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Pressure demagnetization of the Martian crust: Ground truth from SNC meteorites

Natalia S. Bezaeva,[1,2] Pierre Rochette,[1] Jérôme Gattacceca,[1] Ravil A. Sadykov,[3,4] and Vladimir I. Trukhin[2]

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[1] We performed hydrostatic pressure demagnetization experiments up to 1.3 GPa on Martian meteorites: nakhlite NWA998 (magnetite-bearing), basaltic shergottites NWA1068 (pyrrhotite-bearing) and Los Angeles (titanomagnetite-bearing) as well as terrestrial rocks: rhyolite (hematite-bearing) and basalt (titanomagnetite-bearing), using a new non-magnetic high-pressure cell. The detailed description of measuring techniques and experimental set-up is presented. We found that under 1.3 GPa the samples lost up to 54% of their initial saturation isothermal remanent magnetization (IRM). Repeated loading resulted in a further decrease of magnetization of the samples. Our experiments show that the resistance of IRM to hydrostatic pressure is not exclusively controlled by the remanent coercivity of the sample, but is strongly dependant on its magnetic mineralogy. There is no simple equivalence between pressure demagnetization and alternating field demagnetization. The extrapolation of these results of pressure demagnetization of IRM of Martian meteorites to the demagnetization of the Martian crust by impacts is discussed. **Citation:** Bezaeva, N. S., P. Rochette, J. Gattacceca, R. A. Sadykov, and V. I. Trukhin (2007), Pressure demagnetization of the Martian crust: Ground truth from SNC meteorites, *Geophys. Res. Lett.*, *34*, L23202, doi:10.1029/2007GL031501.

1. Introduction

[2] Mars Global Surveyor (MGS) mission brought to light the existence of strong remanent magnetizations of the Martian Noachian crust [*Acuña et al.*, 1999]. But there are also Noachian areas with low magnetization related to large impacts basins [*Hood et al.*, 2003]. Understanding the process of impact demagnetization is therefore a key issue in interpreting the natural remanent magnetization (NRM) around Martian impact basins, as well as in other impacted solar system bodies (e.g. Earth, Moon, meteorites). Different authors have studied shock demagnetization effects in the 1–10 GPa range by using experimental simulations on magnetic remanence of model rocks or pure minerals [e.g., *Hornemann et al.*, 1975; *Martelli and Newton*, 1977; *Cisowski and Fuller*, 1978; *Kletetschka et al.*, 2004; *Gattacceca et al.*, 2006; *Gattacceca et al.*, 2007; *Louzada et al.*, 2007]. The main problems of such experiments are dynamic pressure calibration and the deciphering of the effect of deviatoric versus hydrostatic stresses. Indeed, it is known that NRM is more sensitive to deviatoric than hydrostatic stress [*Nagata*, 1966; *Pearce and Karson*, 1981]. Moreover shock may permanently modify the intrinsic magnetic properties (e.g., coercivity, see *Gattacceca et al.* [2007]), thus complicating the interpretation. Static experiments are better suited to tackle these problems but were until recently limited to the low-pressure range (<0.1 GPa, e.g., *Pearce and Karson* [1981]). More recently, *Rochette et al.* [2003] used a pyrrhotite sample compressed up to 3 GPa in a piston cylinder press and remeasured isothermal remanent magnetization (IRM) after pressure release. This scheme has the disadvantage of needing a new sample and a few days of experiment per each pressure value. *Gilder et al.* [2006] performed IRM measurements of pure magnetite under pressure using a diamond anvil non-magnetic cell up to 4.5 GPa. This experiment is restricted to pure strongly magnetic minerals due to the minute sample size. Both experiments, by using a solid confining media, generate some deviatoric stress on the sample.

[3] We present here the first results of static experiments on pressure demagnetization of IRM up to 1.3 GPa in a non-magnetic high-pressure cell that uses a liquid confining media, thus ensuring pure hydrostatic pressure on the sample. Moreover the relatively large sample size allows to measure rocks representative of natural processes. According to common consensus SNC (Shergotty-Nakhla-Chassigny) meteorites originate from Mars [e.g., *McSween and Treiman*, 1998] and represent the only source (except MGS data) of information about the magnetic properties of the Martian crust [*Rochette et al.*, 2005; *Rochette*, 2006]. We therefore used our cell to study SNC samples as well as terrestrial rocks.

2. Samples and Measuring Techniques

2.1. Experimental Setup

[4] The experimental setup was designed to measure magnetization of rock samples under hydrostatic pressure using a high-pressure cell allowing direct measurement in a cryogenic magnetometer. The cell was made of the non-magnetic "Russian alloy" (NiCrAl) and withstood maximum pressure up to 1.3 GPa. The whole device was 29.5 mm in diameter and 64 mm in height. The sample chamber was 6 mm in diameter and 20 mm in height. R. A. Sadykov et al. (unpublished manuscript, 2007) describe in

[1]Centre Européen de Recherche et d'Enseignement des Géosciences de l'Environnement, CNRS, Aix-Marseille Université, Aix en Provence, France.

[2]Faculty of Physics, M. V. Lomonosov Moscow State University, Moscow, Russia.

[3]Institute for Nuclear Research, Russian Academy of Sciences, Moscow, Russia.

[4]Institute for High Pressure Physics, Russian Academy of Sciences, Troitsk, Russia.

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Table 1. Main Magnetic Properties of Investigated Samples[a]

Sample Name	Magnetic Carrier	M_{rs}, Am^2/kg	B_{cr}, mT	B_{cr}/B_c	M_{rs}/M_s	Loss at 1.3 GPa, %	Loss at 1.3 GPa, %
NWA998 nakhlite	magnetite	$171 \cdot 10^{-3}$	41	1.73	0.240	6	6 (1)
NWA1068 shergottite	pyrrhotite	$99.4 \cdot 10^{-3}$	134	1.54	0.470	16	19 (2)
Los Angeles shergottite	titanomagnetite	$360 \cdot 10^{-3}$	55	1.41	0.490	23	71 (12)
Rhyolite	hematite	$2.31 \cdot 10^{-3}$	406	3.56	0.477	27	23 (2)
Basalt	titanomagnetite	$97.8 \cdot 10^{-3}$	18.4	3.54	0.157	54	59 (2)

[a]For NWA998, NWA1068, and Los Angeles, the hysteresis data are from *Rochette et al.* [2005]; for rhyolite and basalt samples the hysteresis data are from *Gattacceca et al.* [2007]. In the last column number between brackets indicates the number of pressure cycles between 0 and 1.3 GPa. M_{rs}: saturation remanent magnetization (= SIRM); B_{cr}: coecivity of remanence; B_c: coercivity; M_s saturation magnetization.

more details this clamp cell design (see auxiliary material).[1] A sample was placed into a teflon container filled with polyethylsiloxane (PES-1) liquid allowing to convert the uniaxial pressure on the pistons into a pure hydrostatic pressure. The pressure cell permitted using samples with maximum height of 15 mm and diameter of 6 mm, fitting into the teflon container. After pressure application on the cell with a press (Graseby Specac 15011) at a given load, pressure was locked inside with two screws. Pressure inside the cell was calibrated using a manganin sensor, whose resistance is a known function of pressure. For this we carried out an experiment on the cell with the sensor and a sample inside, following the same procedure as the other experiments. Due to friction, actual pressure is about 10% less than the pressure estimated from the known external load. The press with the cell inside was placed at the center of three pairs of perpendicular Helmholtz coils connected to stabilized DC supplies. The magnetic field in the area of investigated sample was monitored using a flux-gate magnetometer. Due to the presence of mobile metallic parts in the press it was not possible to reduce the ambient field below 5 μT. Measurements of the magnetic moment of the sample under pressure were performed by inserting the pressure cell with sample in a 2G Enterprises DC SQUID magnetometer placed in a magnetically shielded room. This magnetometer allowed the measurement of magnetic moment up to 10^{-4} Am^2 with a noise level of 10^{-11} Am^2.

[5] Measurements of magnetic remanence of the cell without sample after pressure application were carried out to check its intensity and stability. Cell magnetic moment remained stable after 5–10 min of relaxation in the zero magnetic field inside the magnetometer and gave values increasing with pressure, from 3 to $4.5 \cdot 10^{-8}$ Am^2. Remanence direction changed from nearly along the cylinder axis to nearly perpendicular to it from 0 to 1. 3 GPa. As the exposure to laboratory field did not change the cell remanence after relaxation, the magnetization of the sample could be corrected for the magnetization of the cell. In the present study the magnetic moments of the samples were always at least two orders of magnitude above cell magnetic moment. Before pressure application samples were given a saturation IRM (SIRM) in a 3T magnetic field. After placement of the sample into the high-pressure cell, sample magnetic moment was measured with SQUID magnetometer after waiting 5–10 minutes for cell relaxation. Pressure was increased stepwise (8 steps up to 1.3 GPa) and remanence measured after each step.

[1]Auxiliary materials are available in the HTML. doi:10.1029/2007GL031501.

[6] Pressure was then decreased to zero and remanence measured again. The specimen was then extracted from the pressure cell and its residual remanence (IRM_P) was demagnetized in alternating fields (AF) up to 150 mT. Sample was then saturated again in a 3T magnetic field and was demagnetized by AF. In some cases maximum pressure (1.3 GPa) was applied and removed repeatedly several times to check for the effect of repeated loading on magnetic remanence.

2.2. Description of Samples

[7] Magnetic measurements were carried out on three different Martian meteorites: Los Angeles, NWA998 and NWA1068, representative of various magnetic mineralogies as described by *Rochette et al.* [2001, 2005]. The main carrier of magnetic remanence in the nakhlite NWA998 is Ti-poor pseudo-single-domain (PSD) magnetite while in the Los Angeles and NWA1068 basaltic shergottites it is single-domain (SD) titanomagnetite (with a Curie point at 150°C) and SD pyrrhotite, respectively. In order to compare pressure demagnetization of Martian and terrestrial rocks we also present here two volcanic rocks that were used in the shock experiments of *Gattacceca et al.* [2007]. The first sample is a PSD titanomagnetite bearing basalt from the Bas-Vivarais area, France (described by *Rochette et al.* [1993]), with the same Curie point as Los Angeles. The second sample is a near SD hematite-bearing rhyolite from the Esterel range, France (described by *Vlag et al.* [1997]). Magnetic properties of all samples are presented in Table 1.

3. Experimental Results

[8] IRM pressure demagnetization curves of the three different Martian meteorites and two terrestrial rocks are presented in Figure 1. AF demagnetization curves of uncompressed IRM are presented in Figure 2. Unloading the sample from 1.3 to 0 GPa resulted in an increase of IRM (0–15%). A second (up to 12 times for Los Angeles) loading to 1.3 GPa significantly reduced the IRM (see Table 1), except for the rhyolite which shows a 4% increase.

4. Discussion

[9] Martian meteorite with SD titanomagnetite (Los Angeles) shows the largest pressure demagnetization among the studied Martian meteorites: 23% of SIRM is lost under 1.3 GPa. The terrestrial basalt with the same magnetic carrier (titanomagnetite) but lower coercivity (PSD) was demagnetized by 54%. On the other hand for the other samples the loss at 1.3 GPa does not increase with decreas-

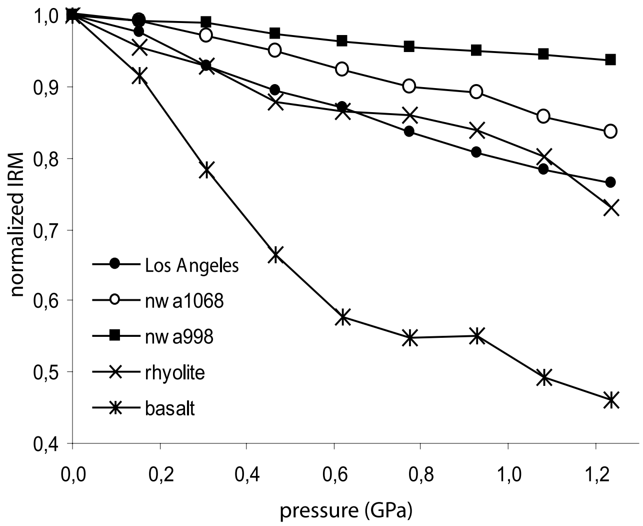

Figure 1. Pressure demagnetization of Martian meteorites and terrestrial rocks (IRM(p = 0) = SIRM and normalized IRM(p = 0) = 1).

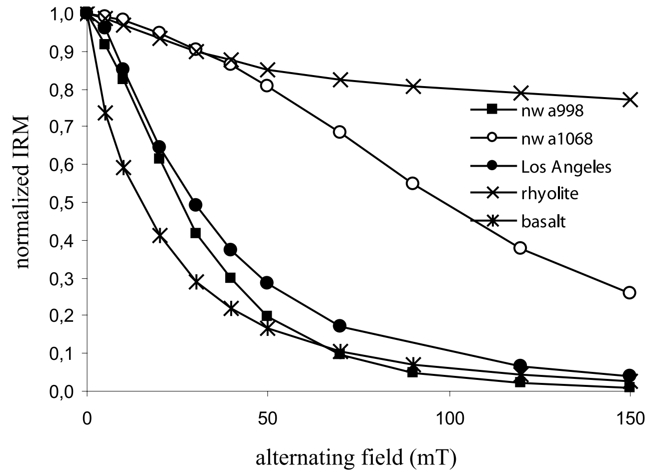

Figure 2. Alternating field demagnetization of Martian meteorites and terrestrial rocks (IRM(AF = 0) = SIRM and normalized IRM(AF = 0) = 1).

ing coercivity (see Table 1). By order of resistance to pressure we find NWA998 (Ti-poor PSD magnetite), NWA1068 (near SD pyrrhotite), Los Angeles (SD titanomagnetite), the rhyolite (near SD hematite) and the basalt (PSD titanomagnetite). This demonstrates that the proportionality relationship between the resistance to pressure demagnetization and coercivity, found by *Kletetschka et al.* [2004] does not hold in general. It holds only for a specific mineralogy, i.e. a specific magnetoelastic constant. In particular Ti-poor magnetite is much more resistant than titanomagnetite as exemplified by NWA998 with respect to Los Angeles, despite the fact that NWA998 has a lower coercivity than Los Angeles.

[10] To represent how the pressure demagnetization affects the different coercivity fractions of the IRM, we computed the demagnetization loss ε,

$$\varepsilon = (\mathrm{IRM_B} - \mathrm{IRM_{BP}})/\mathrm{IRM_B} \times 100,$$

as a function of alternating field B (Figure 3). $\mathrm{IRM_B}$ and $\mathrm{IRM_{BP}}$ correspond to IRM values before and after pressure application respectively. $\mathrm{IRM_B}(B = 0)$ = SIRM and $\mathrm{IRM_{BP}}(B = 0) = \mathrm{IRM_P}$.

[11] In all cases ε decreases with increasing AF, indicating that pressure preferentially demagnetized the lower coercivity fractions. However the 1.3 GPa pressure application does significantly affect the high coercivity part of IRM as exemplified in the case of Los Angeles. As we can see from Table 1 repeated load results in further decrease of magnetic moment. So the repeated loads may be responsible for this "hard" demagnetization in Los Angeles.

[12] Contrary to what was observed in the shock experiments with higher pressures [*Gattacceca et al.*, 2007] we did not observed any change in coercivity of the sample after pressure application. As we found the same values of SIRM before and after exposure to pressure demagnetization, grains are probably not affected by irreversible changes in crystalline structure. The pressure demagnetization is more likely due to domain wall displacements.

[13] The present data on IRM pressure demagnetization of Martian meteorites cannot be directly extrapolated to the demagnetization of the NRM of the Martian crust by impacts, as NRM (likely a thermoremanence) is more resistant to pressure effects than IRM in agreement with its higher coercivity [*Gattacceca et al.*, 2007]. Moreover, our hydrostatic experiments only roughly model the dynamic stress variations recorded during impact shock waves. The shock wave is likely to produce a larger demagnetization than hydrostatic pressure because of the deviatoric stresses present in the shock wave [*Nagata*, 1966;

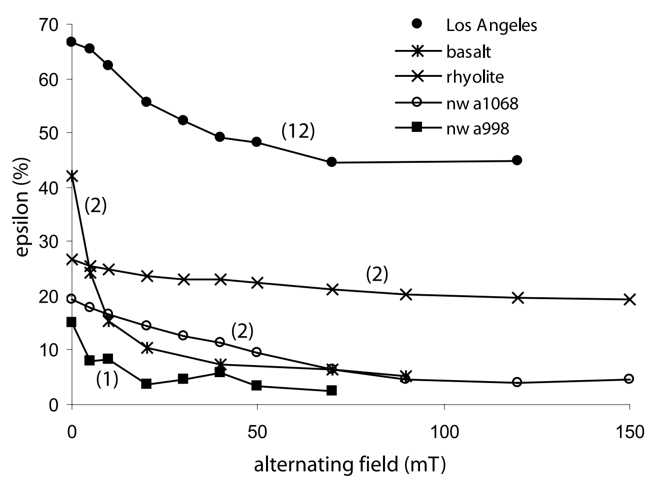

Figure 3. Curves of demagnetization loss ε versus alternating field B. (The chaotic behavior of NWA998 is due to the low demagnetization rate for this coercivity window. We consider that ε is not significant if $\mathrm{IRM_{BP}}/\mathrm{IRM_B} < 0.1$, all such points are deleted from the plot.) The number between brackets associated to each curve indicates the number of cycles between 0 and 1.3 GPa.

Pearce and Karson, 1981] and of the stress oscillations [*Dunlop and Ozdemir*, 1997] that qualitatively corresponds to the multiple loading in our experiments. Repeated loading of the sample in our experiments could also be interpreted as an analogue of multiple impacts.

[14] One may also question the significance of our results in terms of intrinsic effect of hydrostatic pressure on magnetization. Indeed it is likely that, besides the fact that the sample is isotropically loaded, deviatoric stress still exists at the grain scale due to grain anisotropy and mechanical heterogeneities (e.g. grains boundaries). However our results can qualitatively rank the different Martian lithologies in terms of pressure sensitivity. As the shergottites have been recently suggested to represent the Noachian crust [*Bouvier et al.*, 2005], it is worth noting that pyrrhotite and titanomagnetite bearing basaltic shergottites are more pressure sensitive than nakhlites. We can also use our results to estimate in situ effects of hydrostatic load. This is important in the Martian case as the crustal magnetization is thought to be carried on a thick section of the crust (up to 50 km; see *Langlais et al.* [2004]). As Mars shows a pressure gradient of 1.5 GPa per 100 km, we can conclude that the steady pressure demagnetization at 50 km depth is minor (5–15% for IRM according to Figure 1).

5. Conclusions

[15] Through the demagnetization by hydrostatic pressure of various Martian and terrestrial lithologies we have shown that there is no general pressure demagnetization law as a function of remanent coercitivity. Our experiments showed that magnetic mineralogy of samples plays a key role in pressure demagnetization behavior of those samples.

[16] We have shown that there is no simple equivalence between pressure demagnetization and alternating field demagnetization (coercivity). Despite having similar demagnetization effects we can not directly compare pressure and alternating field in terms of process and link pressure values to AF values giving the same degree of demagnetization. AF demagnetizes all grains with coercivity under AF value whereas pressure demagnetizes grains with a large spectrum of coercivity. With pressure, the degree of demagnetization can be different over the coercivity range. Mostly, repeated application of the same pressure demagnetization step results in further decrease of magnetic moments (see Table 1) whereas repeated application of the same AF demagnetization step does not change magnetic moments.

[17] Another aspect of the present results is to bring for the first time ground truth on pressure sensitivity of Martian rocks. Considering the rarity of Martian material it is excluded to perform on them shock experiments that are often destructive and require rather large volume. We can use our hydrostatic results on SNC, compared to the same experiment on terrestrial rocks, to evaluate the representativity of terrestrial analogues of the Martian crust (as studied by e.g., *Gattacceca et al.* [2006, 2007]).

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N. S. Bezaeva, J. Gattacceca, and P. Rochette, Centre Européen de Recherche et d'Enseignement des Géosciences de l'Environnement, CNRS, Aix-Marseille Université, BP-80, F-13545 Aix en Provence, Cedex 4, France. (bezaeva@cerege.fr)

R. A. Sadykov, Institute for Nuclear Research, Russian Academy of Sciences, Prospekt 60-letiya Oktyabrya 7a, Moscow 117312, Russia.

V. I. Trukhin, Faculty of Physics, M. V. Lomonosov Moscow State University, Leninskiye gori, GSP-2, Moscow 119992, Russia.