynamo did not die out altogether in the Ediacaran? Such questions are
453 unfortunately beyond the scope of the present study to answer but motivate urgent future
454 investigations that may yield transformative insights into our planet's history.

455

456 **5 Conclusions**

457 We report new high quality palaeointensities from eleven sites of the western end of the
458 Grenville dyke swarm. Success rates of palaeointensity experiments were substantially
459 lowered by thermochemical alteration during laboratory heating as well as MD behaviour of
460 the studied rocks. The resulting palaeointensity values range between 1.4 - 7.6 μT (0.33 – 1.76

461 * 10^{22} Am²). Most of these estimates agree well with results from other palaeointensity studies
462 from this time period (Bono et al., 2019; Shcherbakova et al., 2020) but several
463 palaeointensities are substantially lower with values being similar to values of Mars' crustal
464 field (Johnson et al., 2020). This opens up questions about what geodynamo regimes could
465 result in such weak fields or, to take it one step further, if such weak fields could even suggest
466 an inactive geodynamo. The behaviour of VGPs with almost identical angular dispersion
467 around two clusters argues for the presence of an equatorial dipole field. In contrast, the
468 consistency of presented ultra-low intensities along dykes for both directional components
469 supports the idea of an unstable and/or transitional field in the Ediacaran (Bono et al., 2019;
470 Halls et al., 2015). The presented data are another argument for the Ediacaran field behaving
471 strangely but at this point we are still struggling to define it. However, data for the early to
472 mid-Ediacaran is still scarce and a better coverage of palaeointensity and
473 magnetostratigraphic data would be immensely useful to better characterise the
474 geomagnetic field in the Ediacaran.

475

476 **Acknowledgements**

477 This study was supported by The Leverhulme Trust (Research Leadership Award, RL-2016-
478 080). We would like to thank R.K. Bono, G. Paterson and S. Lloyd for helpful discussions. Henry
479 Halls would like to thank Alan Lovette for helpful field assistance in finding outcrops of dykes
480 that formed negative topographic features and for many of the paleomagnetic
481 measurements.

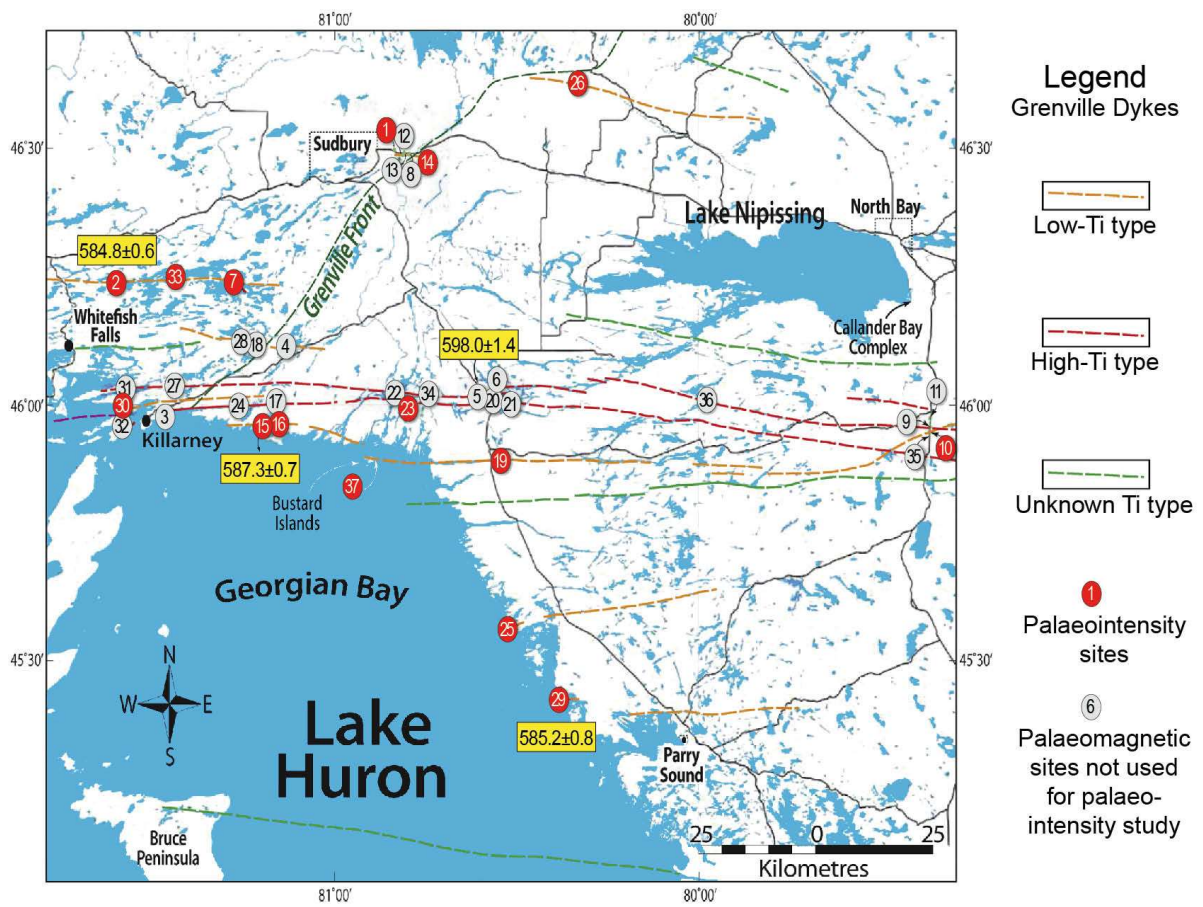
482

483 **Data availability:**

484 *Data has been uploaded to a private repository in the MagIC database and will be accessible*
485 *after publication. Currently, data from this study are available from the corresponding author*
486 *on reasonable request.*

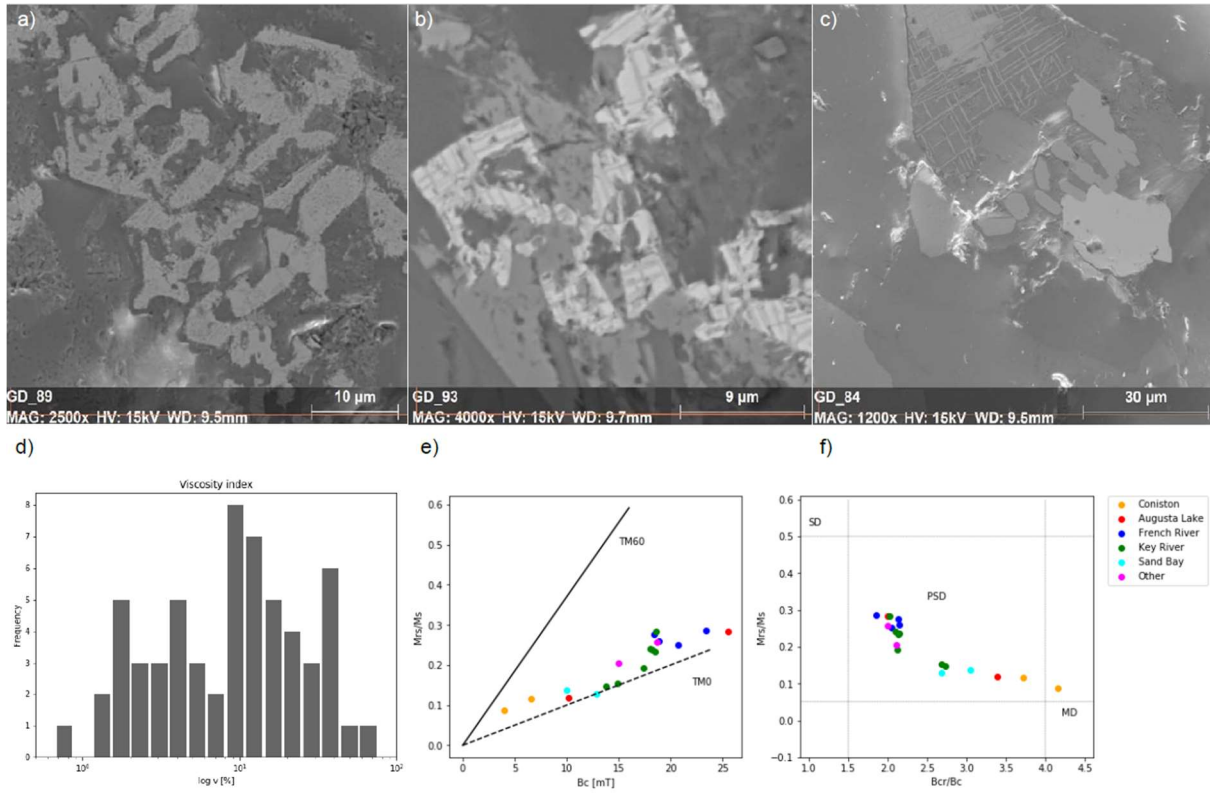
487

488 **Figures:**



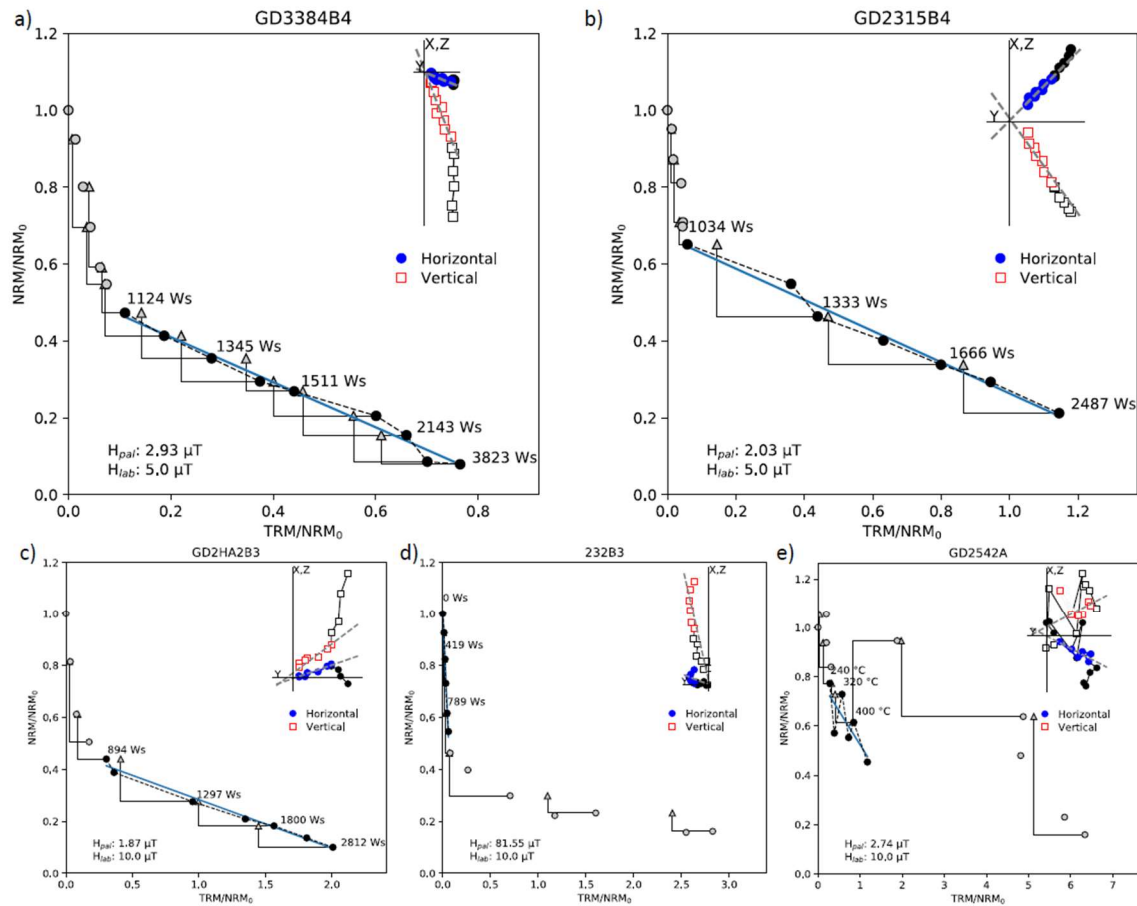
489

490 **Figure 1:** Map of southeast Ontario, showing the Grenville dykes and locations of sites studied
491 in Halls et al. (2015) and sites selected for this palaeointensity study; from Halls et al. (2015)



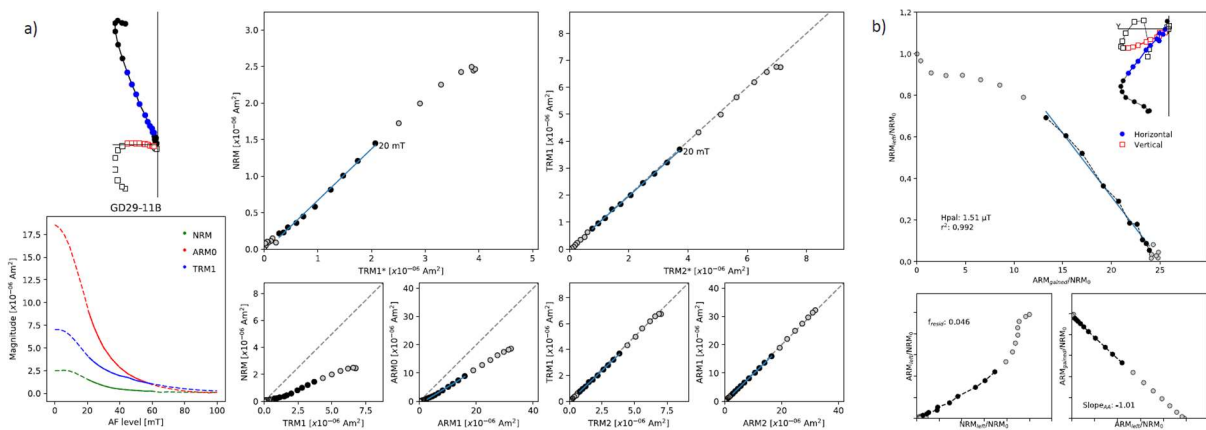
492

493 **Figure 2:** SEM backscatter images: (a) GD23, (b) GD29, examples of dendritic TM grains, (c)
494 GD02, coarse Tm grain, showing magnetite-ilmenite exsolution. (d) frequencies of viscosity
495 indexes of all sites, (e) Hysteresis parameters and (f) Day plot of representative samples of
496 each site.



497

498 **Figure 3:** Arai plots of Thellier experiment results. a)-d) show microwave experiments, e)
 499 shows a result of a thermal Thellier experiment. For more details, readers are referred to
 500 section 3 of the text.



501

502 **Figure 4:** a) Representative example of a successful Shaw experiment: left column shows the
 503 orthogonal plot of the demagnetisation of NRM (top) and absolute values of magnetisation

504 during AF demagnetisations of NRM, ARM0 and TRM1 (bottom). middle and right columns
 505 show results from first and second laboratory heating, respectively; b) example of a successful
 506 pseudo-Thellier experiment: pseudo-Arai plot with orthogonal projection, demag-demag plot
 507 (NRM vs ARM) and ARM-ARM plot (acquisition vs. demagnetisation).

508

509 **Tables:**

Dyke	Site	Age [Ma]	Directions (Halls et al 2015)					Intensities [μ T]				$pTheI$ [μ T]	VDM [10^{22} Am ²]		QPI
			N/n	Dec	Inc	a95	k	n/n _{PI}	Methods	PI	Std	PI	VDM	Std	
Augusta Lake	GD02	584	9/8	120.7	23.1	6.7	69	12/2	MW+S	2.1	0.04	-	0.52	0.01	7
	GD33	± 0.6	8/6	135.6	31.2	8.8	60	17/5	MW	2.6	0.62	1.9	0.60	0.14	7
Coniston	GD01		5/3	115.5	17.1	32.4	16	4/0	-	-	-	-	-	-	-
	GD14		13/3	95.5	23.8	26.3	13	6/1	MW	7.6	-	-	1.76	-	6
French River	GD23	598 ± 1.4	6/3	132.7	32.6	8.7	202	24/13	MW+S	1.8	0.67	-	0.42	0.15	7
Key River	GD10		10/9	137.7	75.8	6.7	61	14/0	-	-	-	-	-	-	-
	GD15		5/5	140.9	66.8	6.7	130	12/1	MW	5.6	-	-	0.88	-	6
	GD16		6/6	163.8	76.7	14.4	22	10/1	MW	4.3	-	1.3	0.67	-	6
	GD19	587	9/5	100.6	56.2	11.9	42	11/0	-	-	-	-	-	-	-
Sand Bay	GD37	± 0.7	8/7	143.1	71.7	5.1	141	6/3	S	5.3	1.57	3.1	0.78	0.23	7
	GD29	585 ± 0.8	9/9	297	40.1	5.2	97	41/7	MW+S	3.7	0.69	4.1	0.80	0.15	8
Other Grenville Dykes	GD07		7/6	134.8	1.2	11.2	37	1/0	-	-	-	-	-	-	-
	GD25		9/9	136.5	9.6	8.8	36	14/2	MW+S	6.8	5.82	-	1.74	1.41	7
	GD26		9/6	130.1	23.8	10.1	45	7/2	S	1.4	0.40	1.5	0.33	0.10	6
	GD30		10/6	139	67.3	8.4	64	3/0	-	-	-	-	-	-	-

510

511 **Table 1:** Summary of palaeointensity results of the Grenville dykes: Dyke/Site: dyke/site name, Ages
 512 and directional information from Halls et al. (2015), N/n_{int}: total number of specimens/ number of
 513 successful results, Methods: type of method that contributed successful results (MW: microwave
 514 Thellier, TH: thermal Thellier, S: Shaw), PI: palaeointensity results in μ T, Std: standard deviation of
 515 palaeointensity results (*: standard error of single specimen result instead of multiple specimen

516 result), pTheI: palaeointensity results in μT from pseudo-Thellier experiments. Values are shown as
517 comparison to the results from the heating methods but are not used in the calculation of the mean
518 palaeointensities. VDM: virtual dipole moment in 10^{22}Am^2 , Std: standard deviation of VDM results,
519 QPI values of sites.

520

521 **References:**

522 Abrajevitch, A., Van der Voo Rob, R., 2010. Incompatible Ediacaran paleomagnetic directions
523 suggest an equatorial geomagnetic dipole hypothesis. Earth Planet. Sci. Lett.

524 <https://doi.org/10.1016/j.epsl.2010.02.038>

525 Aubert, J., Wicht, J., 2004. Axial vs. equatorial dipolar dynamo models with implications for
526 planetary magnetic fields. Earth Planet. Sci. Lett. <https://doi.org/10.1016/S0012->

527 [821X\(04\)00102-5](https://doi.org/10.1016/S0012-821X(04)00102-5)

528 Bazhenov, M.L., Levashova, N.M., Meert, J.G., Golovanova, I. V., Danukalov, K.N., Fedorova,
529 N.M., 2016. Late Ediacaran magnetostratigraphy of Baltica: Evidence for Magnetic Field
530 Hyperactivity? Earth Planet. Sci. Lett. <https://doi.org/10.1016/j.epsl.2015.12.015>

531 Biggin, A.J., Paterson, G.A., 2014. A new set of qualitative reliability criteria to aid inferences
532 on palaeomagnetic dipole moment variations through geological time. Front. Earth Sci.

533 <https://doi.org/10.3389/feart.2014.00024>

534 Biggin, A.J., Perrin, M., Dekkers, M.J., 2007. A reliable absolute palaeointensity
535 determination obtained from a non-ideal recorder. Earth Planet. Sci. Lett.

536 <https://doi.org/10.1016/j.epsl.2007.03.017>

537 Biggin, A.J., Piispa, E.J., Pesonen, L.J., Holme, R., Paterson, G.A., Veikkolainen, T., Tauxe, L.,

- 538 2015. Palaeomagnetic field intensity variations suggest Mesoproterozoic inner-core
539 nucleation. *Nature*. <https://doi.org/10.1038/nature15523>
- 540 Bono, R.K., Tarduno, J.A., 2014. A stable ediacaran earth recorded by single silicate crystals
541 of the ca. 565 Ma sept-Îles intrusion. *Geology*. <https://doi.org/10.1130/G36247.1>
- 542 Bono, R.K., Tarduno, J.A., Nimmo, F., Cottrell, R.D., 2019. Young inner core inferred from
543 Ediacaran ultra-low geomagnetic field intensity. *Nat. Geosci.*
544 <https://doi.org/10.1038/s41561-018-0288-0>
- 545 Brown, M.C., Gratton, M.N., Shaw, J., Holme, R., Soler, V., 2009. Microwave palaeointensity
546 results from the Matuyama-Brunhes geomagnetic field reversal. *Phys. Earth Planet.*
547 *Inter.* <https://doi.org/10.1016/j.pepi.2008.11.001>
- 548 Day, R., Fuller, M., Schmidt, V.A., 1977. Hysteresis properties of titanomagnetites: Grain-size
549 and compositional dependence. *Phys. Earth Planet. Inter.*
550 [https://doi.org/10.1016/0031-9201\(77\)90108-X](https://doi.org/10.1016/0031-9201(77)90108-X)
- 551 Doubrovine, P. V., Veikkolainen, T., Pesonen, L.J., Piispa, E., Ots, S., Smirnov, A. V., Kulakov,
552 E. V., Biggin, A.J., 2019. Latitude Dependence of Geomagnetic Paleosecular Variation
553 and its Relation to the Frequency of Magnetic Reversals: Observations From the
554 Cretaceous and Jurassic. *Geochemistry, Geophys. Geosystems*.
555 <https://doi.org/10.1029/2018GC007863>
- 556 Driscoll, P.E., 2016. Simulating 2 Ga of geodynamo history. *Geophys. Res. Lett.*
557 <https://doi.org/10.1002/2016GL068858>
- 558 Evans, D.A.D., Raub, T.D., 2011. Chapter 7: Neoproterozoic glacial palaeolatitudes: A global
559 update. *Geol. Soc. Mem.* <https://doi.org/10.1144/M36.7>

- 560 Gissinger, C., Petitdemange, L., Schrunner, M., Dormy, E., 2012. Bistability between
561 equatorial and axial dipoles during magnetic field reversals. *Phys. Rev. Lett.*
562 <https://doi.org/10.1103/PhysRevLett.108.234501>
- 563 Halls, H.C., Lovette, A., Hamilton, M., Söderlund, U., 2015. A paleomagnetic and U-Pb
564 geochronology study of the western end of the Grenville dyke swarm: Rapid changes in
565 paleomagnetic field direction at ca. 585Ma related to polarity reversals? *Precambrian*
566 *Res.* <https://doi.org/10.1016/j.precamres.2014.11.029>
- 567 Halls, H.C., McArdle, N.J., Gratton, M.N., Hill, M.J., Shaw, J., 2004. Microwave
568 paleointensities from dyke chilled margins: A way to obtain long-term variations in
569 geodynamo intensity for the last three billion years. *Phys. Earth Planet. Inter.*
570 <https://doi.org/10.1016/j.pepi.2004.03.013>
- 571 Hawkins, L.M.A., Anwar, T., Shcherbakova, V. V., Biggin, A.J., Kravchinsky, V.A., Shatsillo, A.
572 V., Pavlov, V.E., 2019. An exceptionally weak Devonian geomagnetic field recorded by
573 the Viluy Traps, Siberia. *Earth Planet. Sci. Lett.*
574 <https://doi.org/10.1016/j.epsl.2018.10.035>
- 575 Hill, M.J., Shaw, J., 1999. Palaeointensity results for historic lavas from Mt Etna using
576 microwave demagnetization/remagnetization in a modified Thellier-type experiment.
577 *Geophys. J. Int.* <https://doi.org/10.1046/j.1365-246X.1999.00980.x>
- 578 Hyodo, H., Dunlop, D.J., 1993. Effect of anisotropy on the paleomagnetic contact test for a
579 Grenville Dike. *J. Geophys. Res.* <https://doi.org/10.1029/92JB02915>
- 580 Johnson, C.L., Mittelholz, A., Langlais, B., Russell, C.T., Ansan, V., Banfield, D., Chi, P.J.,
581 Fillingim, M.O., Forget, F., Haviland, H.F., Golombek, M., Joy, S., Lognonné, P., Liu, X.,

- 582 Michaut, C., Pan, L., Quantin-Nataf, C., Spiga, A., Stanley, S., Thorne, S.N., Wieczorek,
583 M.A., Yu, Y., Smrekar, S.E., Banerdt, W.B., 2020. Crustal and time-varying magnetic
584 fields at the InSight landing site on Mars. Nat. Geosci. [https://doi.org/10.1038/s41561-](https://doi.org/10.1038/s41561-020-0537-x)
585 [020-0537-x](https://doi.org/10.1038/s41561-020-0537-x)
- 586 Kirschvink, J.L., Ripperdan, R.L., Evans, D.A., 1997. Evidence for a large-scale reorganization
587 of Early Cambrian continental masses by inertial interchange true polar wander.
588 Science (80-.). <https://doi.org/10.1126/science.277.5325.541>
- 589 Kissel, C., Laj, C., 2004. Improvements in procedure and paleointensity selection criteria
590 (PICRIT-03) for Thellier and Thellier determinations: Application to Hawaiian basaltic
591 long cores. Phys. Earth Planet. Inter. <https://doi.org/10.1016/j.pepi.2004.06.010>
- 592 Kodama, K.P., Carnes, L.K., Tarduno, J.A., Berti, C., 2019. Palaeointensity of the 1.3 billion-yr-
593 old Gardar basalts, southern Greenland revisited: No evidence for onset of inner core
594 growth. Geophys. J. Int. <https://doi.org/10.1093/gji/ggz126>
- 595 Kulakov, E. V., Sprain, C.J., Doubrovine, P. V., Smirnov, A. V., Paterson, G.A., Hawkins, L.,
596 Fairchild, L., Piispa, E.J., Biggin, A.J., 2019. Analysis of an Updated Paleointensity
597 Database (QPI-PINT) for 65–200 Ma: Implications for the Long-Term History of Dipole
598 Moment Through the Mesozoic. J. Geophys. Res. Solid Earth.
599 <https://doi.org/10.1029/2018JB017287>
- 600 Landeau, M., Aubert, J., Olson, P., 2017. The signature of inner-core nucleation on the
601 geodynamo. Earth Planet. Sci. Lett. <https://doi.org/10.1016/j.epsl.2017.02.004>
- 602 Lawley, E.A., 1970. The intensity of the geomagnetic field in Iceland during neogene polarity
603 transitions and systematic deviations. Earth Planet. Sci. Lett.

- 604 [https://doi.org/10.1016/0012-821X\(70\)90076-2](https://doi.org/10.1016/0012-821X(70)90076-2)
- 605 Leonhardt, R., Soffel, H.C., 2002. A reversal of the Earth's magnetic field recorded in mid-
606 Miocene lava flows of Gran Canaria: Paleointensities. *J. Geophys. Res. Solid Earth*.
607 <https://doi.org/10.1029/2001jb000949>
- 608 Li, Z.X., Evans, D.A.D., Halverson, G.P., 2013. Neoproterozoic glaciations in a revised global
609 palaeogeography from the breakup of Rodinia to the assembly of Gondwanaland.
610 *Sediment. Geol.* <https://doi.org/10.1016/j.sedgeo.2013.05.016>
- 611 McCausland, P.J.A., Hodych, J.P., 1998. Paleomagnetism of the 550 Ma Skinner Cove
612 volcanics of western Newfoundland and the opening of the Iapetus Ocean. *Earth
613 Planet. Sci. Lett.* [https://doi.org/10.1016/S0012-821X\(98\)00171-X](https://doi.org/10.1016/S0012-821X(98)00171-X)
- 614 McCausland, P.J.A., Van der Voo, R., Hall, C.M., 2007. Circum-Iapetus paleogeography of the
615 Precambrian-Cambrian transition with a new paleomagnetic constraint from Laurentia.
616 *Precambrian Res.* <https://doi.org/10.1016/j.precamres.2007.03.004>
- 617 Meert, J.G., 2014. Ediacaran-Early Ordovician paleomagnetism of Baltica: A review.
618 *Gondwana Res.* <https://doi.org/10.1016/j.gr.2013.02.003>
- 619 Meert, J.G., Van Der Voo, R., Payne, T.W., 1994. Paleomagnetism of the Catoclin volcanic
620 province: a new Vendian- Cambrian apparent polar wander path for North America. *J.
621 Geophys. Res.* <https://doi.org/10.1029/93JB01723>
- 622 Murthy, G.S., 1971. The Paleomagnetism of Diabase Dikes from the Grenville Province. *Can.
623 J. Earth Sci.* <https://doi.org/10.1139/e71-075>
- 624 Paterson, G.A., 2011. A simple test for the presence of multidomain behavior during
625 paleointensity experiments. *J. Geophys. Res. Solid Earth*.

626 <https://doi.org/10.1029/2011JB008369>

627 Paterson, G.A., Heslop, D., Muxworthy, A.R., 2010. Deriving confidence in paleointensity
628 estimates. *Geochemistry, Geophys. Geosystems*.

629 <https://doi.org/10.1029/2010GC003071>

630 Paterson, G.A., Heslop, D., Pan, Y., 2016. The pseudo-Thellier palaeointensity method: New
631 calibration and uncertainty estimates. *Geophys. J. Int.*

632 <https://doi.org/10.1093/gji/ggw349>

633 Paterson, G.A., Tauxe, L., Biggin, A.J., Shaar, R., Jonestrask, L.C., 2014. On improving the
634 selection of Thellier-type paleointensity data. *Geochemistry, Geophys. Geosystems*.

635 <https://doi.org/10.1002/2013GC005135>

636 Prévot, M., 1981. Some aspects of magnetic viscosity in subaerial and submarine volcanic
637 rocks. *Geophys. J. R. Astron. Soc.* <https://doi.org/10.1111/j.1365-246X.1981.tb05952.x>

638 Riisager, P., Riisager, J., 2001. Detecting multidomain magnetic grains in Thellier
639 palaeointensity experiments. *Phys. Earth Planet. Inter.* [https://doi.org/10.1016/S0031-](https://doi.org/10.1016/S0031-9201(01)00236-9)
640 [9201\(01\)00236-9](https://doi.org/10.1016/S0031-9201(01)00236-9)

641 Robert, B., Besse, J., Blein, O., Greff-Lefftz, M., Baudin, T., Lopes, F., Meslouh, S., Belbadaoui,
642 M., 2017. Constraints on the Ediacaran inertial interchange true polar wander
643 hypothesis: A new paleomagnetic study in Morocco (West African Craton).
644 *Precambrian Res.* <https://doi.org/10.1016/j.precamres.2017.04.010>

645 Rolph, T.C., Shaw, J., 1985. A new method of palaeofield magnitude correction for thermally
646 altered samples and its application to Lower Carboniferous lavas. *Geophys. J. R. Astron.*
647 *Soc.* <https://doi.org/10.1111/j.1365-246X.1985.tb05124.x>

- 648 Shaar, R., Tauxe, L., 2015. Instability of thermoremanence and the problem of estimating
649 the ancient geomagnetic field strength from non-single-domain recorders. Proc. Natl.
650 Acad. Sci. U. S. A. <https://doi.org/10.1073/pnas.1507986112>
- 651 Shcherbakov, V.P., Gribov, S.K., Lhuillier, F., Aphinogenova, N.A., Tsel'movich, V.A., 2019. On
652 the Reliability of Absolute Palaeointensity Determinations on Vasaltic Rocks Bearing a
653 Thermochemical Remanence. J. Geophys. Res. Solid Earth.
654 <https://doi.org/10.1029/2019JB017873>
- 655 Shcherbakova, V. V., Bakhmutov, V.G., Thallner, D., Shcherbakov, V.P., Zhidkov, G. V., Biggin,
656 A.J., 2020. Ultra-low palaeointensities from East European Craton, Ukraine support a
657 globally anomalous palaeomagnetic field in the Ediacaran. Geophys. J. Int. 220, 1928–
658 1946. <https://doi.org/10.1093/gji/ggz566>
- 659 Shcherbakova, V. V., Biggin, A.J., Veselovskiy, R. V., Shatsillo, A. V., Hawkins, L.M.A.,
660 Shcherbakov, V.P., Zhidkov, G. V., 2017. Was the Devonian geomagnetic field dipolar or
661 multipolar? Palaeointensity studies of Devonian igneous rocks from the Minusa Basin
662 (Siberia) and the Kola Peninsula dykes, Russia. Geophys. J. Int.
663 <https://doi.org/10.1093/gji/ggx085>
- 664 Smirnov, A. V., Kulakov, E. V., Foucher, M.S., Bristol, K.E., 2017. Intrinsic paleointensity bias
665 and the long-term history of the geodynamo. Sci. Adv.
666 <https://doi.org/10.1126/sciadv.1602306>
- 667 Smirnov, A. V., Tarduno, J.A., 2005. Thermochemical remanent magnetization in
668 Precambrian rocks: Are we sure the geomagnetic field was weak? J. Geophys. Res. Solid
669 Earth. <https://doi.org/10.1029/2004JB003445>

- 670 Tanczyk, E.I., Lapointe, P., Morris, W.A., Schmidt, P.W., 1987. A paleomagnetic study of the
671 layered mafic intrusion at Sept- Iles, Quebec (Canada). Can. J. Earth Sci.
672 <https://doi.org/10.1139/e87-135>
- 673 Tauxe, L., Gee, J.S., Steiner, M.B., Staudigel, H., 2013. Paleointensity results from the
674 Jurassic: New constraints from submarine basaltic glasses of ODP Site 801C.
675 Geochemistry, Geophys. Geosystems. <https://doi.org/10.1002/ggge.20282>
- 676 Tauxe, L., Pick, T., Kok, Y.S., 1995. Relative paleointensity in sediments: A Pseudo-Thellier
677 Approach. Geophys. Res. Lett. <https://doi.org/10.1029/95GL03166>
- 678 Tauxe, L., Staudigel, H., 2004. Strength of the geomagnetic field in the cretaceous normal
679 superchron: New data from submarine basaltic glass of the troodos ophiolite.
680 Geochemistry, Geophys. Geosystems. <https://doi.org/10.1029/2003GC000635>
- 681 Tsunakawa, H., Shaw, J., 1994. The Shaw method of palaeointensity determinations and its
682 application to recent volcanic rocks. Geophys. J. Int. <https://doi.org/10.1111/j.1365->
683 [246X.1994.tb03999.x](https://doi.org/10.1111/j.1365-246X.1994.tb03999.x)
- 684 Ueno, H., Irving, E., McNutt, R.H., 1975. Paleomagnetism of the Whitestone Anorthosite and
685 Diorite, the Grenville Polar Track, and Relative Motions of the Laurentian and Baltic
686 Shields. Can. J. Earth Sci. <https://doi.org/10.1139/e75-019>
- 687 Ueno, N., 1995. Geomagnetic palaeointensity experiment on igneous and metamorphic
688 rocks from Enderby land in Napier Complex, Antarctica. Proc. NIPR Symp. Antarct.
689 Geosci. 8, 193–200.
- 690 Veikkolainen, T., Pesonen, L.J., 2014. Palaeosecular variation, field reversals and the stability
691 of the geodynamo in the Precambrian. Geophys. J. Int.

692 <https://doi.org/10.1093/gji/ggu348>

693 Wang, D., Van der Voo, R., 2004. The hysteresis properties of multidomain magnetite and
694 titanomagnetite/titanomaghemite in mid-ocean ridge basalts. *Earth Planet. Sci. Lett.*

695 [https://doi.org/10.1016/S0012-821X\(04\)00052-4](https://doi.org/10.1016/S0012-821X(04)00052-4)

696 Yamamoto, Y., Shibuya, H., Tanaka, H., Hoshizumi, H., 2010. Geomagnetic paleointensity

697 deduced for the last 300kyr from Unzen Volcano, Japan, and the dipolar nature of the

698 Iceland Basin excursion. *Earth Planet. Sci. Lett.*

699 <https://doi.org/10.1016/j.epsl.2010.02.024>

700 Yoshihara, A., Hamano, Y., 2004. Paleomagnetic constraints on the Archean geomagnetic

701 field intensity obtained from komatiites of the Barberton and Belingwe greenstone

702 belts, South Africa and Zimbabwe. *Precambrian Res.*

703 <https://doi.org/10.1016/j.precamres.2004.01.003>

704 Yu, Y., Tauxe, L., 2005. Testing the IZZI protocol of geomagnetic field intensity

705 determination. *Geochemistry, Geophys. Geosystems.*

706 <https://doi.org/10.1029/2004GC000840>

707