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# Origin of the central magnetic anomaly at the Houghton impact structure, Canada

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## Abstract

The 23 km-diameter well-preserved Houghton impact structure shows a rather unique combination of a positive magnetic anomaly with a negative gravity anomaly over the center of its central uplift. Using a new ground magnetic dataset and several modeling approaches, we investigate the properties and geometry of its central magnetized source. Our results confirm that a km-sized magnetic body with a narrow near-surface extension is necessary to account for the anomaly. Additional measurements of rock magnetic properties of samples of all lithologies encountered in and outside the crater show that the target sedimentary rocks and the vast majority of the Precambrian basement rocks cannot be the source of the magnetic anomaly. While in larger impact structures such magnetic anomalies are often explained by magmatic mafic intrusions or highly magnetic glass lenses in the impact melt rocks, we propose that impact-generated hydrothermal activity enhanced the magnetization of the highly-porous unmelted uplifted basement rocks. Such a

process may be considered for the interpretation of the geophysical signature of planetary impact craters.

*Keywords:* Haughton impact structure, magnetic anomaly, rock magnetism, modeling, hydrothermalism

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## 1. Introduction

Impact cratering processes on planetary surfaces involve excavation and displacement of target formations that lead to a transient cavity (Melosh, 1989). Then, depending on the projectile size, a central uplift of variably shocked and heated target rocks forms, in addition to crater wall collapse through faulting (Grieve, 1987). Interaction of hot impact-affected rocks with surface fluids can subsequently lead to post-impact hydrothermal activity. Magmatism, sedimentation and erosion are also important processes on the Earth and other planets leading to obscure many impact-related features (e.g., Frey et al., 2001; Buczkowski et al., 2005). In such cases, geophysical investigations can provide constraints on the buried structure (e.g., Pilkington and Grieve, 1992; Pilkington and Hildebrand, 2000; Langlais and Thébault, 2011).

The Haughton impact structure, located Devon Island, Nunavut, Canada (Figure 1) is one of the best preserved medium-size (23 km apparent crater diameter) complex impact structures on Earth. The impact occurred 39 Ma ago in a target formation composed of a  $\sim$ 2-km thick sequence of Lower Paleozoic sedimentary rocks of the Arctic Platform overlying Precambrian metamorphic basement of the Canadian Shield (Robertson and Sweeney, 1983; Bischoff and Oskierski, 1988; Hickey et al., 1988; Metzler et al., 1988; Re-

21 deker and Stoeffler, 1988; Osinski et al., 2005b). Clast-rich impact melt rocks  
22 filled the crater and impact-related hydrothermal activity took place (Osinski  
23 et al., 2005a). A large amount of these impactites were eroded or covered by  
24 lacustrine sediments during the Miocene (the Haughton Formation; Figure  
25 1). Later, glacial-interglacial erosion periods modified the landscape, which  
26 is now only affected by seasonal glacial and periglacial processes.

27 Ground magnetic and gravity measurements were carried out within the  
28 central part of the crater by Robertson and Sweeney (1983) and Pohl et al.  
29 (1988). A 24 km diameter negative Bouguer gravity anomaly was discovered  
30 over the crater with a local minimum over the center. The latter coincides  
31 with an intense positive magnetic anomaly. This feature is not common for  
32 impact structures of the same size or smaller (Pilkington and Grieve, 1992;  
33 Henkel et al., 2002; Shah et al., 2005; Ugalde et al., 2007). Using a simple  
34 cone-shape model with a 1-km deep planar root, Pohl et al. (1988) suggested  
35 that this low-density highly magnetic local volume of rock underwent large  
36 degassing increasing the porosity, and acquired a coherent remanent magneti-  
37 zation during cooling after the shock. Robertson and Sweeney (1983) did not  
38 exclude shock remanent magnetization or normal magnetic contrast between  
39 the Precambrian basement rocks and the Paleozoic sedimentary formations.  
40 Both authors modeled the impact structure with a central uplift, which was  
41 partially evidenced by seismic investigations performed on the western flank  
42 of the crater (Scott and Hajnal, 1988). Subsequent structural mapping has  
43 shown that Haughton possesses a morphologically subdued central uplift fea-  
44 ture (Osinski and Spray, 2005). Recent airborne magnetic surveys delivered  
45 additional data that cover the whole crater but no detailed modeling of the

46 central magnetic anomaly has been done (Glass et al., 2012, and references  
47 within). Therefore the exact geometry of the top of this body, as well as its  
48 geological significance, are still unknown.

## 49 **2. Methods**

50 In 2010, we performed a new ground magnetic survey to better character-  
51 ize the central magnetic anomaly of the Haughton impact structure. A Geo-  
52 metrics G858 MagMapper cesium vapor magnetometer was used to measure  
53 the intensity of the geomagnetic field vector (total-field, TF) with a sensor at  
54 1.90 m above the ground. A 1 Hz sampling was used along kilometeric walk-  
55 ing profiles crossing the peak of the magnetic anomaly (located in 424677E,  
56 8367960N, and denoted 'M' on Figure 1 and 2A). Additional discrete mea-  
57 surements were performed. Using our own base station and the data from  
58 the Resolute magnetic observatory (distant of 170 km from Haughton), we  
59 were able to properly remove the external and core field signals from our TF  
60 measurements to produce a magnetic field anomaly map of the Haughton  
61 central area at ground level (Figure 2B). Finally, a local 120 m-long E-W  
62 gradient profile was performed over the maximum of the magnetic anomaly  
63 (Figure 2C).

64 These ground magnetic field data were used to investigate, by model-  
65 ing, the magnetized source of this anomaly. Three modeling approaches  
66 were used. First, the possible sources of the magnetic anomaly were forward  
67 modeled using uniformly magnetized rectangular prisms (e.g., Quesnel et al.,  
68 2008). Their parameters (top and bottom depths, lateral coordinates, total  
69 magnetization) were adjusted by a semi-automatic trial and error approach,

70 starting from a single deep (and thick) body (that creates the long wave-  
 71 length of the anomaly). New subsequent (shallow) prisms were then added  
 72 (and parameters adjusted) until a satisfying model was reached. To estimate  
 73 the model accuracy, the following *Fit* parameter was used:

$$Fit = 100 * \left[ 1 - \sqrt{\frac{(\Delta B_{obs} - \Delta B_{pred})^2}{\Delta B_{obs}^2}} \right] \quad (1)$$

74 where  $\Delta B_{obs}$  and  $\Delta B_{pre}$  correspond to the observed and predicted magnetic  
 75 anomaly values, respectively. The second approach is the Standard Euler  
 76 Deconvolution (SED) method (Reid et al., 1990), which was applied over the  
 77 maximum of the anomaly (but excluding the local gradient profile). It gives  
 78 a range of source depth rather than a real and single source depth. The  
 79 last modeling approach is the GM-SYS module of the GEOSOFT Inc Oasis  
 80 montaj software (mainly based on Campbell, 1983). It was applied to all  
 81 profiles to invert parameters of 2D bodies with more 'geological' shapes.

82 The magnetic susceptibility and natural remanent magnetization (NRM)  
 83 of 289 samples of the Lower Paleozoic sediments (186 from inside and 103  
 84 from outside the impact structure) and 337 samples of the Precambrian base-  
 85 ment (316 from inside and 21 from outside the crater) were measured using  
 86 either a portable SM30 susceptibilimeter from ZH Instruments, or a KLY2  
 87 susceptibilimeter from AGICO for susceptibility, and for NRM either a 2G  
 88 Entreprises 775R magnetometer or a Molspin Minispin magnetometer (for  
 89 large samples). In order to recover the direction of the paleofield at the time  
 90 of the impact, we sampled 20 decimetric oriented clasts (13 of Paleozoic sed-  
 91 iments and 7 of basement) from two sites located 3 km apart in the clast-rich  
 92 impact-melt rocks (Figure 1). Thermal (in a MMTD furnace) or alternating

93 field (with an on-line system in the 2G magnetometer) stepwise demagneti-  
94 zations were used. Characteristic remanent magnetizations were evaluated  
95 by principal component analysis using Paleomac software (Cogné, 2003).

### 96 **3. Results and Interpretation**

#### 97 *3.1. Magnetic field anomaly*

98 The central magnetic anomaly reaches a maximum of 900 nT (point M in  
99 Figure 1 and 2B) while the surroundings globally show a negative magnetic  
100 anomaly signal (down to -100 nT). The total wavelength of the central mag-  
101 netic anomaly corresponds to about 1.2 km in all directions, except towards  
102 SE where a positive signal is still present. This wavelength is coherent with  
103 Pohl et al. (1988), but these authors did not precisely locate the maximum  
104 and underestimated the amplitude of the magnetic anomaly. Glass et al.  
105 (2012, and references within) also did not show details about this anomaly,  
106 although it still appears to be intense at the altitude of their airborne survey.

107 The maximum of the ground anomaly consists of a local sharp transition  
108 where a 20 nT/m vertical gradient was measured (Figure 2C). This suggests  
109 that the top of the magnetized source occurs at very shallow depths in or  
110 below the impact melt sheet at Anomaly Hill.

#### 111 *3.2. Acquisition time and type of magnetization*

112 The clasts of the impact melt rock should have acquired their magnetiza-  
113 tion during cooling of the melt sheet in the geomagnetic field at the time of  
114 impact (e.g., Koch et al., 2012). Indeed, the melt rock matrix contains car-  
115 bonate melts with liquidus temperatures of about 500-600°C (Osinski et al.,

116 2005c), which is above the maximum blocking temperature of most clasts  
117 whose magnetic mineralogy is dominated by pyrrhotite with a Curie tem-  
118 perature of 320°C (see Figure 1 of the Supplementary Material). The 20  
119 clasts sampled in the impact melt rocks show a single component of mag-  
120 netization (Figure 3). The two sites provide undistinguishable mean paleo-  
121 omagnetic directions. The mean direction computed from individual clast  
122 direction is  $I=71^\circ$ ,  $D=341^\circ$  ( $\alpha_{95}=6.4^\circ$ ,  $n=16/20$ ; see Figure 2 of the Supple-  
123 mentary Material), in agreement with a normal polarity regional direction  
124 of the time-averaged geomagnetic field at the impact location at  $\sim 39$  Ma  
125 ( $I=80^\circ$ ,  $D=-34^\circ$ ; Besse and Courtillot, 2002). This direction is also close  
126 to the direction of the geomagnetic field in 2010 ( $I=87^\circ$ ,  $D=-41^\circ$ ; Finlay  
127 et al., 2010). Therefore, because induced (i.e., due to the present field) and  
128 measured remanent magnetizations have similar directions (angle of  $16^\circ$ ; see  
129 Figure 2 of the Supplementary Material), their sum (total magnetization)  
130 was used in the following models regardless of their respective contribution.

### 131 *3.3. Rock magnetism*

132 The measurements of magnetic susceptibility and NRM on rock sam-  
133 ples show that all Palaeozoic sedimentary rocks, both inside and outside the  
134 crater, have total magnetization intensity weaker than 0.01 A/m (Table 1  
135 and Figure 4). Pyrrhotite is the main magnetic carrier of the remanent mag-  
136 netization of these rocks (Figure 3 of the Supplementary Material). Among  
137 the 337 basement samples, only 4% (12 samples), both from inside and out-  
138 side the crater, show total magnetization intensities larger than 1 A/m with a  
139 maximum of 7.4 A/m for a basaltic clast from the melt rock (Figure 5). These  
140 12 samples are either granites or diorites from outside the crater, or unusual



141 clasts from the melt rock in the crater: basalts, diorites, mafic rocks. These  
142 clasts come from the Precambrian basement (where they occur as dykes and  
143 lenses) underlying the Palaeozoic sediments. The corresponding formations  
144 are minor in the Precambrian shield (Frisch and Trettin, 1991) as in the clast  
145 population (Redeker and Stoeffler, 1988). In the few clasts found nearby the  
146 magnetic center (Anomaly Hill area), both magnetite (and/or maghemite)  
147 and pyrrhotite can carry these large (mainly remanent) magnetizations (see  
148 Figures 4 and 5 of the Supplementary Material). Most basement samples, in  
149 particular all the gneisses, have low susceptibility (median value  $3.0 \cdot 10^{-4}$  SI)  
150 and low NRM (median value  $1.2 \cdot 10^{-2}$  A/m), resulting in a low total magne-  
151 tization (median value  $2.9 \cdot 10^{-2}$  A/m). The median (mean) Koenigsberger  
152 ratio ( $Q$ , remanent magnetization intensity over induced magnetization in-  
153 tensity) for basement clasts from inside the crater is 0.88 (resp. 1.82), in-  
154 dicating an approximately equivalent contribution of induced and remanent  
155 magnetization to the total magnetization (Table 1). It is noteworthy that the  
156 Koenigsberger ratio is significantly lower for basement samples collected out-  
157 side the crater than for samples collected inside the crater (median  $Q=0.11$   
158 against 0.88, respectively). This is explained by the nature of the NRM of  
159 the samples from inside the crater. Indeed, these samples all come as clasts  
160 from the impact melt rock layer. Therefore, they carry a thermoremanent  
161 magnetization (TRM) acquired soon after the impact, whereas the remanent  
162 magnetization of basement samples from outside the crater may not be a  
163 TRM, but rather a chemical remanent magnetization (CRM), and moreover  
164 it was affected by more than 500 Myrs of viscous decay.

165 *3.4. Magnetic source modeling*

166 After forward modeling, the best combination corresponds to 12 prisms  
167 that comprise a  $0.7 \text{ km}^3$  magnetized source body with a root at 1 km depth  
168 (see Table 1 of the Supplementary Material; Figure 6). It fits the data with  
169  $\sim 90\%$  of accuracy, while this value reaches 98% for the local anomaly profile  
170 (see Figure 6 of the Supplementary Material). The resulting root is similar  
171 to the root suggested by Pohl et al. (1988), indicating that only the top  
172 core of the central uplift is more magnetized than the surrounding rocks. At  
173 shallow depths (i.e., less than 30 m), several small isolated magnetized bodies  
174 are necessary to explain the local variations of the magnetic anomaly and  
175 vertical gradient signals over the maximum. Most prisms need a 1.5 A/m  
176 magnetization to account for the anomaly, while the most superficial ones  
177 require a larger intensity that can reach 2.3 A/m (Table 1 of Supplementary  
178 Material).

179 Using a structural index of 1, a window size of 700 m, a depth uncertainty  
180 less than 20% and the interpolated magnetic anomaly grid as inputs, the SED  
181 resulting solutions beneath the maximum of the anomaly are clustered with  
182 depths ranging from 25 to 45 m (see Figure 7 of the Supplementary Material).  
183 The local gradient profile was not considered during the SED calculations,  
184 leading to a deeper top of the source than expected at local scale. Then  
185 it only confirms that the top of the magnetic part of the central uplift is  
186 shallower than about 35 m.

187 Using the GM-SYS module of the GEOSOFT Inc Oasis montaj software,  
188 the best model consists in a  $\sim 1 \text{ km}^3$  magnetic body with superficial branches  
189 buried in the first 10 m below the surface (see Figure 8 of the Supplementary

190 Material). Therefore, all modeling approaches suggest that the source of this  
191 magnetic anomaly is a large magnetized body with a 1 km-deep root and  
192 with a top composed of several isolated (but close) volumes that could be  
193 accessible by a 20-30 m deep drilling. This large body was already suggested  
194 by the model of Pohl et al. (1988) – same root, same mean magnetization  
195 – but no details about the shape and depth of its top were shown.

#### 196 **4. Discussion**

197 Our investigations confirm that Palaeozoic sediments cannot contribute  
198 to the central magnetic anomaly at Haughton impact structure. Unexposed  
199 crystalline basement bodies are sometimes suggested as the magnetic source  
200 for impact magnetic anomalies (Ugalde et al., 2007). However, the low mean  
201 total magnetization of the basement samples and the rarity of samples pos-  
202 sessing sufficient total magnetization argue against this hypothesis for the  
203 Haughton case. Furthermore, the vast majority of the basement clasts in  
204 the melt rocks inside the crater are gneisses (Redeker and Stoeffler, 1988),  
205 suggesting that the uplifted basement is mostly composed of gneisses, which  
206 have total magnetizations well below 1 A/m. The modeled pyramidal ge-  
207 ometry of the source and its expected high porosity also argue against the  
208 hypothesis of an unexposed silicate impact melt body for the source of the  
209 magnetic anomaly, as proposed for Chesapeake Bay (Shah et al., 2005) or  
210 Ries (Pohl et al., 2010) craters. Therefore, we suggest that an additional  
211 impact-related process is required to enhance the induced and/or remanent  
212 magnetization of the rocks responsible for the anomaly. Experimental works  
213 have shown that shock can modify the intrinsic magnetic properties of rocks

214 (e.g., Gattacceca et al., 2007), and lead to demagnetization or remagnetiza-  
215 tion (e.g., Gattacceca et al., 2010). Our rock magnetic measurements show  
216 that the dominant ferromagnetic mineral in the Paleozoic sedimentary tar-  
217 get rocks and in the gneissic basement is pyrrhotite (see Figures 1 and 3 of  
218 the Supplementary Material). Pyrrhotite should have been largely remagne-  
219 tized by impact pressures above 1 GPa (Rochette et al., 2003; Gilder et al.,  
220 2011). The intrinsic magnetic properties of pyrrhotite are also affected by  
221 shock (Gattacceca et al., 2007; Louzada et al., 2007). But neither the shock  
222 magnetization (less efficient than thermoremanent magnetization; see Gat-  
223 tacceca et al., 2008), nor the expected modifications can make the studied  
224 rocks magnetic enough to account for the observed magnetic anomaly.

225 Therefore, we suggest that the best hypothesis to account for this central  
226 magnetic anomaly is post-impact hydrothermal alteration of a basement core  
227 of the Haughton central uplift (Figure 7). There is widespread evidence of  
228 hydrothermal activity at Haughton, dominated by vugs and veins of calcite,  
229 sulfides and sulfates (Osinski et al., 2005a). Indeed the hot superficial im-  
230 pactite layer provided the heat that, in combination with the meteoric, post-  
231 impact lake and ground water circulation, led to hydrothermal alteration of  
232 the target rocks. Most of the outcrops showing hydrothermal-derived min-  
233 eralizations are present at the base of the melt layer, which is best exposed  
234 around the edge of the central uplift (Osinski et al., 2005a). However, the  
235 more intense fracturing and porosity at the center of the Haughton structure  
236 - suggested by the low gravity anomaly, the low densities of highly-shocked  
237 basement clast densities (Singleton et al., 2011) and by the heterogeneity  
238 of the shallow part of our magnetic model - would have favored water cir-

239 culation and associated hydrothermal activity at this location (Figure 7).  
240 Hydrothermal activity has been suggested as the process that creates the  
241 main magnetic anomaly in large impact craters, such as Chicxulub (Pilking-  
242 ton and Hildebrand, 2000; Escobar-Sanchez and Urrutia-Fucugauchi, 2010).  
243 For intermediate- and small-size craters, if post-impact hydrothermal activ-  
244 ity occurred, to our knowledge, this process has not been invoked to pro-  
245 duce a central magnetic anomaly to date (e.g., Pilkington and Grieve, 1992;  
246 Kenkmann et al., 2005).

247 Over Precambrian formations affected by low grade metamorphism and  
248 mineralization, pyrrhotite-bearing rocks and magnetite-bearing rocks are of-  
249 ten responsible for local intense aeromagnetic anomalies (Airo, 2002). Inter-  
250 estingly, the remanence of such rocks is mainly carried by pyrrhotite (with low  
251 susceptibilities), while magnetite have high susceptibility and low remanence  
252 (Direen et al., 2008). Such remanence is often acquired through crystalliza-  
253 tion of pyrrhotite during retrograde low-level metamorphism when sulfides  
254 are remobilized in shear zones. On the other hand, magnetite is thought to  
255 be produced during prograde metamorphism (Direen et al., 2008). It means  
256 that local accumulations of any of these two minerals can occur during a  
257 mild thermal event, like in post-impact hydrothermal activity. Sulfides and  
258 sulfates are abundant in hydrothermal-derived parageneses around the edge  
259 of the central uplift at Haughton (Osinski et al., 2005a), which argues for  
260 post-impact crystallization of pyrrhotite (Kontny et al., 2007) in other areas  
261 of the crater, like the core of the central uplift. Magnetite of hydrother-  
262 mal origin can also contribute to magnetic anomalies in impact craters, in  
263 association with pyrrhotite (Mang et al., 2012). These newly-formed min-

erals carry a chemical or thermo-remanent magnetization. Using reasonable assumptions about the intrinsic magnetic properties of pyrrhotite and magnetite as well as about the efficiency of chemical and thermo-remanent magnetization, it appears that the crystallization of about 0.1 wt% (resp. 0.6 wt.%) of magnetite (resp. pyrrhotite) is enough to account for a total magnetization of 1 A/m (see explanations at the end of the Supplementary Material). Such concentrations are consistent with those measured in some other hydrothermal systems (see Clark, 1999, and references within). An alternative impact-generated process that can lead to the production of fine-grained magnetite is shock-wave decomposition of the target mafic minerals (Pohl et al., 2010). This could explain the presence of magnetite in the rare highly-shocked strongly-magnetized clasts (see Figures 4 and 5 of the Supplementary Material), and, perhaps, associated with pyrrhotite in the uplifted gneissic basement.

If hydrothermalism occurred and enhanced the initial magnetization of the target rocks at Haughton, it implies that a similar process may have occurred on other terrestrial planets. It requires the presence of subsurface water, of an ambient magnetic field and of iron-bearing rocks in the target crust. Concerning subsurface water, among the possible candidates excepted the Earth, only Mars and possibly some asteroids are suitable. Indeed geological and morphological indices of impact-generated hydrothermalism have been observed on Mars (e.g., Marzo et al., 2010; Squyres et al., 2012; Osinski et al., in press (online publication Oct 2012)). Numerical modeling is also suggestive of its occurrence for Martian craters (Abramov and Kring, 2005; Rathbun and Squyres, 2002). However, the high-altitude of the mag-

289 netic field measurements made on Mars prevents the detection of magnetic  
290 anomalies over impact craters with diameter less than 200 km (Langlais and  
291 Thébault, 2011). Therefore, new low-altitude surveys are necessary to unveil  
292 such central magnetic anomalies. Nevertheless, hydrothermalism has proba-  
293 bly occurred and enhanced the crustal magnetization on early Mars not only  
294 after impacts (external origin), but also because large heat sources from the  
295 early Martian mantle are expected and may have led to the hydrothermal al-  
296 teration of a large part of the crust (e.g., Meunier et al., 2012; Quesnel et al.,  
297 2009). However, near-surface hydrothermal activity in impact craters (some-  
298 times with lake) constitutes an excellent place for the initiation, development  
299 and long-term existence for life (e.g., Cockell et al., 2010).

## 300 **5. Conclusion**

301 The objective of this study was to determine the nature of the magnetized  
302 source responsible for the intense central magnetic anomaly observed within  
303 the Haughton impact structure by investigating the geometry and magneti-  
304 zation of this source. We found that no target rocks at Haughton possess a  
305 total magnetization large enough to account for the magnetic anomaly, ex-  
306 cept a small fraction (4%) of unusual basement samples. The high density of  
307 these rare samples is incompatible with the negative gravity anomaly. It is  
308 therefore probable that the actual magnetic source lithology is not exposed  
309 in the crater and is not found as basement clasts within the melt rock sheet.  
310 Using different modeling methods, we concluded that the magnetic source  
311 body has a 1-km deep root but a very superficial top (less than 10 m) that  
312 could be drilled in a future mission. The enhancement of the total mag-

313 netization of this source is best explained by impact-generated hydrother-  
314 mal activity favored by intense fracturing and high porosity of the uplifted  
315 basement (low gravity anomaly), leading to the crystallization of magnetite  
316 and/or pyrrhotite. Our model thus explains the association of a large central  
317 magnetic anomaly with a low gravity anomaly over this impact structure.  
318 This may help for the interpretation of the low-altitude geophysical signa-  
319 ture of impact craters on other planets, and confirms that impact structures  
320 are ideal places for life. In addition, our study shows that the impact took  
321 place during a normal polarity interval of the geomagnetic field, providing  
322 an additional chronologic constraint.

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Figure 1: Location and geology of the Haughton impact structure. On top-left location map, RES corresponds to the RESolute magnetic observatory located at 170 km from the Haughton crater. On the central map, the surveyed area is indicated by the white broken box with letter 'M' indicating the peak of the central magnetic anomaly, located at E 424677 m, N 8367960 m in the North American Datum 1927 system. This 'M' point is used as origin for all magnetic maps and profiles. The two stars indicate the two sites where the impact melt rock was sampled for paleomagnetism, while the dashed circles represent the external and internal limits of a well-fractured zone with radial and concentric faults. Lithologies: 1, Quaternary fluvial and fluvio-glacial sediments; 2, Haughton formation (lacustrine sediments); 3, Impact melt rocks; 4 and 5, Upper Ordovician to Silurian Allen Bay formation, Middle (dolomite) and Lower (limestone) Members, respectively; 6, Middle to Upper Ordovician Thumb Mountain formation (limestone and dolomite); 7, Middle Ordovician Bay Fiord formation (anhydrite and dolomite); 8, Lower to Middle Ordovician Eleanor River formation (limestone and dolomite).

Figure 2: Raw total-field measurements (A), interpolated magnetic field anomaly map (B) and local vertical gradient E-W profile over the magnetic center (C). The background on (A) shows the topographical contours of the studied area, while point 'M' denotes the maximum of the magnetic anomaly, chosen as origin for all maps and profiles. The location of the gradient profile (C) is shown by a thin black line over this maximum on the anomaly map (B). On (C), HP means High Probe signal (1.5 m above the ground), LP means Low Probe signal (0.5 m above the ground), while GD means Gradient signal. 'f.m.c.' means 'from magnetic center'.

Figure 3: Orthogonal projection plots of stepwise thermal demagnetization data of sedimentary and basement clasts from the impact melt rock. Open and solid symbols are projections on vertical and horizontal planes, respectively. Demagnetization steps are indicated in °C. NRM means Natural Remanent Magnetization.

Figure 4: Mean total (induced + remanent) magnetization ( $M_{tot}$ ) and Koenigsberger ratio ( $Q$ , ratio of remanent over induced magnetization) of all target formations. The range is indicated by the 'error' bars. The gray area in the total magnetization diagram indicates the range of values (1.5-2.3 A/m) necessary for the source of the magnetic anomaly (see text). Acronyms are: AB, Allen Bay formation; TM, Thumb Mountain formation; BF, Bay Fiord formation; ER, Eleanor River formation; CO, other Cambrian-Ordovician sedimentary formations sampled outside the crater; PB, Precambrian Basement.

Figure 5: Frequency histogram of total magnetization for 337 basement samples from Devon Island (316 from inside the crater, 21 from outside the crater). Note the log scale for total magnetization.

Figure 6: 3D view showing the resulting 12-prisms forward model of the magnetic source of the central magnetic anomaly observed at Haughton. The brown level of each uniformly magnetized prism depends on its magnetization (brown scale in A/m). The surface over these prisms corresponds to the interpolated magnetic anomaly grid seen on Figure 2B with relief and color level depending on the anomaly intensity (in nT). The top right panel zooms in the superficial prisms. Same origin as for the map of Figure 1. 'f.m.c.' means 'from magnetic center'.

Figure 7: Schematic representation of the post-impact hydrothermal alteration that enhanced the magnetization in the porous core (dotted area) of the central uplift due to concentrations of newly-formed pyrrhotite and/or magnetite mineralizations.

## \*Highlights (for review)

- We investigate the source of the central magnetic anomaly at the Haughton impact structure.
- A large magnetized body with superficial branches and with 1.5 to 2.3 A/m magnetization intensities is discovered.
- No visible rocks can carry such magnetization.
- Post-impact hydrothermal activity at the top of the central uplift may have enhanced the initial basement rock magnetization.

Figure 1  
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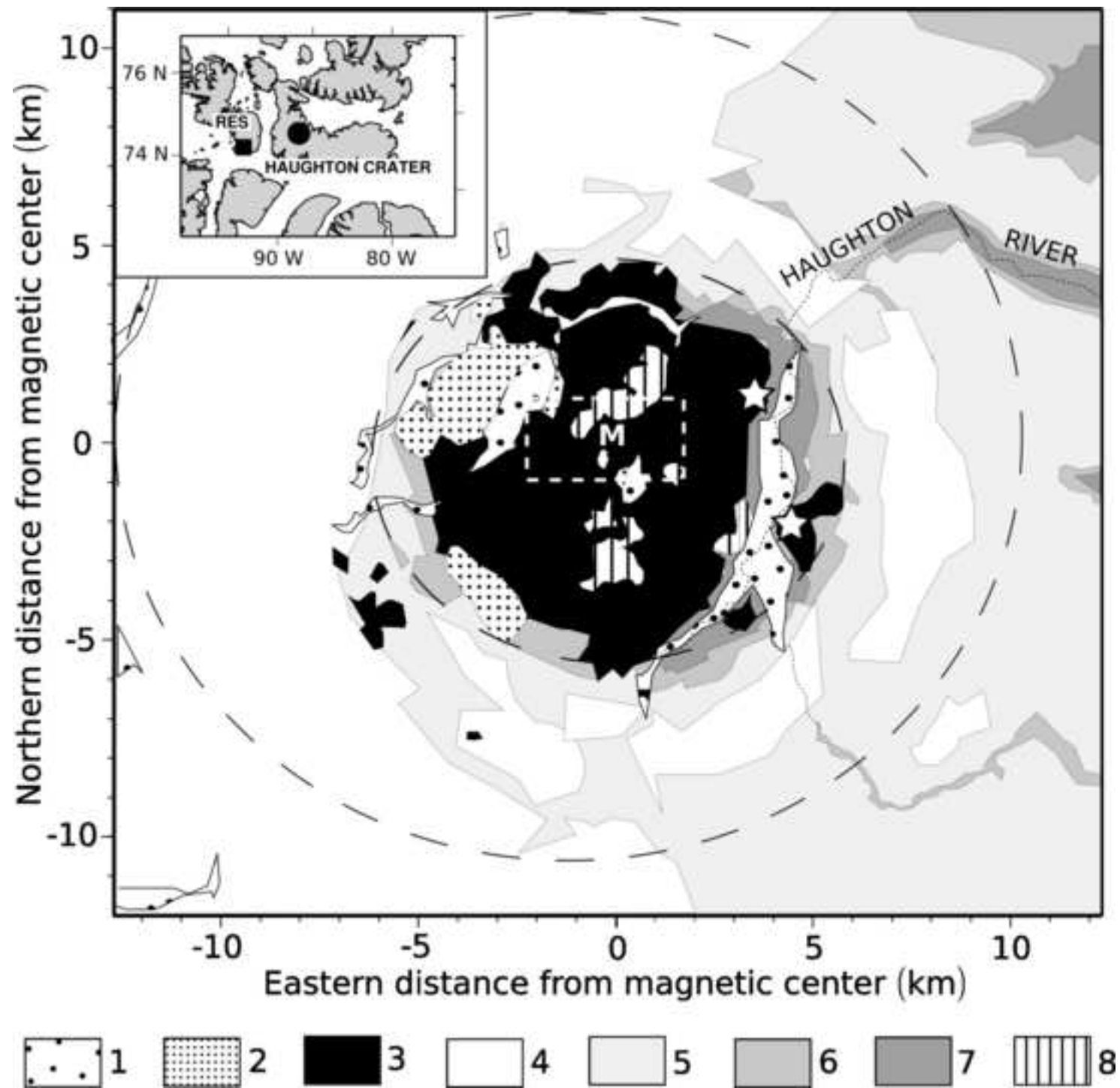


Figure 2

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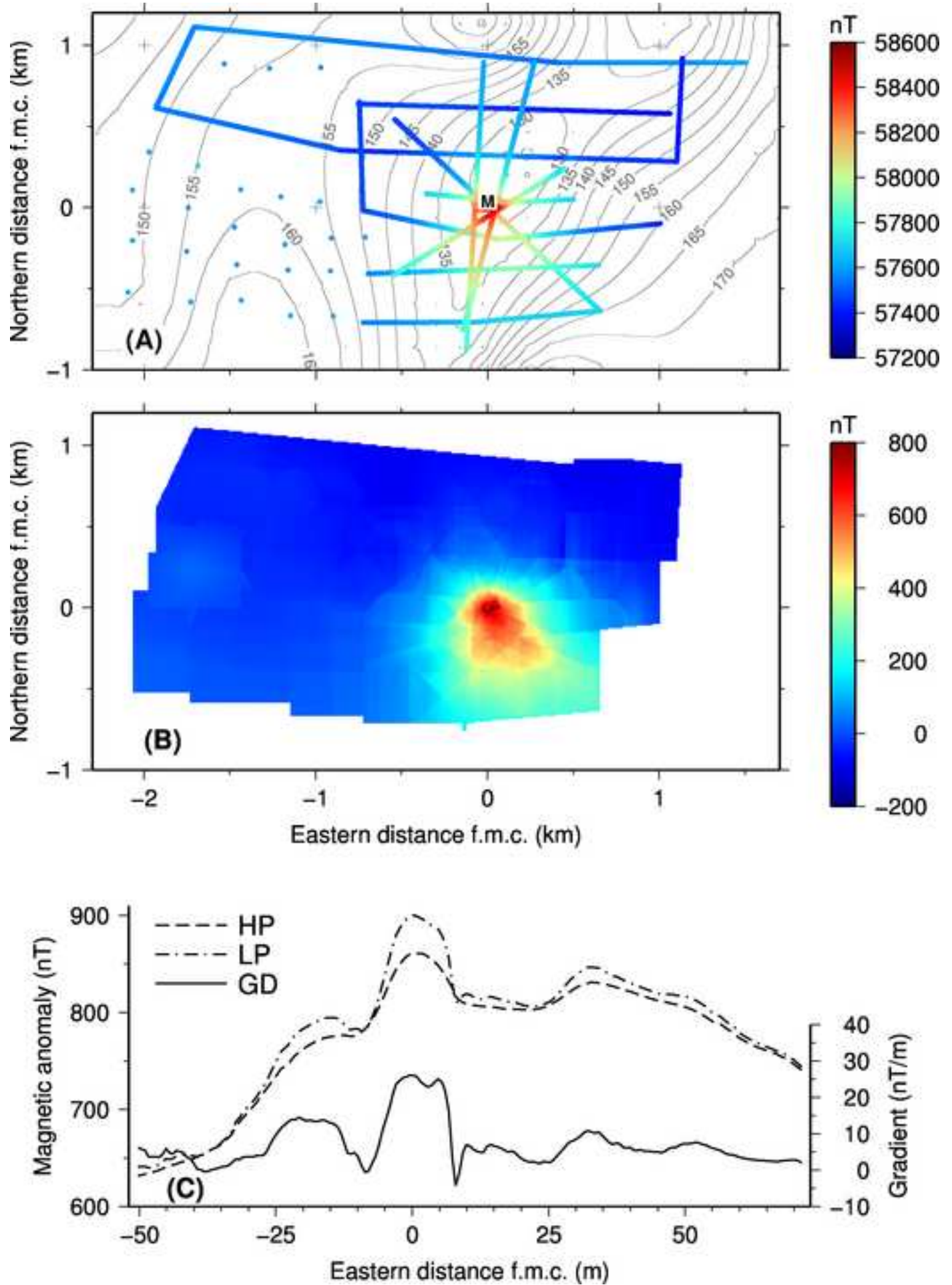


Figure 3  
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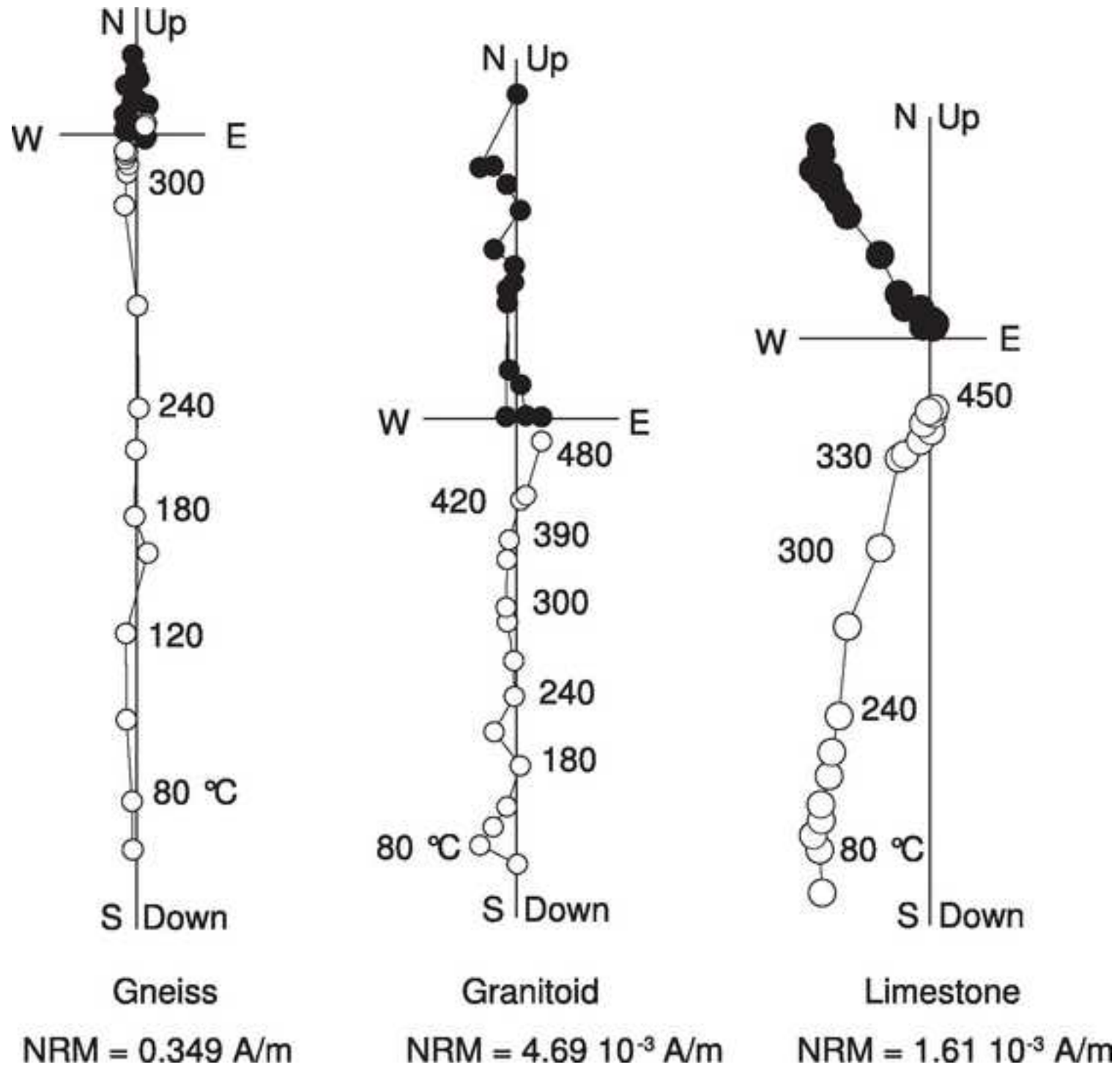


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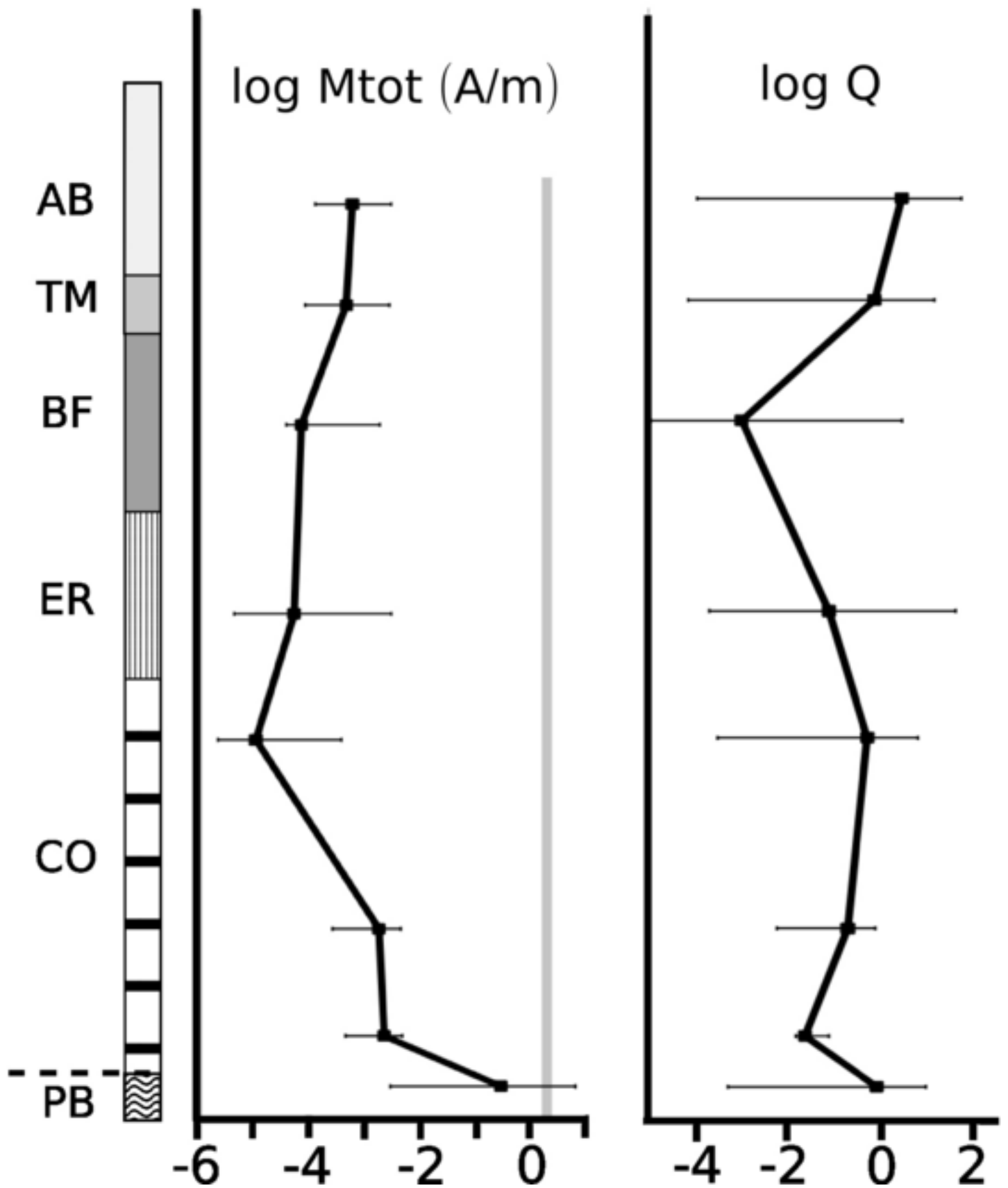
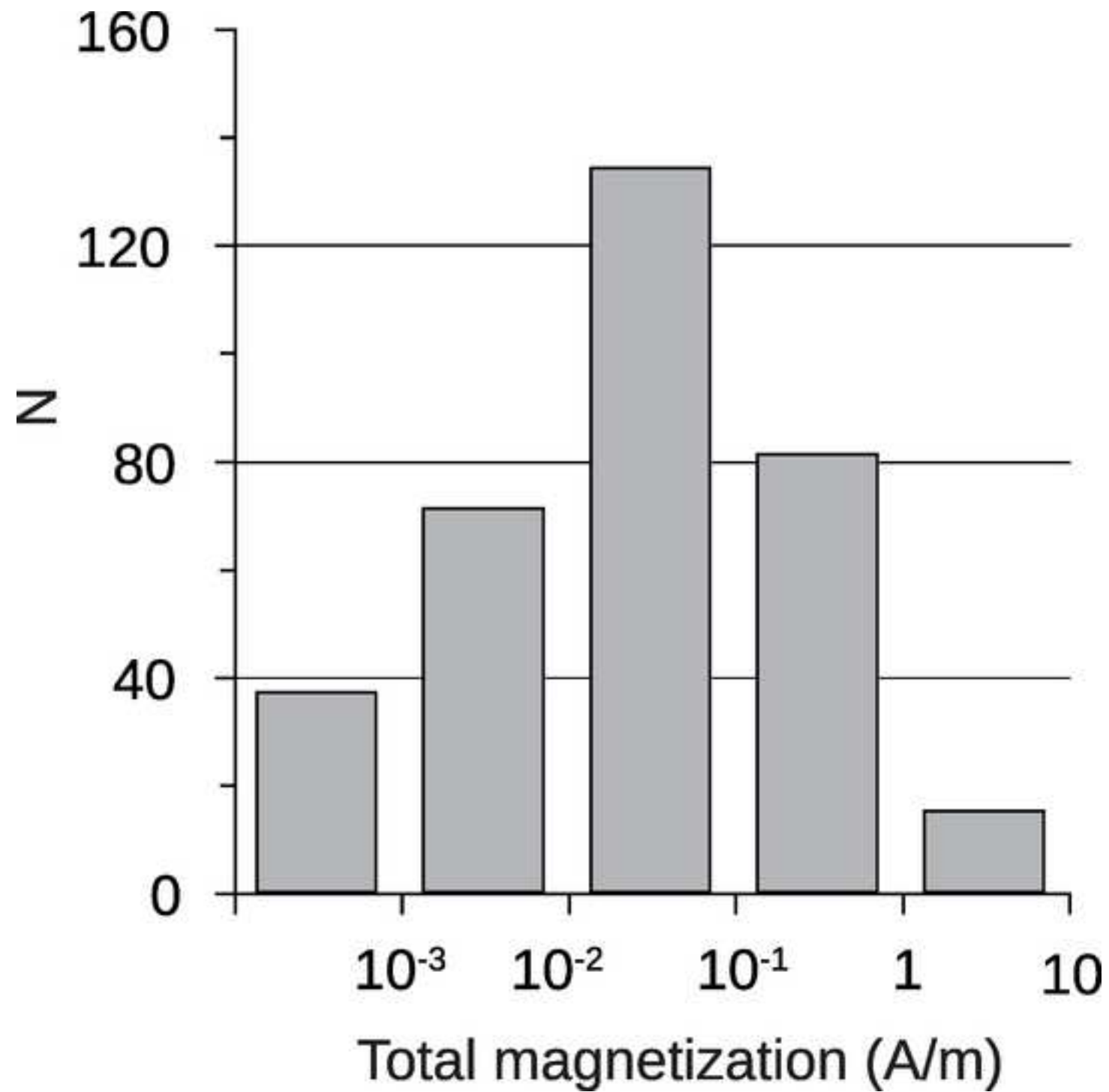


Figure 5  
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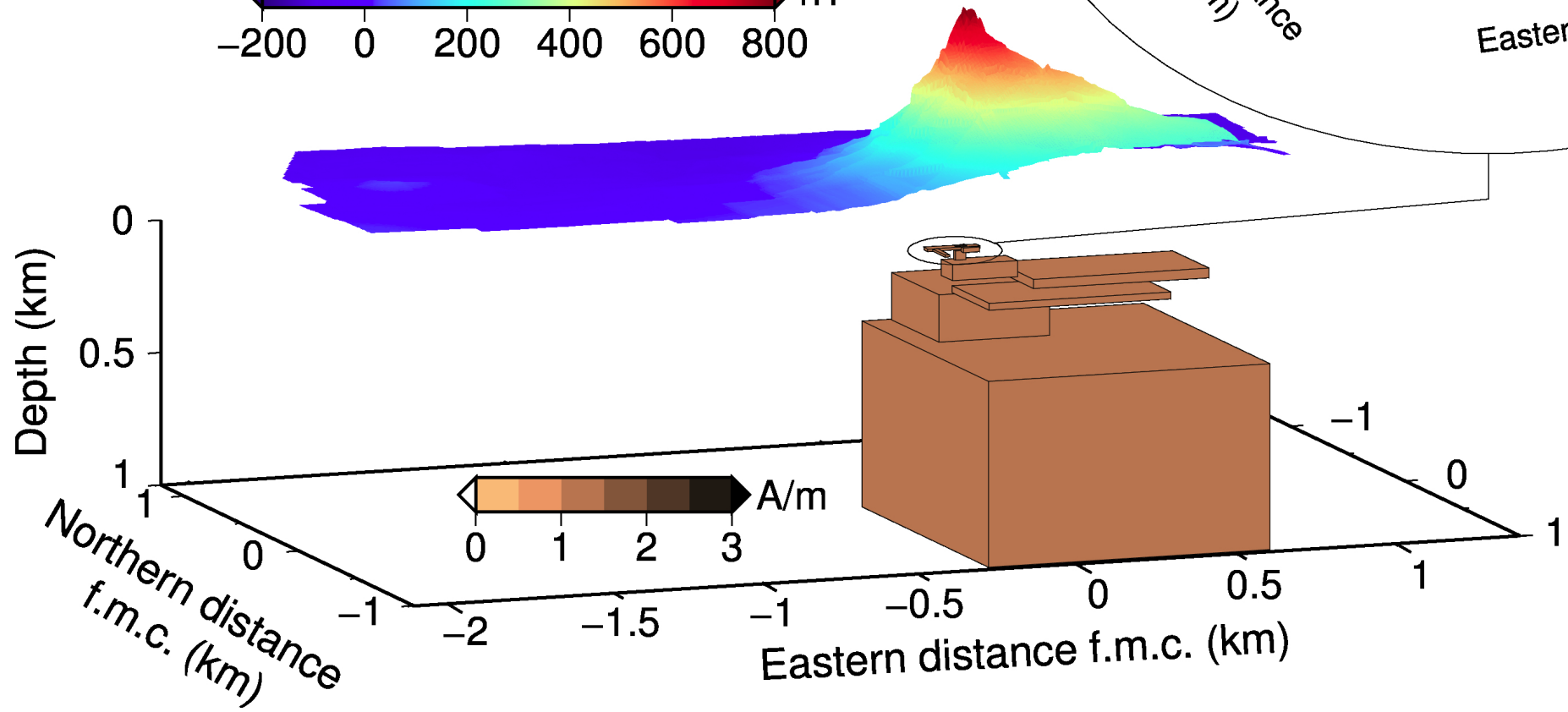
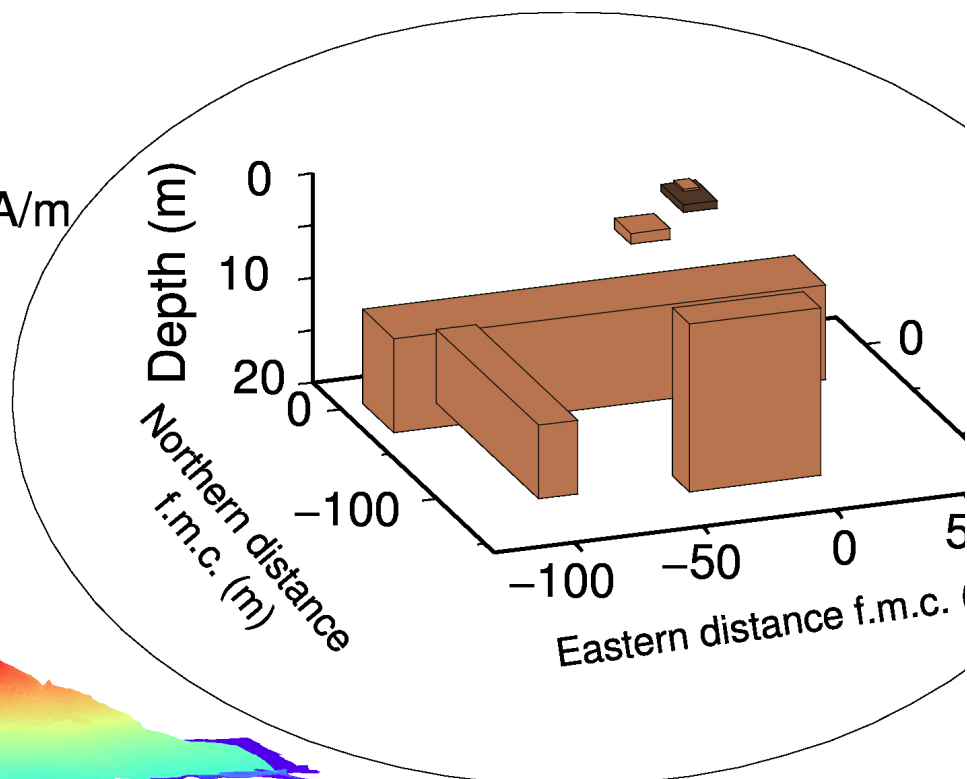
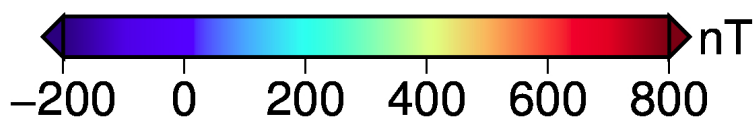
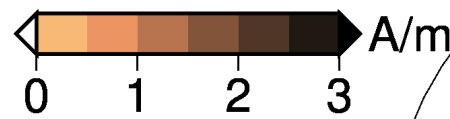


Figure 7  
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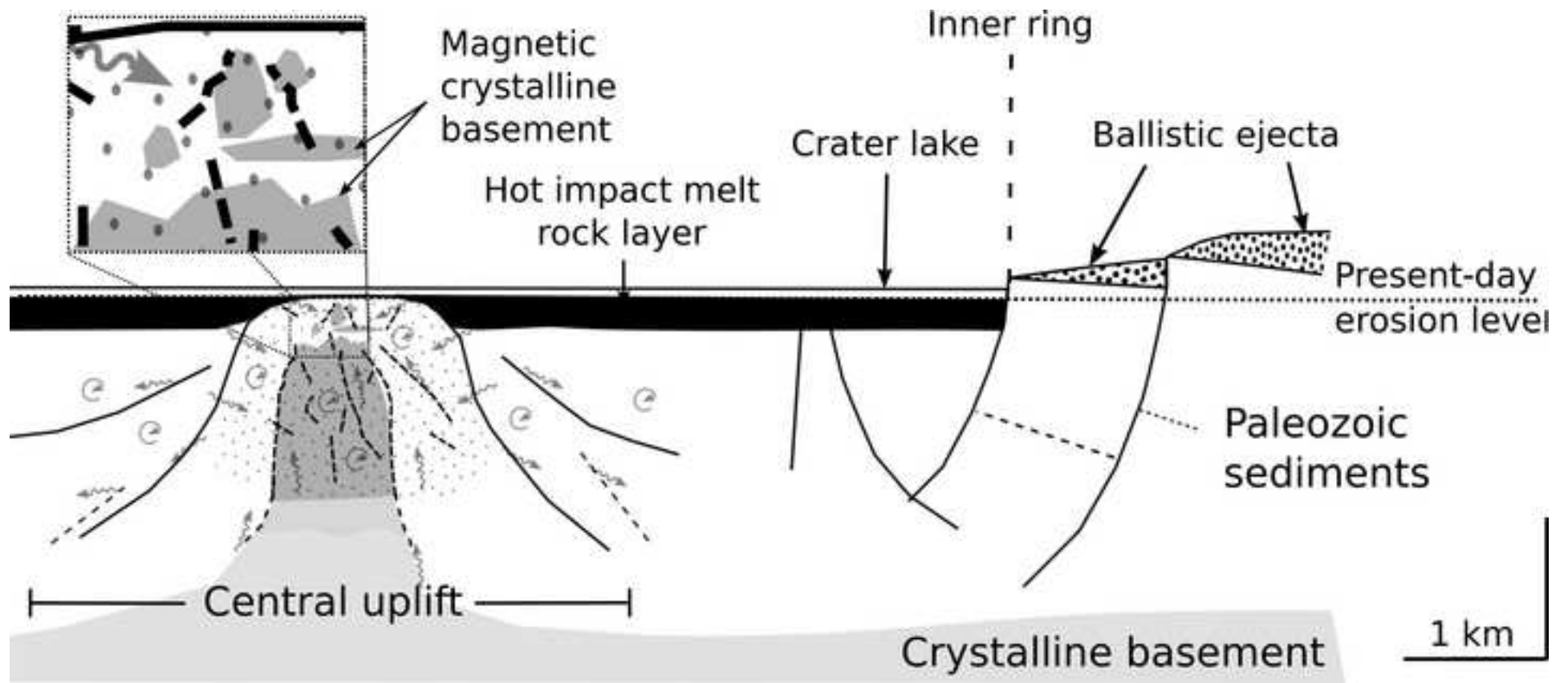


Table 1: Magnetic properties of Haughton target rocks.

Name of formation	N	K ( $10^{-3}$ SI)		NRM ( $10^{-3}$ A/m)		Mtot ( $10^{-3}$ A/m)		Q	
		Mean	Max	Mean	Max	Mean	Max	Mean	Max
Allen Bay	25	0.006	0.025	0.347	1.523	0.601	2.524	2.74	52.89
Thumb Mountain	42	0.006	0.038	0.186	0.688	0.460	2.419	0.73	14.02
Bay Fiord	39	-0.004	0.026	0.095	0.654	0.069	1.854	0.04	2.80
Eleanor River	108	-0.005	0.007	0.200	3.343	0.051	3.086	0.07	39.86
Blanley Bay	20	-0.002	0.007	0.085	0.249	0.010	0.367	0.50	6.22
Cass Fiord	24	0.038	0.060	0.142	0.458	1.858	2.802	0.19	0.75
Bear Point	11	0.050	0.057	0.051	0.138	2.319	2.689	0.02	0.07
Basement (inside)	316	2.260	89.90	97.00	3610.	215.0	7350	1.82	9.21
Basement (outside)	21	7.400	89.20	88.50	476.0	443.0	4740	0.33	1.18

N, number of studied samples; K, magnetic susceptibility; NRM, Natural Remanent Magnetization; Mtot, total (induced + remanent) magnetization intensity; Q, Koenigsberger ratio (remanent over induced magnetization ratio).