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Preservation and detectability of shock-induced magnetization

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Citation: Tikoo, Sonia M. et al. "Preservation and Detectability of Shock-Induced Magnetization: PRESERVATION AND DETECTABILITY OF SRM." *Journal of Geophysical Research: Planets* 120.9 (2015): 1461–1475.

As Published: <http://dx.doi.org/10.1002/2015JE004840>

Publisher: American Geophysical Union (AGU)

Persistent URL: <http://hdl.handle.net/1721.1/106491>

Version: Author's final manuscript: final author's manuscript post peer review, without publisher's formatting or copy editing

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Peer Reviewed

Title:

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Journal Issue:

Journal of Geophysical Research: Planets,

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Publication Date:

July 29, 2015

Series:

[UC Berkeley Previously Published Works](#)

Permalink:

<http://escholarship.org/uc/item/09v7b29x>

DOI:

<https://doi.org/10.1002/2015JE004840>

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1 **Preservation and detectability of shock-induced magnetization**

2
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15 *Final version for publication in Journal of Geophysical Research - Planets*

16 *Accepted 07/24/2015*

17 *DOI:10.1002/2015JE004840*
18

19 **Abstract**

20
21 An understanding of the effects of hypervelocity impacts on the magnetization of natural samples
22 is required for interpreting paleomagnetic records of meteorites, lunar rocks, and cratered planetary
23 surfaces. Rocks containing ferromagnetic minerals have been shown to acquire shock remanent
24 magnetization (SRM) due to the passage of a shock wave in the presence of an ambient magnetic
25 field. In this study, we conducted pressure remanent magnetization (PRM) acquisition experiments
26 on a variety of natural samples as an analog for SRM acquisition at pressures ranging up to 1.8
27 GPa. Comparison of the alternating field (AF) and thermal demagnetization behavior of PRM
28 confirms that AF demagnetization is a more efficient method for removing SRM overprints than
29 thermal demagnetization because SRM may persist to unblocking temperatures approaching the
30 Curie temperatures of magnetic minerals. The blocking of SRM to high temperatures suggests that
31 SRM could persist without being eradicated by viscous relaxation over geologic timescales.
32 However, SRM has been rarely observed in natural samples likely because of two factors: [1] other
33 forms of impact-related remanence (e.g., thermal remanent magnetization from impact-related
34 heating or chemical remanent magnetization from post-impact hydrothermal activity) are often
35 acquired by target rocks that overprint SRM, and [2] low SRM acquisition efficiencies may prevent
36 SRM from being distinguished from the underlying primary remanence or other overprints due to
37 its low magnetization intensity.
38

39 1. Introduction

40 The ubiquity of hypervelocity impact events throughout solar system history motivates an
41 understanding of the effects of impacts on both terrestrial and extraterrestrial rocks. In the context
42 of paleomagnetism, shock remagnetization is expected to occur in any geologic environment that
43 has been subjected to impacts. Shock remanent magnetization (SRM) may be acquired nearly
44 instantaneously as the shock wave from an impact passes through a rock in the presence of a
45 magnetic field [*Nagata, 1971; Pohl et al., 1975*]. SRM is usually aligned with the ambient
46 magnetizing field with an intensity proportional to the field strength for weak planetary fields (~1-
47 2500 μT) [*Nagata, 1971; Gattacceca et al., 2008; Gattacceca et al., 2010a*]. Therefore, SRM is
48 capable of recording long-lived core dynamo magnetic fields as well as transient fields such as
49 those hypothesized to be generated or amplified by impact plasmas [*Srnka, 1977; Crawford and*
50 *Schultz, 1993; Hood and Artemieva, 2008*]. SRM has been proposed as a potential source for the
51 natural remanent magnetization (NRM) present in some lunar samples [*Cisowski et al., 1976;*
52 *Gattacceca et al., 2010b*] and meteorites [*Weiss et al., 2010*] as well as for secondary
53 magnetization components present in rocks from terrestrial impact craters (e.g., *Halls [1979]*). In
54 the absence of an ambient field, shock waves can demagnetize rocks [*Nagata, 1971; Gattacceca*
55 *et al., 2006*]. Shock demagnetization may be responsible for the modification of magnetic
56 anomalies observed in the Martian [*Hood et al., 2003*] and lunar crust [*Halekas et al., 2002*].

57 SRM may be acquired in multiple ways that depend on the nature of the ferromagnetic
58 grains within a rock. Shock waves remagnetize multidomain (MD) grains through the
59 rearrangement of domain walls [*Bogdanov and Vlasov, 1966; Nagata, 1973*]. In single domain
60 (SD) grains, shock-induced stresses introduce magnetoelastic energy that can exceed the
61 anisotropy energy associated with a pre-existing remanent magnetization and impart a new
62 magnetization [*Hodych, 1977; Dunlop and Ozdemir, 1997*]. Shock pressures in excess of the
63 Hugoniot elastic limit (typically ~3 GPa for silicates) may introduce crystallographic defects that
64 result in irreversible changes to intrinsic magnetic properties and can impart magnetic anisotropy
65 [*Gattacceca et al., 2007; Louzada et al., 2007; Gilder and Le Goff, 2008; Mang et al., 2013*]. As
66 rocks experience decompression, these effects combine to impart rocks with SRM or its
67 hydrostatic analog, pressure remanent magnetization (PRM). Note that in the literature another
68 term, piezoremanent magnetization (also abbreviated as PRM; e.g., *Nagata and Carleton [1968]*
69 and *Gattacceca et al. [2010a]*), has been inconsistently used to describe remanence induced

70 through either hydrostatic or non-hydrostatic pressure. The mechanism of PRM acquisition may
71 differ somewhat from that of SRM for at least two reasons. First, only weak deviatoric stresses are
72 present when rocks are pressurized quasi-hydrostatically in the laboratory [*Nagata, 1966; Martin*
73 *and Noel, 1988*]. Second, the pressurization time in typical PRM experiments (>10 s) is longer
74 than the duration of laser shock ($\sim 10^{-9}$ to 10^{-8} s) or typical natural impact events ($\sim 10^{-3}$ to 1 s) that
75 would impart SRM. Nevertheless, we consider PRM to be a good analog for SRM, at least for
76 peak pressures $< \sim 2$ GPa. Similar behavior between PRM and SRM at these pressures has been
77 observed in acquisition experiments on some lunar rocks and the Allende meteorite that have
78 shown that these samples acquire similar intensities of PRM and SRM at equivalent pressures
79 [*Nagata, 1971; Gattacceca et al., 2010b; Carporzen et al., 2011*]. Pressure experiments on natural
80 pyrrhotite also indicate that variations in non-hydrostaticity do not significantly affect
81 magnetization intensity and other magnetic properties at these pressures [*Gilder et al., 2011*].

82 The acquisition of SRM and PRM and the response of these remanences to alternating field
83 (AF) demagnetization have been described in several studies [*Gattacceca et al., 2007; Gattacceca*
84 *et al., 2008; Gattacceca et al., 2010a*]. PRM and SRM are recorded preferentially in the low
85 coercivity fraction of magnetic grains and can therefore be removed more efficiently using
86 progressive AF demagnetization than other forms of remanence such as thermoremanent
87 magnetization (TRM), anhysteretic remanent magnetization (ARM, often used as a room-
88 temperature analog for TRM), and saturation isothermal remanent magnetization (SIRM). In
89 contrast, the thermal demagnetization behaviors of SRM and PRM have not yet been studied in
90 detail, with the exception of some preliminary analyses of FeNi-bearing lunar materials [*Cisowski*
91 *et al., 1973; Gattacceca et al., 2010b*].

92 Because nearly all meteorites and rocks from cratered planetary surfaces (including the
93 lunar samples from the Apollo missions) have experienced some level of shock, it is important to
94 understand the effects of shock on remanent magnetization, especially at relatively low pressures
95 where petrographic evidence of shock may not be observed (< 5 GPa [*Stoffler et al., 2006*]). The
96 magnetization of rocks submitted to pressures < 2 GPa is of particular interest because the volume
97 of target rocks shocked to < 2 GPa during hypervelocity impacts is ~ 2 -3 times the volume of target
98 rocks shocked to pressures > 2 GPa (estimated from *Robertson and Grieve [1977]* and *Louzada*
99 *and Stewart [2009]*). In this study, we determine the relative intensities of PRM acquired at
100 pressures < 2 GPa compared to TRM for samples representing a wider range of rock types and

101 ferromagnetic mineralogies than previously studied. In addition, we investigate the thermal
102 demagnetization properties of PRM and compare them to those of other remanences. We utilize
103 our results to assess the paleomagnetic stability of PRM and SRM over geologic timescales and to
104 provide a framework for identifying SRM in natural samples.

105

106 2. Samples and Methods

107 2.1. Samples and handling

108 We analyzed the PRM properties of a variety of terrestrial and extraterrestrial samples with
109 different magnetic mineralogies and rock magnetic properties including FeNi alloys, magnetite,
110 pyrrhotite, titanomagnetite, and combinations of these ferromagnetic minerals (Table 1). Our
111 terrestrial samples include a titanomagnetite-bearing Pleistocene basalt (BB) from Chanteuges,
112 Haute-Loire, France and a magnetite-bearing microdiorite (EE) from the Esterel range, France,
113 whose rock magnetic properties were previously characterized by *Gattacceca et al.* [2007; 2008].
114 We also studied a Mesoproterozoic titanomagnetite-bearing diabase (DeI3-6) collected from a dike
115 associated with the Osler Volcanic Group [*Swanson-Hysell et al.*, 2014] within the Slate Islands
116 impact crater, Canada. Numerous extraterrestrial samples were also analyzed. Our magnetite- and
117 pyrrhotite-bearing samples include the Martian meteorite Tissint [*Gattacceca et al.*, 2013], the
118 CV3 carbonaceous chondrite Allende (e.g., *Carporzen et al.* [2011]), the Rumuruti-like (R)
119 chondrite PCA 91002, and the CM carbonaceous chondrites Cold Bokkeveld, Mighei, Murchison,
120 Murray, Nogoya, and Paris [*Cournede et al.*, 2015]. We also analyzed the FeNi-bearing mare
121 basalts 15556 [*Tikoo et al.*, 2012], 12022 [*Tikoo et al.*, 2014], 10020 [*Shea et al.*, 2012], 10017
122 and 10049 [*Suavet et al.*, 2013], ordinary chondrites NWA 6490 and NWA 7621, and basaltic
123 eucrite ALHA81001 [*Fu et al.*, 2012]. Finally, we studied the magnetite-bearing CV3 chondrite
124 Kaba and the pyrrhotite-bearing R chondrite LAP 03639. Sample handling, PRM acquisition
125 experiments, and rock magnetic experiments were conducted within magnetically shielded rooms
126 (ambient DC field < 250 nT) in paleomagnetism laboratories at the Massachusetts Institute of
127 Technology (MIT), CEREGE (Aix-en-Provence, France), and the University of California,
128 Berkeley.

129

130 2.2. PRM acquisition

131 Specimens ranging in mass from 30-350 mg were imparted with a PRM using a nearly
132 nonmagnetic pressure cell in the presence of a controlled laboratory field [*Gattacceca et al.*,
133 2010b]. Prior to PRM acquisition, the samples were demagnetized using AF or thermal
134 demagnetization. We were able to fully demagnetize (i.e., residual magnetization was <95% of the
135 original NRM) all samples except for LAP 03639 (residual was ~15% of the original NRM) and
136 DeI3-6 (which acquired spurious gyroremanent magnetization during AF demagnetization with
137 intensity ~10% of the original NRM). Any residual magnetizations present were removed from
138 the PRM data by vector subtraction. Following demagnetization pre-treatment, specimens were
139 placed in an 8 mm × 20 mm Teflon capsule and submerged in polyethylsiloxane fluid. The capsule
140 was then placed in a nearly nonmagnetic piston-cylinder pressure cell made of the alloy
141 Ni₅₇Cr₄₀Al₃ [*Sadykov*, 2008]. The cell has a magnetic moment of 2×10^{-8} Am² and is designed to
142 allow hydrostatic loading up to 1.8 GPa. A solenoidal coil wrapped around the pressure cell was
143 used to produce a dc magnetic field oriented along the long axis of the cell. Field intensities were
144 calibrated using a Hall probe. A known magnetic field (500 μT, 750 μT, or 800 μT) was applied
145 to the cell. The cell was then loaded with pressures ranging between 0.18 and 1.8 GPa using a
146 Specac 15-ton manual hydraulic press. Specimens were held at pressure for ~1 minute and then
147 the load was released in the presence of the applied magnetic field.

148

149 2.3. Magnetic analyses

150 Following PRM acquisition, we measured the acquired magnetization and then demagnetized
151 specimens using either stepwise AF or thermal methods. To compare the demagnetization behavior
152 of PRM to that of other forms of remanence, additional specimens from each parent rock sample
153 (subjected to the same pre-treatment as the PRM specimen) were given TRM, ARM, and/or SIRM.
154 ARM was applied with a 0.1 mT dc bias field and a 290-300 mT ac field. SIRM was applied in a
155 pulse field of 900 mT for FeNi and magnetite-bearing samples and a 3 T field for pyrrhotite-
156 bearing samples. Measurements of magnetization, progressive demagnetization, and rock
157 magnetic experiments were carried out using 2G Enterprises Superconducting Rock
158 Magnetometers at MIT, CEREGE, or UC Berkeley. The MIT and UC Berkeley magnetometers
159 are equipped with automated sample handling and AF demagnetization equipment [*Kirschvink et*
160 *al.*, 2008]. Thermal demagnetization was conducted in ASC Scientific ovens with residual
161 magnetic fields <5 nT. Nearly all thermally demagnetized samples were given an AF pre-treatment

162 of 1.5 mT prior to thermal demagnetization to remove any weak isothermal remanent
163 magnetization (IRM) that may have been acquired from the pressure cell solenoid (500-800 μ T dc
164 field). Two exceptions to this protocol were made for the ordinary chondrites NWA 6490 and
165 NWA 7629 due to their exceptionally low coercivities (they lose ~40% of ARM by AF 1.5 mT).
166 FeNi-bearing samples were thermally demagnetized in a controlled oxygen fugacity atmosphere
167 using a calibrated H₂-CO₂ mixture at MIT to avoid alteration of the magnetic carriers [*Suavet et*
168 *al.*, 2014]. Hysteresis properties were measured on Princeton Instruments vibrating sample
169 magnetometers at CEREGE and the Institute for Rock Magnetism at the University of Minnesota.

170
171

172 3. Results

173

174 3.1. PRM acquisition

175 We define PRM acquisition efficiency (α) as the ratio of PRM to TRM acquired in the
176 same ambient field. One of the goals of our study was to determine how α varies as a function of
177 ferromagnetic mineralogy, domain state, and pressure. Observations from a wide range of samples
178 and ferromagnetic mineralogies suggest the following generalized relationship between TRM and
179 SIRM:

180

$$181 \quad \text{TRM} \approx \frac{\text{SIRM} \cdot B}{3000 \mu\text{T}} \quad (1)$$

182

183 where B is the strength of the ambient field in μ T [*Kletetschka et al.*, 2003; *Gattacceca and*
184 *Rochette*, 2004]. Numerous studies indicate that this relationship is generally accurate to within a
185 factor of ~2-3 [e.g., *Kletetschka et al.*, 2003; *Gattacceca and Rochette*, 2004; *Tikoo et al.*, 2014;
186 *Weiss and Tikoo*, 2014]. To avoid thermochemical alteration from heating, we did not impart the
187 PRM specimens with laboratory TRM. Instead, we used equation (1) and PRM intensities from
188 our acquisition experiments to estimate α (i.e., PRM/TRM) for all samples. We normalized PRM
189 efficiency data from after AF demagnetization to 2 mT ($\alpha_{2 \text{ mT}}$) to remove any viscous contributions
190 imparted by the pressure cell solenoid from the PRM data. In a natural setting, SRM or PRM
191 acquired by such a low-coercivity (<2 mT) fraction of ferromagnetic grains would likely be

192 eradicated by other secondary processes such as the acquisition of viscous remanent magnetization
193 (VRM). Therefore, we do not anticipate that our overall conclusions regarding PRM and SRM
194 efficiency in nature would change substantially by the exclusion of PRM data at these lowest AF
195 levels.

196 PRM acquisition efficiency ($\alpha_{2\text{ mT}}$) generally increased with peak pressure for all samples.
197 However, we observed that different samples had vastly different efficiencies for the same pressure
198 level, ranging between negligible PRM acquisition for ALHA81001 ($\alpha_{2\text{ mT}} = 5 \times 10^{-5}$) to substantial
199 PRM acquisition for EE ($\alpha_{2\text{ mT}} = 0.22$) at a peak pressure of 1.8 GPa (Fig. 1). We observed that
200 samples with lower remanent coercivities (B_{cr}) typically had higher $\alpha_{2\text{ mT}}$ values than samples with
201 lower B_{cr} values (Fig. 2), consistent with the observation that PRM is preferentially acquired by
202 low coercivity grains. Considering $\alpha_{2\text{ mT}}$ values in conjunction with the hysteresis data suggests
203 that rocks with larger populations of MD grains have higher PRM efficiencies, at least for samples
204 containing a single ferromagnetic mineralogy (e.g., only magnetite or only FeNi) (Fig. 3). For
205 example, the magnetite-bearing microdiorite EE and titanomagnetite-bearing basalt BB have
206 higher PRM efficiencies than the more SD-like (i.e., having higher M_{rs}/M_s and lower B_{cr}/B_c)
207 samples DeI3-6 and Kaba (Figs. 2 and 3). Similarly, among FeNi-bearing samples, the MD lunar
208 basalts and ordinary chondrite NWA 6490 have higher PRM efficiencies than the more SD-like
209 eucrite ALHA81001. However, this relationship between PRM efficiency and domain state is
210 difficult to determine when comparing samples with more than one magnetic carrier mineral (such
211 as rocks with both magnetite and pyrrhotite). In such cases, differences in both PRM efficiency
212 and bulk hysteresis properties may be associated with variations in the relative concentrations of
213 each ferromagnetic mineral present between samples (Section 4.3).

214

215

216 3.2. AF demagnetization of PRM

217 We conducted AF demagnetization experiments of laboratory PRM acquired at a range of
218 pressures ≤ 1.8 GPa on at least one specimen from each rock studied. Consistent with previous
219 experiments focused on SRM demagnetization behavior [Gattacceca *et al.*, 2007; Gattacceca *et*
220 *al.*, 2008; Gattacceca *et al.*, 2010a], we found that PRM was confined to lower AF levels (< 20 -50
221 mT, depending on the sample) than SIRM, TRM, and ARM. The median destructive field (MDF:
222 the AF amplitude required to remove half of a remanence) of PRM increased with applied pressure,

223 but always remained lower than those of TRM, ARM, and SIRM at the pressures studied (Fig. 4).
224 The AF levels necessary to remove the laboratory-induced remanences were correlated with the
225 domain states of the samples. For example, PRM was removed more efficiently from the
226 multidomain remanence carriers of the microdiorite sample EE than from the remanence carriers
227 of diabase sample DeI3-6, which have rock magnetic behavior characteristic of pseudo-single
228 domain grains, even though low-titanium magnetite is the primary magnetic carrier for both
229 samples. Therefore, our PRM acquisition and AF demagnetization results both suggest that PRM
230 is preferentially acquired by low coercivity, multidomain grains.

231

232 3.3. *Thermal demagnetization of PRM*

233 We conducted thermal demagnetization of PRM on selected specimens representing each
234 group of ferromagnetic mineralogies. We found that for all samples, PRM persisted to unblocking
235 temperatures approaching the Curie temperatures (or at least the maximum unblocking
236 temperatures of SIRM) of the ferromagnetic minerals (Fig. 5). For example, the magnetite-bearing
237 samples did not lose 95% of the PRM overprint until they were heated to temperatures >500 °C
238 (Fig. 5a, b, c, f). The FeNi-bearing ordinary chondrites NWA 6490 and NWA 7629 lost 95% of
239 their PRM at ~500-550 °C (Fig. 5d). While this temperature is well below the 780 °C Curie point
240 of kamacite ($\text{Fe}_{0.95-1}\text{Ni}_{0-0.05}$), the fact that the SIRM demagnetizes at the same low temperature
241 indicates that these samples either experienced thermochemical alteration during heating or that
242 the remanence carriers in these samples are made of other FeNi alloys with higher Ni contents
243 such as martensite ($\text{Fe}_{0.75-0.95}\text{Ni}_{0.05-0.25}$) which could demagnetize at similarly low temperatures
244 depending on Ni content [Swartzendruber *et al.*, 1991]. In all cases, PRM (acquired at pressures
245 up to 1.8 GPa) had lower median unblocking temperatures than TRM or ARM although this
246 difference was less pronounced than that seen in the MDF values in the AF demagnetization data.
247 The median unblocking temperatures of PRM generally increased with pressure, but did not
248 exceed those of SIRM, which also had a lower median unblocking temperature than TRM and
249 ARM (Fig. 5).

250

251

252 4. Discussion

253 4.1. Demagnetization properties of pressure-induced remanence

254 Although PRM is easily removed at relatively low AF levels compared to other forms of
255 remanence such as full ARM and TRM (Fig. 4), we found that during thermal demagnetization
256 PRM (and SRM) may persist and overlap with higher coercivity magnetizations over nearly the
257 full range of unblocking temperatures in shocked samples (Fig. 6). This result confirms that AF
258 demagnetization methods are more efficient at removing PRM and SRM overprints from rocks
259 than thermal demagnetization. Therefore, if thermal demagnetization is conducted without prior
260 AF pre-treatment, both the primary remanence and any present SRM overprints could be removed
261 simultaneously. Paleomagnetic studies aiming to retrieve paleointensities from Thellier-Thellier
262 style experiments or other thermal methods from the primary (pre-shock) remanence of shocked
263 samples should ideally include an AF pre-treatment prior to thermal demagnetization to ensure
264 that any putative SRM overprints are identified and cleaned from samples properly.

265 Alternatively, if the goal of a paleomagnetic study is to test whether or not a secondary
266 impact-related remanence is SRM, the distinct AF and thermal demagnetization behavior of SRM
267 (based on its analog, PRM) provides a framework for distinguishing SRM from other forms of
268 remanence such as VRM from long-term exposure to the terrestrial field or thermoviscous
269 remanent magnetization (TVRM) from heating produced in impact settings as a result of
270 significant shock pressures [Stewart *et al.*, 2007]. For relatively high Curie temperature
271 ferromagnetic minerals such as near-stoichiometric magnetite (~580 °C), both VRM and TVRM
272 would likely be removed well below the Curie temperatures during thermal demagnetization in
273 SD and pseudo-single domain (PSD) samples, whereas SRM would persist to higher temperatures.
274 Distinguishing SRM from other remanences may also be challenging for predominantly MD
275 samples because unblocking tail effects could potentially cause all of these forms of remanence to
276 not fully demagnetize until near the Curie temperature [Xu and Dunlop, 1994].

277

278 4.2. Mechanism(s) behind PRM and SRM acquisition and implications for their paleomagnetic 279 stability

280 Given the relative ease of removing PRM and SRM using AF demagnetization, the
281 persistence of PRM to relatively high unblocking temperatures during thermal demagnetization

282 experiments requires explanation. In this section, we discuss how this behavior results from the
283 various mechanisms by which PRM and SRM are acquired. Following the treatment of *Dunlop et*
284 *al.* [1969] for magnetization acquired under uniaxial compression, we first discuss how PRM and
285 SRM may be acquired by and preserved in SD grains according to Néel theory [*Néel*, 1955]. We
286 then discuss the acquisition and preservation of these remanences in MD grains.

287

288 4.2.1. Single domain samples

289 Ferromagnetic grains preferentially retain magnetization along certain directional axes
290 within crystals called easy axes. Grains are remagnetized when the energy barrier preserving an
291 initial magnetization is overcome such that magnetization is re-acquired along a different easy axis
292 or in an antipodal direction along the same axis. The net anisotropy energy of a grain is the sum
293 of the magnetocrystalline, magnetostriction (shape), and magnetoelastic (stress) anisotropy
294 energies [*Dunlop and Ozdemir*, 1997]. Several changes in magnetic anisotropy have been observed
295 to occur when rocks are pressurized:

296 When ferromagnetic grains are hydrostatically compressed, the constants of
297 magnetocrystalline anisotropy (K_1 and K_2 for a cubic crystal structure) have been observed to
298 decrease with increasing pressure while the magnetostriction constants (λ_{100} and λ_{111}) increase with
299 pressure for magnetite [*Nagata and Kinoshita*, 1967]. These constants change at different rates in
300 response to pressure (K_1 and K_2 decrease less rapidly than λ_{100} and λ_{111} increase with pressure)
301 [*Nagata and Kinoshita*, 1967]. Therefore, even though there is no preferred compression axis, the
302 resulting change in total anisotropy can lead to remagnetization.

303 Uniaxial compression experiments demonstrate that remanence anisotropy and magnetic
304 susceptibility strengthen in the direction perpendicular to a uniaxial compression axis and weaken
305 along the axis parallel to the compression (e.g., *Nagata* [1970] and *Gilder and Le Goff* [2008]).
306 These changes indicate that uniaxial compression introduces stress anisotropy to ferromagnetic
307 grains. As pressure increases, the contribution of stress anisotropy energy increases relative to the
308 magnetocrystalline and shape anisotropy energies (that are simultaneously changing as a result of
309 compression, as discussed in the hydrostatic case above). Considering only shape anisotropy, the

310 total energy for a spheroidal SD grain under a uniaxial stress σ applied parallel to the elongation
 311 axis is:

$$312 \quad E_{tot} = -\mu_0 V \overline{M}_s \cdot \overline{H}_0 + \frac{1}{2} \mu_0 V M_s \left[(N_b - N_a) M_s - \frac{3\lambda_s \sigma}{\mu_0 M_s} \right] \sin^2 \theta \quad (2)$$

313 where μ_0 is the permeability of free space, V is the grain volume, \overline{M}_s is the spontaneous
 314 magnetization, \overline{H}_0 represents the applied field, N_a and N_b are the demagnetizing factors when \overline{M}_s
 315 is oriented parallel or perpendicular to the long axis of the grain, λ_s is the magnetostriction (i.e.,
 316 the magnetization-induced change in shape of a grain), and θ is the angle that \overline{M}_s is rotated away
 317 from the easy axis by \overline{H}_0 (Equation 16.10 of *Dunlop and Ozdemir* [1997]). The expression within
 318 the brackets represents the microcoercivity (i.e., the critical field above which the spontaneous
 319 magnetization in a grain will undergo an irreversible rotation to another stable orientation).

320 For shape anisotropy alone, in the absence of pressure, the microcoercivity of a magnetic
 321 grain, H_K , is equal to $(N_b - N_a)M_s$. Adding stress parallel to the elongation axis of a spheroidal
 322 reduces the microcoercivities of magnetic grains to a new value, $H'_K = H_K - \frac{3\lambda_s \sigma}{\mu_0 M_s}$ (Equation 16.11
 323 of *Dunlop and Ozdemir* [1997]). In contrast, applying uniaxial stress perpendicular to the
 324 elongation axis will result in an increase in coercivity. Application of stress at intermediate angles
 325 would shift the anisotropy from a uniaxial to a non-uniaxial form. The effect of stress on
 326 microcoercivity suggests that imparting SRM may be analogous to imparting IRM in the absence
 327 of pressure. During decompression, the bulk anisotropy of a grain will progressively return to its
 328 natural (stress-free) state. If an ambient magnetic field is present, the grain will be remagnetized
 329 as the spontaneous magnetization aligns itself with the anisotropy easy axis that has the lowest
 330 angular deviation from the field direction.

331 During TRM acquisition, SD grains acquire remanent magnetization as they cool through
 332 their respective blocking temperatures. Blocking temperatures vary depending on grain volume
 333 and microcoercivity. In contrast, as demonstrated above, acquisition of other forms of remanence
 334 such as IRM, PRM, and SRM are principally dependent on coercivity rather than grain volume.
 335 The recording of PRM and SRM by coercivity may explain why these remanences are removed
 336 more efficiently than TRM by AF demagnetization and also why the thermal demagnetization
 337 curves of PRM and IRM qualitatively resemble each other more than they resemble TRM (Fig. 5).

338 The persistence of PRM and SRM to nearly the Curie temperature during thermal demagnetization
339 experiments likely occurs because the blocking/unblocking temperature distribution in rocks is
340 skewed towards the Curie temperature (see Fig. 8.15 of *Dunlop and Ozdemir* [1997]). The
341 hyperbolic nature of blocking temperature contours demonstrates that large grains with low
342 microcoercivities can have equally high unblocking temperatures as smaller grains with higher
343 coercivities. As such, it is possible for a relatively low coercivity magnetization such as PRM or
344 SRM to persist up to temperatures approaching the Curie temperature.

345

346

347 4.2.2. *Pseudo-single domain and multidomain samples*

348 While the theoretical frameworks for characterizing SD remanence and demagnetization
349 properties are well-described, rocks with purely SD magnetic grains are rare in natural settings.
350 PSD and MD grains, which make up the majority of magnetic carriers in most natural samples, do
351 not strictly adhere to the predictions of Néel theory. Rather than by the rotation of spontaneous
352 magnetization (which occurs for SD grains), MD grains magnetize and demagnetize by motions
353 and pinning of domain walls in their interiors. Translation of domain walls requires little energy
354 and can be accomplished in relatively low fields [*Dunlop and Ozdemir*, 1997]. As a result, MD
355 grains are characterized by low coercivities. Our data reveal that samples with low coercivity MD
356 grains have a much higher acquisition efficiency of PRM than samples with a greater concentration
357 of SD grains (Figs. 2,3). This is exemplified by the 3 orders of magnitude difference in $\alpha_{2\text{mT}}$ values
358 between the MD ordinary chondrites and lunar samples ($\sim 10^{-2}$) and that of the eucrite ALHA81001
359 ($\sim 10^{-5}$), which has a substantial population of SD grains (Fig. 3). In general, SD grains are unlikely
360 to carry significant amounts of PRM or SRM, at least for the range of pressures explored in this
361 study.

362 Our results demonstrate that PRM and SRM are dominantly acquired via stress-induced
363 motions of domain walls in MD grains. In PSD and MD grains, PRM may be acquired by the
364 stress-induced nucleation of domain walls [*Boyd et al.*, 1984]. Domain walls stabilize at local
365 energy minima that are often correlated with domain wall pinning localities within the crystal
366 structure [*Muxworthy and Williams*, 2006]. Due to the low coercivity nature of MD grains, domain

367 wall configurations associated with PRM and SRM are easily disrupted by AF demagnetization.
368 This allows PRM and SRM to be removed at relatively low AF levels compared to other forms of
369 remanence such TRM and SIRM, which occupy grains spanning the entire range of coercivities
370 present in a sample. During thermal demagnetization, MD remanence will persist until
371 temperatures are high enough that thermal fluctuations are sufficiently large for domain walls to
372 move to new local energy minima [Muxworthy and Williams, 2006]. Another factor that may
373 contribute to the persistence of PRM to high unblocking temperatures is that, in contrast to SD
374 grains, MD grains do not have discrete unblocking temperatures [Dunlop and Ozdemir, 2000;
375 2001]. Laboratory experiments showed that partial TRM (pTRM) imparted to MD magnetite-
376 bearing samples between 370 °C and 350 °C began to demagnetize well below the unblocking
377 temperature predicted by Néel theory, $T_B = 350$ °C, and that ~90% of the pTRM was not removed
378 until >150 °C above T_B [Dunlop and Ozdemir, 2000; 2001]. Indeed, for the largest MD grains,
379 laboratory pTRM was not completely removed until the Curie temperature. For comparison, ~90%
380 of laboratory pTRM imparted to SD samples was removed by temperatures of only ~30 °C above
381 T_B . Therefore, the persistence of PRM and SRM to unblocking temperatures approaching the Curie
382 temperature during thermal demagnetization experiments may also be related to the presence of
383 unblocking tails. In summary, the low coercivity nature of MD grains, coupled with the persistence
384 of some domain wall configurations until elevated temperatures during thermal demagnetization,
385 explains why PRM and SRM are more efficiently removed using AF rather than thermal methods
386 in MD samples.

387

388 4.3. *Relationships between PRM efficiency and properties of ferromagnetic minerals*

389 In Section 3.1, we demonstrated that for samples containing a single ferromagnetic
390 mineralogy, rocks with larger populations of MD grains have higher PRM efficiencies than more
391 SD-like samples. However, this trend was not as apparent in rocks which contained mixtures of
392 magnetite and pyrrhotite. It is possible that some variability (factor ~2-3) in PRM efficiency values
393 may be attributed to uncertainties in the SIRM normalization calibration constants used to estimate
394 TRM [equation (1)]. However, the primary factor is likely related to the compositions, domain
395 states, grain defect concentrations, and relative abundances of the dominant remanence carriers in
396 a given sample. Among the CM chondrites, magnetic susceptibilities vary by up to ~2 orders of

397 magnitude [Rochette *et al.*, 2008], suggesting the presence of substantial variations in
398 ferromagnetic mineral assemblages, concentrations, and grain sizes even within this one group.
399 Therefore, it is difficult to determine domain states and, in turn, compare PRM efficiency to
400 domain state variabilities that are occurring in multiple phases.

401 Thermal demagnetization curves of laboratory-induced SIRM may elucidate this issue. For
402 Tissint, ~50% of the initial SIRM remains after the sample is heated above the ~320 °C pyrrhotite
403 Curie temperature (Fig. 2b), whereas only ~10-20% of the initial SIRM remains at the same
404 temperature for all studied CM chondrites (see Fig. 4 of *Cournède et al.* [2015]). The coercivity
405 of magnetite (maximum value ~300 mT) is generally far lower than that of pyrrhotite (maximum
406 value >1 T). Therefore, the higher contribution of (low coercivity) magnetite to the net remanence
407 of Tissint relative to the CM chondrites may explain its higher PRM efficiency.

408 We also observed that MD FeNi samples had similar PRM efficiencies as PSD magnetite-
409 bearing samples. This raises the possibility that magnetite may have different magnetoelastic
410 properties than FeNi alloys. The Poisson ratios (negative ratio of transverse to axial strain) of
411 these minerals are similar: the mean ratio for magnetite is 0.31 [*Chicot et al.*, 2011], whereas the
412 ratios for metallic Fe and FeNi alloys are ~0.28 [*Ledbetter and Reed*, 1973], suggesting that
413 differences in bulk elastic properties are likely not responsible for differences in PRM efficiency
414 between these minerals. Poisson ratios for pyrrhotite range between ~0.12-0.3 [*Louzada et al.*,
415 2010], suggesting that pyrrhotite may have a different elastic response to pressure than magnetite
416 and FeNi alloys. However, significant differences do exist between the magnetic anisotropy
417 coefficient values for magnetite and FeNi. For iron at room temperature and atmospheric pressure,
418 the first term of the magnetocrystalline anisotropy constant (K_1) is 4.8×10^4 J/m³ and the
419 polycrystalline magnetostriction constant at saturation (λ) is -7×10^{-6} . For room-temperature
420 magnetite, $K_1 = -1.35 \times 10^4$ J/m³ and $\lambda = 35.8 \times 10^{-6}$ [*Dunlop and Ozdemir*, 1997]. As discussed in
421 Section 4.2.1, K_1 decreases with pressure while λ increases with pressure. As the total anisotropy
422 of a grain is the sum of its shape, magnetocrystalline, and stress anisotropies, the greater
423 contribution of stress anisotropy relative to magnetocrystalline anisotropy for magnetite (as
424 compared to FeNi) may explain its higher PRM efficiency.

425

426 *4.4. Identifying SRM in natural samples*

427 While many attempts have been made to identify SRM in natural samples [*Robertson,*
428 *1967; Halls, 1979; Jackson and Van der Voo, 1986; Fuller and Cisowski, 1987; Iseri et al., 1989;*
429 *Schmidt and Williams, 1991; Pesonen et al., 1999; Carporzen and Gilder, 2006; Elbra et al., 2007;*
430 *Kontny et al., 2007; Louzada et al., 2008; Raiskila et al., 2011; Carporzen et al., 2012*], reports of
431 confirmed SRM in studied impact craters and extraterrestrial samples are rare to nonexistent. A
432 key question is why SRM has not yet been conclusively identified. An important implication of
433 this study is that SRM may not be readily observed in natural samples because of two factors: [1]
434 SRM may be overprinted by other secondary remanences such as VRM, shock heating or
435 metamorphic TVRM, IRM, or chemical remanent magnetization (CRM) from the creation of new
436 ferromagnetic minerals during post-impact hydrothermal activity [*Quesnel et al., 2013*], and [2]
437 the acquisition efficiency of SRM may be too low for such a magnetization to be distinguishable
438 from an underlying primary remanence (such as primary TRM) or a coexisting secondary
439 remanence.

440 Regarding factor [1], the low coercivity nature of SRM means that it is highly susceptible
441 to IRM overprinting by exposure to, for example, magnets or lightning strikes (e.g., *Carporzen et*
442 *al. [2012]*). It is also possible that SRM is not often observed because the remanence may be at
443 least partially eradicated by viscous relaxation over time. Although SRM is predominantly
444 acquired by multidomain grains, prior observations indicate that single domain and multidomain
445 rocks have similar susceptibilities to viscous acquisition and relaxation [*Dunlop, 1983; Yu and*
446 *Tauxe, 2006*] while PSD grains are less susceptible to viscous effects [*Dunlop, 1983*]. Therefore,
447 the persistence of PRM to unblocking temperatures approaching the Curie temperatures of the
448 magnetic carriers suggests that at least some portion of an acquired SRM may be stable for billions
449 of years according to the predictions of Néel theory [*Pullaiah et al., 1975; Dunlop et al., 2000;*
450 *Weiss et al., 2000; Garrick-Bethell and Weiss, 2010*]. In contrast, TVRM acquired from heating to
451 a few hundred °C would likely eradicate much of any pre-existing SRM, given the prevalence of
452 blocking and unblocking tails in PSD and MD samples. Secondary magnetizations at several
453 impact craters have been attributed to shock-induced TVRM [*Jackson and Van der Voo, 1986;*
454 *Iseri et al., 1989; Schmidt and Williams, 1991; Elbra et al., 2007*].

455 Regarding factor [2], $\alpha_{2\text{ mT}}$ was $\leq \sim 10^{-2}$ (i.e., the 1.8 GPa PRM intensity was $\leq 1\%$ of an
456 equivalent-field TRM intensity) for 17 out of the 19 samples for which we determined PRM
457 efficiencies. This means that a low-intensity SRM component may be difficult to identify if much

458 stronger primary or other secondary remanences are present. However, when the paleofield
459 strength range can be roughly estimated (like for terrestrial rocks), the disparity in the acquisition
460 efficiencies of SRM and thermally activated forms of remanence such as VRM or TVRM provides
461 an avenue with which to distinguish between them in natural samples by conducting paleointensity
462 experiments. When conducting paleointensity experiments that assume a thermal origin of
463 remanence in samples, thermally activated forms of remanence should yield paleointensities on
464 order of the expected paleofield strength at the time of the impact. In contrast, SRM would yield
465 paleointensities at minimum an order of magnitude weaker than the expected paleofield due to its
466 low acquisition efficiency (at least for the ≤ 2 GPa pressure range investigated in this study).

467

468 5. Conclusions

469

470 Our results indicate that an impact-related magnetization component may be attributable
471 to (< 2 GPa) SRM if *(i)* it can be efficiently cleaned via AF demagnetization, *(ii)* it persists to near-
472 Curie unblocking temperatures during thermal demagnetization, and *(iii)* its inferred paleointensity
473 (determined in a paleointensity experiment assuming a thermal origin of remanence) is
474 significantly weaker than that of the magnetic field in which it was acquired (although the
475 paleofield intensity is unlikely to be known a priori for extraterrestrial samples). However, the low
476 acquisition efficiency and low coercivity nature of PRM relative to other forms of remanence may
477 result in any record of SRM being obscured by other magnetization components. Furthermore,
478 other impact-related processes such as hydrothermal alteration or shock heating may produce
479 additional magnetizations that could overprint pre-existing SRM. Therefore, conclusive
480 identification of SRM in natural samples would likely require studying rocks that have a fortuitous
481 combination of relatively high PRM acquisition efficiencies ($> \sim 10\%$ of TRM), minimal post-
482 impact hydrothermal alteration, and low peak shock pressures (to preclude significant TVRM from
483 shock heating). We are not currently aware of any unambiguous natural examples of such samples.

484

485 Acknowledgements

486

487

488 Data from this study is available upon request to the corresponding author. We thank the JSC staff
489 and the Curation and Analysis Planning Team for Extraterrestrial Materials for allocating our lunar
490 samples. We thank B. Carbone for administrative support. We thank two anonymous reviewers
491 for their constructive comments. S. M. T., B. P. W., and C. S. thank the Brown-Massachusetts
492 Institute of Technology (MIT) NASA Lunar Science Institute and Solar System Exploration
493 Research Virtual Institute, the NASA Lunar Advanced Science and Exploration Program, and the
494 NASA Solar System Workings Program.. B. P. W. and J. G. thank the MIT-France Seed Funds
495 Program, the Projet International de Coopération Scientifique Program, and the People Programme
496 (Marie Curie Actions) of the European Union under Research Executive Agency Grant 298355.
497 B. P. W. thanks the Miller Institute for Basic Research in Science for support. N. L. S.-H. and J.
498 G. thank the France-Berkeley Fund program. N. L. S.-H. and S. M. T. were supported by NSF
499 Division of Earth Sciences Grant 1316395. S. M. T. was also supported by a Visiting Fellowship
500 at the Institute for Rock Magnetism, University of Minnesota.

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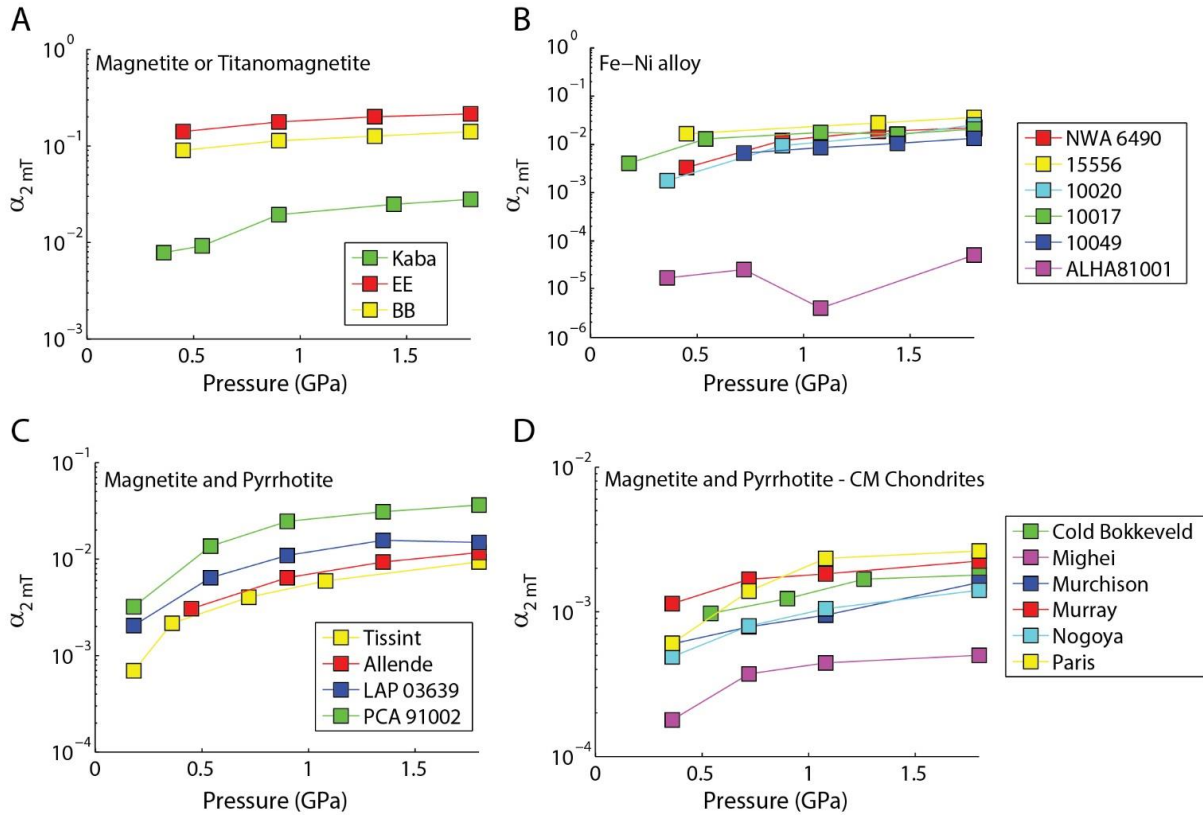


Fig. 1. PRM efficiency relative to TRM at AF 2 mT (α_2 mT) at pressures ranging up to 1.8 GPa for samples of various magnetic mineralogies. (A) Samples with magnetite and titanomagnetite. (B) Samples with Fe-Ni alloy. (C) Samples with pyrrhotite alone or a mixture of magnetite and pyrrhotite. (D) CM chondrite samples with magnetite and pyrrhotite.

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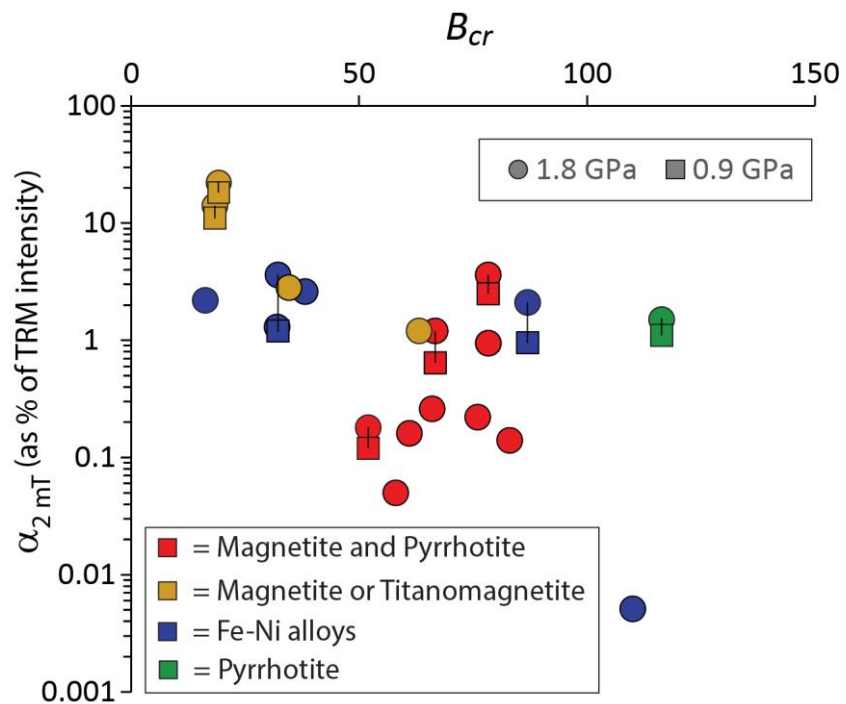


Fig.2. PRM efficiency versus remanent coercivity. The ordinate gives the PRM efficiency relative to TRM at an AF level of 2 mT ($\alpha_{2 \text{ mT}}$), expressed here as a percentage of the intensity of TRM acquired in an equivalent ambient field, for PRM applied at 1.8 GPa (dark gray circles) and 0.9 GPa (light gray squares). The abscissa gives the remanent coercivity (B_{cr}). PRM data collected from the same subsamples at different pressures are joined by lines.

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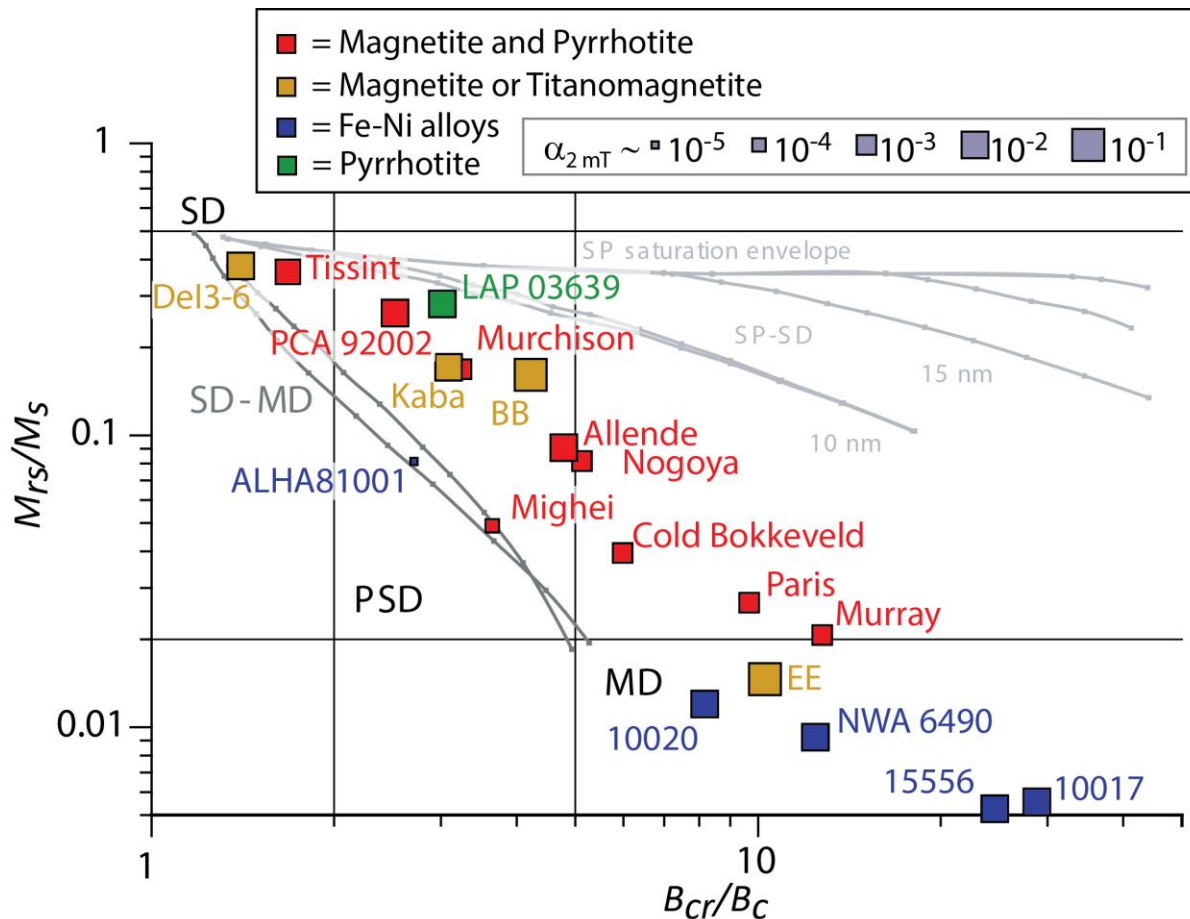


Fig. 3. Dunlop-Day plot of hysteresis parameters. The ordinate gives the magnitude of the saturation remanent magnetization (M_{rs}) divided by the magnitude of the saturation magnetization (M_s). The abscissa gives the remanent coercivity (B_{cr}) divided by the coercive force (B_c). Squares denote sample positions. Red symbols denote samples with a combination of magnetite and pyrrhotite ferromagnetic mineralogies. Yellow symbols denote samples with magnetite or titanomagnetite. Blue symbols denote samples with Fe-Ni alloys. Green symbols denote samples with pyrrhotite. Symbol sizes are scaled according to their PRM efficiency ($\alpha_{2\text{ mT}}$). Straight black vertical and horizontal lines divide the plot into rectangular regions representing single domain (SD), pseudo-single domain (PSD), and multidomain (MD) regimes.

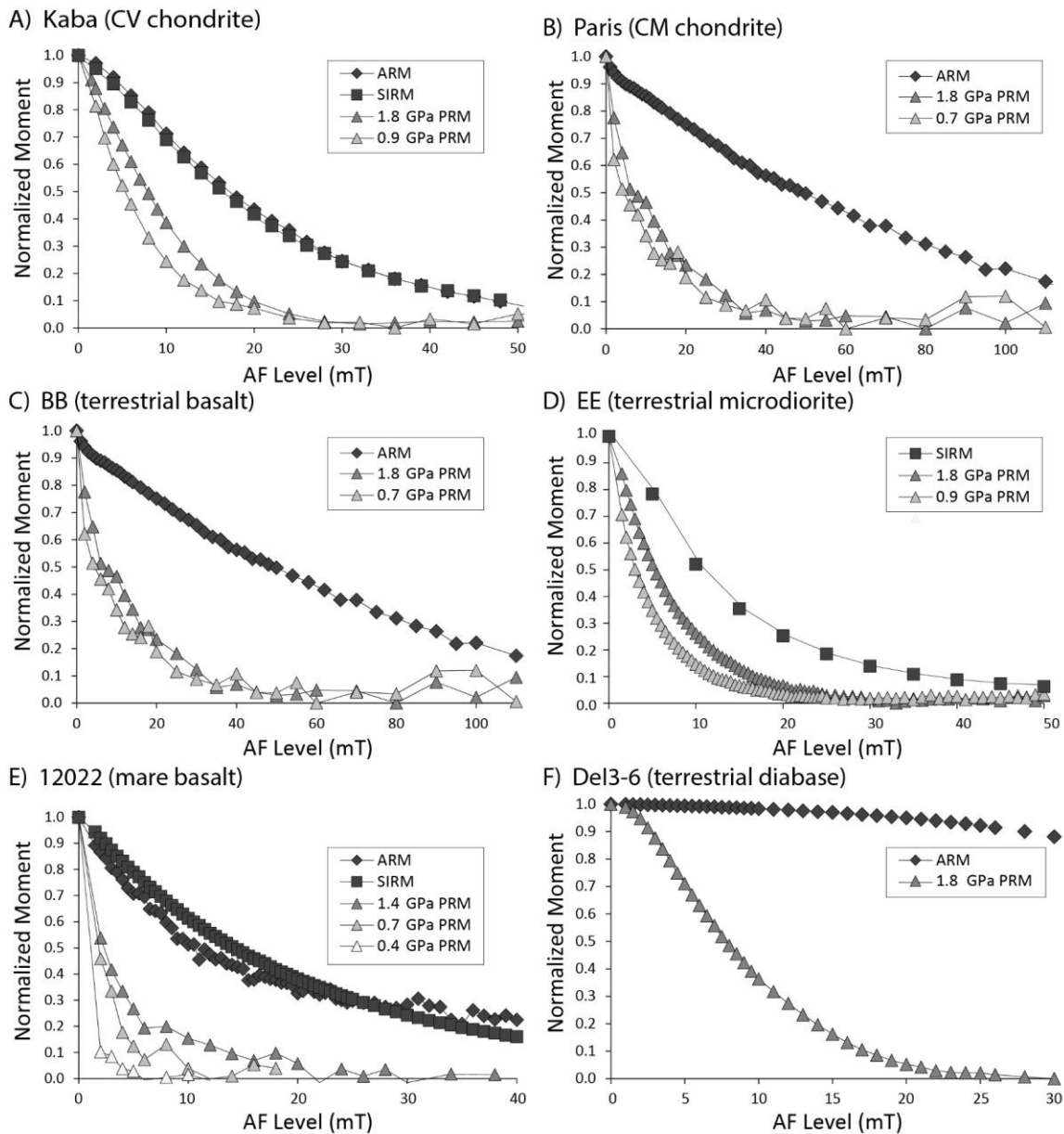


Fig. 4. AF demagnetization of PRM (triangles), TRM (circles), ARM (diamonds), and/or SIRM (squares). The ordinate shows the normalized magnetic moment and the abscissa shows the corresponding AF level. Shown samples (magnetic mineralogies) include: (A) CV3 chondrite Kaba (magnetite), (B) CM chondrite Paris (magnetite and pyrrhotite), (C) terrestrial basalt BB (titanomagnetite), (D) terrestrial microdiorite EE (magnetite), (E) mare basalt 12022 (FeNi), and (F) terrestrial diabase Del3-6 (magnetite).

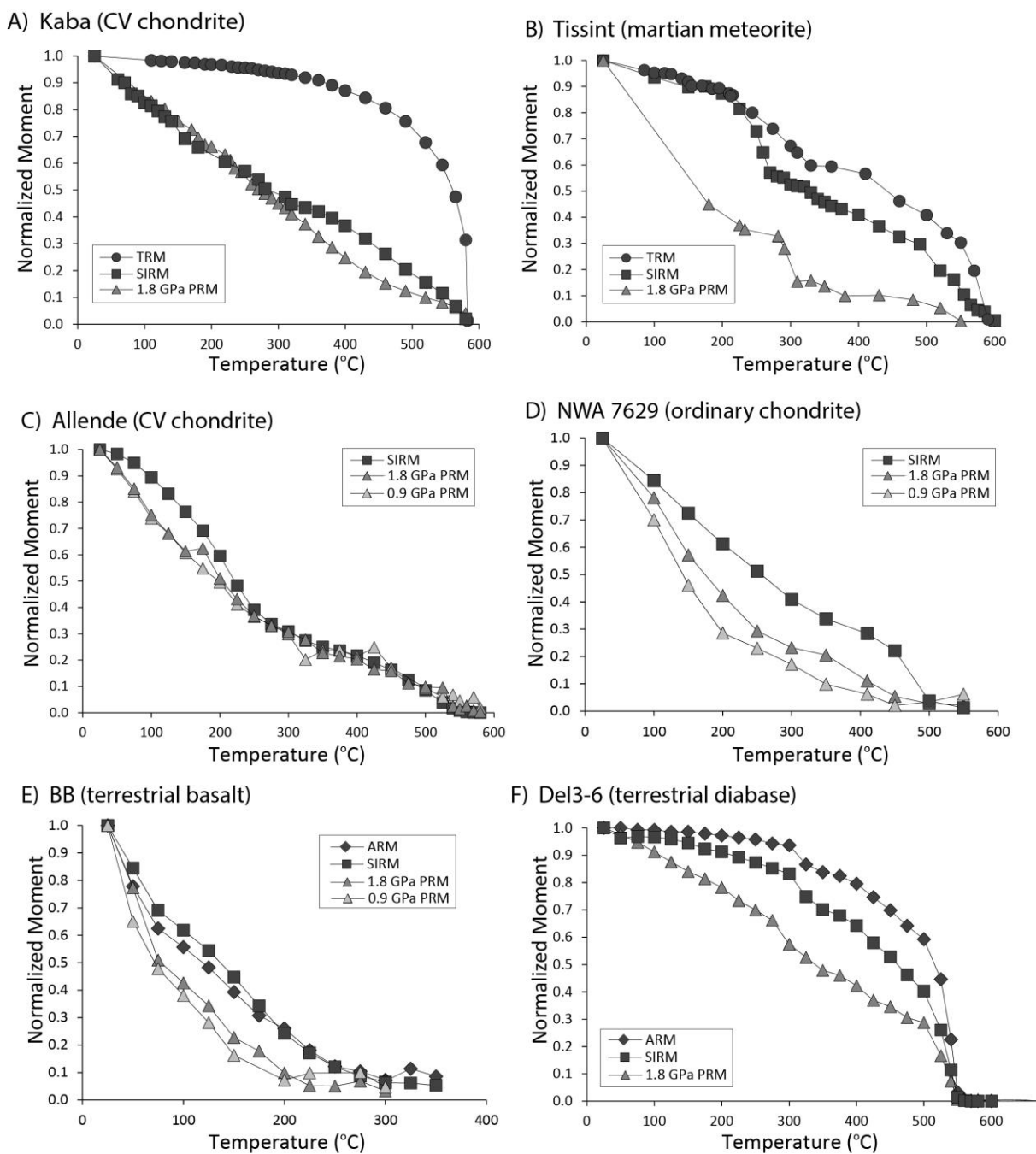


Fig. 5. Thermal demagnetization of PRM (triangles), TRM (circles), ARM (diamonds), and/or SIRM (squares). The ordinate shows the normalized magnetic moment and the abscissa shows the corresponding temperature step. Shown samples (magnetic mineralogies) include: (A) CV3 chondrite Kaba (magnetite), (B) Martian meteorite Tissint (magnetite and pyrrhotite), (C) CV3 chondrite Allende (magnetite and pyrrhotite), (D) NWA 7629 (FeNi). (E) basalt BB (titanomagnetite), and (F) diabase Del3-6 (titanomagnetite).

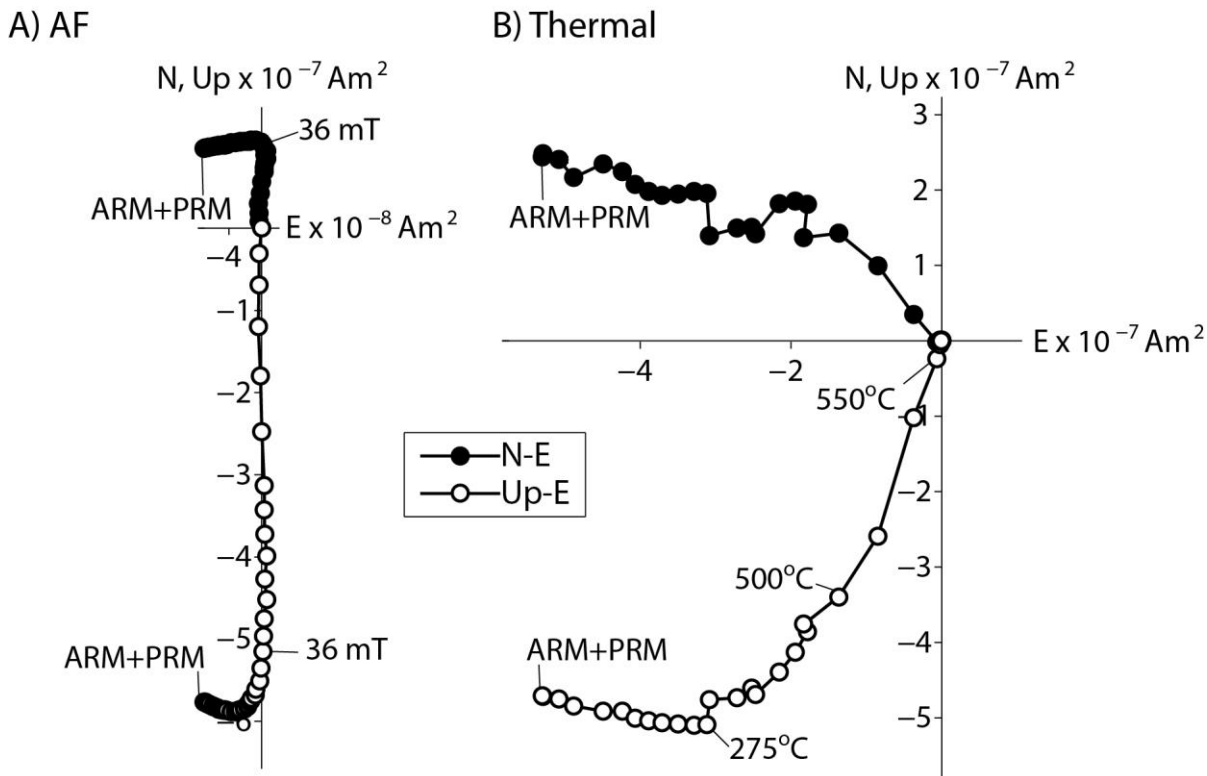


Fig. 6. Demagnetization of ARM with an overlying, orthogonally applied, PRM for two subsamples of terrestrial diabase sample DeI3-6. **(A)** AF demagnetization. **(B)** thermal demagnetization. Open and closed circles represent projections of the NRM vector onto the vertical (Z-E) and horizontal planes (N-E), respectively. Selected AF amplitude and temperature steps are labeled. PRMs were imparted using a 500 μT dc field at 1.8 GPa peak pressure while ARMs were imparted using a 100 μT dc field in a peak ac field of 300 mT. Disparities in remanence intensities between the two subsamples may be attributed to differences in sample mass and vector subtraction of a spurious gyroremanent magnetization component at the end of the AF demagnetization experiment in part (A).

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Table 1. Magnetic and petrophysical properties of samples.

| Sample Name | Magnetic carrier | M_{rs} (Am ² /kg) | M_s (Am ² /kg) | M_{rs}/M_s | B_{cr} (mT) | B_c (mT) | B_{cr}/B_c | $\alpha_{2\text{ mT}}$ (1.8 GPa) | $\alpha_{0.9\text{ mT}}$ (0.9 GPa) | Reference |
|-------------------------------|-----------------------|-----------------------------------|--------------------------------|----------------------|------------------|---------------|--------------|--|--|--------------------------|
| NWA 7629 (H5 chondrite, W1) | Fe-Ni alloy | 5.3×10^{-2} | 4.52 | 1.2×10^{-1} | 31 | 2.5 | 12.4 | - | - | this study |
| NWA 6490 (L5 chondrite, W1) | Fe-Ni alloy | 3.3×10^{-1} | 31.3 | 9.2×10^{-3} | 16 | 1.3 | 12.5 | 2.2×10^{-2} | - | this study |
| 15556 (lunar mare basalt) | Fe-Ni alloy | 6.1×10^{-4} | 0.10 | 5.2×10^{-3} | 32 | 1.3 | 24.8 | 3.6×10^{-2} | 1.2×10^{-2} | Tikoo et al. (2012) |
| 12022 (lunar mare basalt) | Fe-Ni alloy | 8.8×10^{-4} | 0.08 | 1.2×10^{-2} | 49 | 2.0 | 22.2 | - | - | Tikoo et al. (2014) |
| 10020 (lunar mare basalt) | Fe-Ni alloy | 1.5×10^{-3} | 0.01 | 1.2×10^{-2} | 38 | 4.7 | 8.2 | 2.6×10^{-2} | - | Shea et al. (2012) |
| 10017 (lunar mare basalt) | Fe-Ni alloy | 9.4×10^{-4} | 0.17 | 5.5×10^{-3} | 87 | 3.0 | 29.0 | 2.1×10^{-2} | 9.5×10^{-3} | Suavet et al. (2013) |
| 10049 (lunar mare basalt) | Fe-Ni alloy | 1.2×10^{-3} | - | - | 32 | - | - | 1.3×10^{-2} | - | Suavet et al. (2013) |
| ALHA81001 (eucrite) | Fe-Ni alloy | 3.9×10^{-4} | 0.01 | 8.1×10^{-2} | 110 | 41.0 | 2.7 | 5.1×10^{-5} | - | Fu et al. (2012) |
| Allende (CV3 chondrite) | Pyrrhotite, Magnetite | 5.9×10^{-2} | 0.66 | 9.0×10^{-2} | 67 | 13.8 | 4.8 | 1.2×10^{-2} | 6.4×10^{-3} | Wasilewski (1981) |
| Kaba (CV3 chondrite) | Magnetite | 1.8×10^0 | 10.5 | 1.7×10^{-1} | 35 | 14.5 | 2.4 | 2.8×10^{-2} | - | this study |
| BB (terrestrial basalt) | Titanomagnetite | 9.8×10^{-2} | 0.62 | 1.6×10^{-1} | 18 | 5.2 | 3.5 | 1.4×10^{-1} | 1.1×10^{-1} | Gattacceca et al. (2007) |
| Del3-6 (terrestrial diabase) | Magnetite | 8.0×10^{-1} | 2.13 | 3.8×10^{-1} | 63 | 44.4 | 1.4 | 1.2×10^{-2} | - | this study |
| EE (terrestrial microdiorite) | Magnetite | 2.4×10^{-2} | 0.14 | 1.9×10^{-1} | 19 | 1.9 | 10.3 | 2.2×10^{-1} | 1.8×10^{-1} | Gattacceca et al. (2007) |
| LAP 03639 (R chondrite) | Pyrrhotite | 1.6×10^{-2} | 0.06 | 2.8×10^{-1} | 116 | 38.9 | 3.0 | 1.5×10^{-2} | 1.1×10^{-2} | this study |
| PCA 91002 (R chondrite) | Pyrrhotite, Magnetite | 2.5×10^{-2} | 0.10 | 2.5×10^{-1} | 78 | 32.6 | 2.4 | 3.6×10^{-2} | 2.5×10^{-2} | this study |
| Cold Bokkeveld (CM chondrite) | Pyrrhotite, Magnetite | 6.1×10^{-2} | 1.53 | 4.0×10^{-2} | 52 | 10.0 | 5.4 | 1.8×10^{-3} | 1.2×10^{-3} | Cournede et al. (2015) |
| Mighei (CM chondrite) | Pyrrhotite, Magnetite | 3.6×10^{-2} | 0.71 | 5.0×10^{-2} | 58 | 19.0 | 3.0 | 5.0×10^{-4} | - | Cournede et al. (2015) |
| Murchison (CM chondrite) | Pyrrhotite, Magnetite | 8.9×10^{-2} | 0.52 | 1.7×10^{-1} | 61 | 23.0 | 2.6 | 1.6×10^{-3} | - | Cournede et al. (2015) |
| Murray (CM chondrite) | Pyrrhotite, Magnetite | 1.2×10^{-1} | 5.69 | 2.1×10^{-2} | 76 | 6.0 | 13.4 | 2.2×10^{-3} | - | Cournede et al. (2015) |
| Nogoya (CM chondrite) | Pyrrhotite, Magnetite | 8.7×10^{-2} | 1.05 | 8.3×10^{-2} | 83 | 18.0 | 4.5 | 1.4×10^{-3} | - | Cournede et al. (2015) |
| Paris (CM chondrite) | Pyrrhotite, Magnetite | 1.2×10^{-1} | 4.60 | 2.7×10^{-2} | 66 | 7.0 | 9.6 | 2.6×10^{-3} | - | Cournede et al. (2015) |
| Tissint (shergottite) | Pyrrhotite, Magnetite | 6.2×10^{-2} | 0.17 | 3.6×10^{-1} | 78 | 46.4 | 1.7 | 9.4×10^{-3} | - | Gattacceca et al. (2013) |

Note: The first column contains the sample name and description. The second column contains the dominant magnetic mineralogy. The third and fourth columns contain the saturation remanent magnetization (M_{rs}) and saturation magnetization (M_s), respectively. The fifth column contains M_{rs}/M_s . The sixth and seventh columns contain the remanent coercivity (B_{cr}) and coercivity (B_c) respectively. The eighth column contains B_{cr}/B_c . The ninth and tenth columns contain efficiency of 1.8 GPa and 0.9 GPa PRM relative to TRM, respectively. SIRM normalization was used to obtain TRM estimates for all the PRM efficiency calculations. The eleventh column contains the sources for ferromagnetic mineralogy descriptions and hysteresis parameters.