Evidence for an impact-induced magnetic fabric in Allende, and exogenous alternatives to the core dynamo theory for Allende magnetization

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Abstract—We conducted a paleomagnetic study of the matrix of Allende CV3 chondritic meteorite, isolating the matrix’s primary remanent magnetization, measuring its magnetic fabric and estimating the ancient magnetic field intensity. A strong planar magnetic fabric was identified; the remanent magnetization of the matrix was aligned within this plane, suggesting a mechanism relating the magnetic fabric and remanence. The intensity of the matrix’s remanent magnetization was found to be consistent and low (∼6 μT). The primary magnetic mineral was found to be pyrrhotite. Given the thermal history of Allende, we conclude that the remanent magnetization was formed during or after an impact event. Recent mesoscale impact modeling, where chondrules and matrix are resolved, has shown that low-velocity collisions can generate significant matrix temperatures, as pore-space compaction attenuates shock energy and dramatically increases the amount of heating. Nonporous chondrules are unaffected, and act as heat-sinks, so matrix temperature excursions are brief. We extend this work to model Allende, and show that a 1 km/s planar impact generates bulk porosity, matrix porosity, and fabric in our target that match the observed values. Bimodal mixtures of a highly porous matrix and nominally zero-porosity chondrules make chondrites uniquely capable of recording transient or unstable fields. Targets that have uniform porosity, e.g., terrestrial impact craters, will not record transient or unstable fields. Rather than a core dynamo, it is therefore possible that the origin of the magnetic field in Allende was the impact itself, or a nebula field recorded during transient impact heating.

INTRODUCTION

Carbonaceous chondrite meteorites bear witness to the range of nebular and asteroidal processes that preceded large-scale planetary accretion. These meteorites contain two principal components: abundant submicron and micron-scale matrix materials that form a mineralogically complex aggregate; and millimeter-scale chondrules, the spherical igneous inclusions that give chondrites their name. The Allende meteorite is a member of the CV group of carbonaceous chondrites. Estimates of the age of the solar system are based on analyses of components in Allende and other CV chondrites (Amelin et al. 2009). Allende is arguably the most analyzed rock on Earth, but fundamental aspects of the record of early solar system processes, contained in this meteorite and others, remain poorly understood and a matter of vigorous debate. Their relatively pristine nature has driven the assumption that these meteorites derive from primitive asteroids. However, a body of work—paleomagnetic studies of Allende (Butler 1972; Funaki and Wasilewski 1999; Weiss et al. 2010; Carporzen et al. 2011), numerical modeling (Elkins-Tanton et al. 2011; Sahijpal and Gupta 2011), and compositional associations (Humayun and Weiss 2011)—has prompted the recent suggestion that several
chondrite groups are derived from a large differentiated parent asteroid: an object that had a convecting magma ocean, a liquid metallic core, and an active dynamo field (Weiss et al. 2010; Elkins-Tanton et al. 2011; Humayun and Weiss 2011; Sahijpal and Gupta 2011; Fu et al. 2014).

Three processes are known to control the evolution of chondritic asteroids (1) thermal metamorphism, (2) aqueous alteration, and (3) impact-induced shock metamorphism. All three are significant in interpreting the paleomagnetic record in a meteorite. Shock metamorphism has not been considered a dominant process in the most primitive meteorites, the carbonaceous chondrites: 85% are ranked S1 (“unshocked”; ~4–5 GPa) or S2 (“very weakly shocked”; 5–10 GPa) (Scott et al. 1992). The calibration here (assigning a shock level based on observed shock metamorphic textures, with an estimate of the required shock pressure to generate the textures, and the magnitude of postshock heating) is based on impact recovery experiments on nonporous target rocks or single crystals. Although both Stöffler et al. (1991) and Scott et al. (1992) noted the importance of porosity in determining shock level and impact heating, its significance has rarely been discussed in works applying the Stöffler et al. (1991) criteria to meteorites. This is unfortunate, as porous targets respond very differently to nonporous targets under shock. Porosity compaction attenuates shock energy in an impact and dramatically increases the amount of heating, as energy is expended crushing out the pore space (e.g., Zel’Dovich and Raizer 1967; Kieffer 1971; Ahrens and Cole 1974; Melosh 1989; Sharp and de Carli 2006). The role of porosity is significant when we consider the impact record in carbonaceous chondrite meteorites, as the consensus view is that primordial carbonaceous parent bodies had significant microporosity. In addition, it is particularly important when we consider the paleomagnetic record in meteorites. The interpretation of the paleomagnetism data that underpin the idea that primitive meteorites may come from differentiated asteroids is based on a number of assumptions. A fundamental one—drawing on Stöffler et al. (1991)—is that shock heating was minimal.

Allende is classified as shock stage S1 (Scott et al. 1992). Shock effects in the Stöffler et al. (1991) criteria are estimated based on metamorphic textures in large (>>50–100 μm) chondrule olivines. Shock effects in submicrometer matrix grains are rarely considered. Yet matrix is the host for the magnetic carrier phase. Watt et al. (2006) and Bland et al. (2011) found that the matrix in Allende has a micron-scale fabric. Bland et al. (2011) used fabric analysis to show that the volume of the primary matrix aggregate had been halved in an impact-induced compaction event (determining that the primary matrix porosity, precompaction, was of order 70–80%). In addition to the fabric analysis of meteorite matrix, studies of experimentally synthesized fine-grained material (Blum 2004; Blum and Schrappler 2004), and modeled accreted aggregates (Ormel et al. 2008) indicate that primordial matrix porosities were in the 70–80% range. A review of chondrite porosity data (Sasso et al. 2009; Macke et al. 2011) by Bland et al. (2014) supports this estimate. Static compression experiments indicate that gravitational compression was not significant in asteroids with radii <100 km (Blum 2004). Impact-induced compaction is more efficient, and generates porosities similar to those seen in chondrites (Beitz et al. 2013). Taken together, and given the textural evidence for impact-induced compaction of initially highly porous matrix aggregates (Watt et al. 2006; Bland et al. 2011), the expectation is that primordial asteroids initially had high porosity, and that the dominant porosity-reduction process was impact-induced compaction. What was the effect of that compaction event on Allende matrix? What pressure and temperature did it experience? These questions have implications for our understanding of the paleomagnetic record in Allende and other (compacted) chondritic meteorites.

Although the dichotomy between porous and nonporous targets was well known in the impact community, until recently there had been no numerical studies of shock in materials with a bimodal distribution of porous and nonporous components, i.e., a material approximating a chondritic meteorite: (nominally) zero-porosity spherical chondrules (0.1–1 mm in size) set in a highly porous matrix aggregate composed of submicrometer monomers. In addition, impact simulations have typically been performed studying large-scale collisions or crater-forming events. Bland et al. (2014) and Davison et al. (2016) performed numerical modeling of impacts at sufficient resolution to inform the interpretation of features at 100s μm to cm-scale, i.e., providing an impact simulation baseline appropriate for thin section petrographic studies, or analysis of small meteorite aliquots, and in simulated materials that more closely approximate chondrites. The ability to visualize shock at this “mesoscale,” and observe the effect of a shock wave on low-porosity chondrules set in a high-porosity uncompacted matrix (70–80% porosity) was revealing. Even at relatively low-impact velocities (1–2 km s⁻¹), impact-induced compaction can have a significant effect, and there is significant heterogeneity in shock effects at scales of ~100 μm (Bland et al. 2014; Davison et al. 2016). Most notably, the matrix behaves very differently than chondrules. The mesoscale simulations revealed that
matrix in an Allende compaction scenario would experience much higher postshock temperature increase ($\Delta T_{\text{final}} = 300-400$ K) than chondrules, which are barely heated ($\Delta T_{\text{final}} < 20$ K). Chondrules act as a heat sink—matrix rapidly equilibrates to a bulk postshock temperature $\sim 200$ K lower than matrix $T$ (peak). These impacts would generate negligible shock metamorphic textures in chondrule olivine, consistent with assignment of an S1 shock level for Allende.

There is evidence from previous studies (Watson 1983; Sugiura et al. 1985; Funaki and Wasilewski 1999, 2000; Gattacceca et al. 2005) that in addition to inducing a crystallographic/rock fabric in Allende matrix, impacts also imparted a magnetic fabric. To determine the magnetic fabric, these previous studies measured the anisotropy of magnetic susceptibility (AMS) of Allende matrix and found an oblate magnetic fabric, which is the expected fabric to result from impact. AMS is a popular approach for determining the magnetic fabric due to the speed of measurement (Jackson 1991); however, susceptibility measures the magnetic response of all the minerals in a sample, i.e., both remanence carriers (ferromagnets sensu lato) and nonremanence carriers (paramagnets and diamagnets), and as such does not necessarily reflect the anisotropy of the minerals carrying the natural remanent magnetization (NRM).

Given that the NRM carrier phase is frequently located in matrix, does the magnetic remanence carrier display a magnetic fabric, and what is the effect of matrix heating on the NRM carriers? And if heated, does the magnetic phase record a thermoremanence remanance, and if so what is the origin of the magnetic field? To answer these questions, we report a new magnetic study of Allende. In addition to a standard paleomagnetic study, we also conduct a fabric study of the magnetic remanence plus an ancient-field intensity study (paleointensity) using modern calibrated and noncalibrated, nonheating methods.

**METHODS**

**Paleomagnetic Analysis**

A $60 \times 3 \times 3$ mm section of the Allende meteorite was chosen for analysis, and split into sixteen $\sim 2-3$ mm cubes (Table 1), retaining their relative orientation with respect to each other. To isolate the primary magnetization of the NRM, which is likely to have been superimposed by secondary magnetizations, we applied the standard nonheating paleomagnetic technique of step-wise alternating field (AF) demagnetization up to a maximum alternating field of 120 mT. As the samples were small, to improve signal-to-noise ratios, we measured their remanent magnetization characteristics on a 2G SQUID magnetization at the University of Oxford, fitted with triaxial, static AF demagnetization coils.

To determine the magnetic fabric of the magnetic remanence carriers, we measured the anisotropy of magnetic remanance (AMR). Unlike AMS measurements, AMR measurements isolate the magnetic fabric of the magnetic remanence carriers, and, additionally, AMR is simpler to interpret than AMS; AMS data lead to nonunique interpretations: the magnetic response of small grains (magnetically single-domain [SD]) and larger multidomain (MD) grains is opposite to each other; in AMS measurements, SD grains display “inverse” magnetic anisotropy; in AMR measurements, all grain sizes give the same response (Jackson 1991). The samples were imparted with anhysteretic remanent magnetizations (ARM) in nine individual orientations within the samples, following the protocol of Jelinek (1978). For this we used a peak-alternating field of 200 mT, with a bias field of 100 $\mu$T. Before the measurement, the samples were tumbling-AF-demagnetized using a maximum 100 mT field, followed by static three-axis AF demagnetization up to 200 mT. To impart the ARMs, we used a Detech D-2000 AF Demagnetizer. To improve the signal-to-noise ratio for the AMR study, the 16 samples were combined in orientated pairs, to make eight samples.

To better constrain the homogeneity of the magnetization in samples, estimates of the recorded paleointensity were made. Traditionally, paleointensity measurements are made by replicating the remanence acquisition mechanisms in the laboratory; this essentially means replicating thermoremanence acquisition by heating samples to high temperatures. Generally, meteoritic materials are susceptible to chemical alteration on heating, so nonheating methods are employed, though for thermally stable samples, these methods are generally less accurate (Yu 2006). With the exception of the Preisach paleointensity protocol (Muxworthy and Heslop 2011), all nonheating methods are relative methods that rely on a calibration factor that is often determined by examining material that is of terrestrial origin, e.g., “REM family” methods (Gattacceca and Rochette 2004; Acton et al. 2007). In this paper, we employ the Preisach paleointensity protocol that relies on a first-order model to predict the response and behavior of small magnetic particles in materials, and compare the results to those from the REM$^\prime$ method (Gattacceca and Rochette 2004). The REM$^\prime$ method is the latest development of the REM method; rather than determine just the ratio of the NRM to a laboratory-induced saturating isothermal remanence (SIRM), the REM$^\prime$ method compares the ratio of the NRM and SIRM AF demagnetization spectra, thereby determining
Table 1. Hysteresis parameters and paleofield intensity estimates for the subsamples.

| Sample | Mass (mg) | $H_c$ (mT) | $H_{cr}$ (mT) | $M_{sat}/M_s$ | Range$^a$ (mT) | Steps$^b$ | REM$^c$ (μT) | Preisach$^c$ (μT) |
|--------|-----------|------------|--------------|---------------|----------------|---------|-------------|-----------------|
| a1a    | 83.3      | 18         | 79           | 0.12          | 30–100         | 8       | 13.7 ± 0.4  | 7.8 ± 0.6       |
| a1b    | 69.0      | 13         | 41           | 0.16          |                |         |             |                 |
| a2a    | 88.7      | 20         | 78           | 0.16          | 35–100         | 7       | 11.5 ± 0.6  | 5.7 ± 1.4       |
| a2b    | 77.7      | 22         | 80           | 0.14          | 20–70          | 7       | 11.1 ± 0.3  | 4.0 ± 0.3       |
| a2c    | 73.5      | 21         | 82           | 0.14          | 35–100         | 7       | 11.0 ± 0.4  |                 |
| a2d    | 130.2     | 21         | 85           | 0.14          | 35–100         | 7       | 11.1 ± 0.3  | 5.7 ± 0.1       |
| a2e    | 129.3     | 18         | 83           | 0.10          | 30–100         | 8       | 12.8 ± 1.0  | 6.6 ± 0.9       |
| a2f    | 112.2     | 20         | 82           | 0.12          | 35–100         | 7       | 13.6 ± 1.0  | 5.7 ± 0.6       |
| a2g    | 113.6     | 19         | 84           | 0.13          |                |         |             |                 |
| a2h    | 208.8     | 18         | 80           | 0.13          | 25–100         | 9       | 13.1 ± 1.0  | 6.4 ± 0.7       |
| a3aa   | 64.8      | 19         | 77           | 0.14          | 50–100         | 5       | 13.7 ± 1.1  |                 |
| a3ab   | 82.5      | 18         | 79           | 0.12          | 25–100         | 9       | 12 ± 2      | 5.7 ± 0.6       |
| a3ba   | 49.5      | 19         | 76           | 0.15          | 30–70          | 6       | 13.5 ± 1.5  |                 |
| a3bb   | 64.0      | 20         | 82           | 0.13          | 35–70          | 5       | 12.7 ± 1.0  |                 |
| a4a    | 75.0      | 19         | 81           | 0.14          | 50–120         | 6       | 13.7 ± 0.4  | 7.3 ± 0.1       |
| a4b    | 60.6      | 21         | 81           | 0.15          | 35–100         | 7       | 8.9 ± 1.0   | 4.1 ± 0.2       |

$^a$Range over which the REMc and Preisach estimates were made.

$^b$Number of AF demagnetization steps used in the REMc and Preisach estimations.

$^c$Preisach estimate made using a cooling time of 1 hour to cool from the Curie temperature to ambient temperature.

$^d$No estimate as no clear HC component identified.

$^e$Measured FORC diagram of poor quality.

$^f$No estimate as no clear paleointensity range identified.

a series of REM estimates, one for each AF demagnetization step. Generally in the middle of the spectra, there is usually a plateau of consistent REM estimates. The REM$^c$ intensity is the average of the REM estimates in the plateau region.

Both the REM$^c$ and Preisach techniques assume that the primary NRM is a thermoremanence (TRM) in origin, and not, for example, a thermo-chemical remanent magnetization. For both techniques, the NRMs’ AF demagnetization data are combined with the AF demagnetization data for a laboratory-induced SIRM. No further measurements are needed for the REM$^c$ protocol (Gattacceca and Rochette 2004). For the Preisach method, it is necessary to also measure a series of hysteresis measurements termed first-order reversal curves (FORC) (Roberts et al. 2000). This was done using a Princeton Measurements (now Lakeshore) high-field vibrating sample magnetometer (VSM) at the University of Southampton. In contrast to the original Preisach protocol (Muxworthy and Heslop 2011), where normalization is undertaken by a single SIRM measurement, in this paper, we use SIRM AF demagnetization spectra to normalize (Di Chiara, personal communication). We also measured the standard hysteresis parameters (1) coercive force $H_c$, (2) the remanent coercive force $H_{cr}$, and (3) the reduced remanent saturation magnetization $M_{sat}/M_s$.

To assess the magnetic mineralogy of the remanence carriers, three samples were imparted with a saturation isothermal remanence (SIRM), and continuously thermally demagnetized using an Orion three-axis low-field VSM at Imperial College London.

### Macroscale and Mesoscale Modeling

Simulations of the impact processing on the macro- and mesoscale were performed using the iSALE shock physics code (Amsden et al. 1980; Collins et al. 2004; Wünnemann et al. 2006). Porosity was modeled using the $\varepsilon\zeta$ porous compaction model (Wünnemann et al. 2006; Collins et al. 2011). The ANEOS equation of state table for forsterite (Benz et al. 1989) was used to describe the bulk material in macroscale simulations and both the chondrules and matrix in mesoscale simulations. Lagrangian tracer particles were used to track the peak pressures and temperatures throughout both the macroscale and mesoscale simulations. The macroscale simulations included self-gravity (using the algorithm described in Barnes and Hut [1986]), so the full crater formation and collapse process could be simulated. Mesoscale simulations followed the methodology described in Davison et al. (2016). Randomly placed nonporous chondrules were surrounded by porous matrix. Matrix abundance was set to 70%, and initial matrix porosity was 0.7 (leaving an initial bulk porosity of ~0.5). The initial temperature was set to 400 K. An impact velocity of 1 km s$^{-1}$ was chosen to...
RESULTS

Identification of Primary Magnetization Directions

In 14 of the 16 samples, a high-coercivity (HC) remanent magnetization component with unblocking fields >20 mT was identified; in all samples, the NRM did not fully demagnetize by 120 mT, but the HC component was tending toward the origin (Fig. 1). Principal component analysis (PCA) was used to fit the components (Kirschvink 1980). Plotting the directions on an equal area projection plot (Fig. 2), the HC components are clearly clustered ($\alpha_{95} = 6.5^\circ$), while the more poorly defined low-coercivity (LC) components are scattered. These results are in agreement with previous work for Allende matrix, which identified a HC unidirectional magnetization (e.g., Banerjee and Hargraves 1972; Butler 1972; Nagata 1979; Wasilewski 1981; Sugiura and Strangway 1985; Sugiura et al. 1985; Carporzen et al. 2011; Fu et al. 2014).

Hysteresis Parameters

The samples displayed near consistent hysteresis parameters (Table 1). In terms of domain state, these parameters are indicative of large pseudosingle-domain/MD material. The coercive force values (Table 1) are too high for MD magnetite, and are more indicative of iron sulfides.

Two FORC diagrams are shown in Fig. 3. As a first-order approximation, the x-axis of a FORC diagram can be interpreted as the coercive force distribution, whereas spreading on y-axis is representative of magnetic interactions within the system, both intergrain magnetostatic interactions and/or internal interactions within multidomain grains (Roberts et al. 2014). Figure 3a is representative of all samples, except sample a1b (Fig. 3b). Fig. 3a is typical of iron sulfides, which have relatively higher coercivity distributions than iron oxides and FeNi particles (Roberts et al. 2014). Sample a1b is distinctly different; it had a large chondrule near its surface that is very likely the cause of the anomalous magnetic behavior.

Thermomagnetic Analysis

The magnetization in two of the samples was mostly demagnetized (>95%) by 590–605 K, suggesting the presence of pyrrhotite, which has a Curie temperature of ~595 K (Dekkers 1989), with a high-temperature tail persisting to >750 K (Fig. 4). Pyrrhotite has been previously reported as the primary magnetic mineral in Allende matrix samples (e.g., Fu et al. 2014; Wasilewski 1981; Weiss et al. 2010). Fu et al. (2014) determined a mean matrix composition of Fe$_{6.1}$Ni$_{2.8}$S$_{8.0}$, which for formation at <670 K corresponds to an equilibrium assemblage of pentlandite, troilite, and hexagonal pyrrhotite (Vaughan and Craig 1978). The high-temperature tail is probably magnetite and awaruite as suggested previously (Funaki and Wasilewski 2000); however, it has recently been shown (Tarduno et al. 2016) that Allende matrix is highly unstable to heating, and acquires remanence even on heating in zero-field to...
<620 K; for this to happen requires the creation of a new magnetic phase that is magnetically coupled to the existing remanence carrier. It is possible that the high-temperature tail observed in this study (Fig. 4) is an artifact created during this study.

The other sample (a1b) had not reached its Curie temperature by 1000 K, indicating a Ni-poor FeNi phase (Leedahl et al. 2016); this sample also had a NRM intensity six times greater than the next strongest sample, and was one of the two samples for which HC and LC components were not identified. As stated above, sample a1b had a large chondrule near the surface. Although not thought to be common, Ni-poor FeNi phases have been previously petrographically identified in Allende chondrules (Emmerton et al. 2011).

Anisotropy of ARM

Only eight samples were used to determine the anisotropy, which is statistically low (Tauxe 2010); however, the results from all eight samples were very consistent, especially with respect to the minimum anisotropy axis (Fig. 5). The samples were found to be highly anisotropic (mean $P' = 2.2$ [Jelinek 1981]), displaying a strong planar/oblate anisotropy (foliation, $T = 0.74$ [Jelinek 1981]), within which there is a preferred direction. The anisotropy reported here appears very high compared to what is reported in the literature for other minerals, but these should be compared to the values for pure pyrrhotite samples, e.g., $P' > 40$ (Louzada et al. 2010).

All previous magnetic fabric studies of Allende matrix measured anisotropy of magnetic susceptibility (AMS) (Sugiura et al. 1985; Funaki and Wasilewski 1999, 2000; Gattacceca et al. 2005); all these studies found an oblate anisotropy. Gattacceca et al. (2005) reported an anisotropy value $P \sim 1.09$, which is much lower than the value reported here; however, AMR is known to produce higher anisotropies than AMS, particularly for pyrrhotite-bearing samples (Clement et al. 2008).
Paleointensity Determinations

Both the REM’ method (Gattacceca and Rochette 2004) and the Preisach method (Muxworthy and Heslop 2011; Di Chiara, personal communication) were employed to determine paleointensity estimates (Table 1). For both paleointensity techniques, orthogonal projection plots (Fig. 1) are used to select the AF range of the component of interest, i.e., the HC component. The REM’ paleointensity estimates are simply made by identifying an AF demagnetization range of the HC component for which the NRM/SIRM ratio is relatively constant, averaging this NRM/SIRM ratio, and multiplying the average by 3000 to yield an estimate in micro-Tesla (Gattacceca and Rochette 2004). The REM’ method produced a narrow range of estimates (Table 1), with a mean of 12.2 \( \pm \) 1.4 \( \mu \)T with a 95% confidence interval (CI 95) of 11–13 \( \mu \)T (Table 1).

The Preisach paleointensity method works by using the room-temperature-measured FORC diagram (Fig. 3) to generate a Preisach distribution (Muxworthy and Heslop 2011; Muxworthy et al. 2011b). Using thermally activated Preisach theory, the measured Preisach distribution is used to predict the TRM/SIRM ratio as a function of applied field intensity. The predicted TRM/SIRM ratios are compared with the measured NRM/SIRM ratios to estimate the paleointensity. In a similar manner to the REM’ procedure, to allow for multicomponent magnetizations, the Preisach method determines paleofield estimates for each demagnetization step of the NRM (Fig. 1), and identifies areas of consistency (Di Chiara, personal communication). The Preisach method allows for different cooling rates to be used in the paleointensity calculation. We considered three rates: 6 min, 1 h, and 24 h to cool from the Curie temperature to ambient, though this range of cooling rates only contributed a difference of \(~0.3\) \( \mu \)T to the estimates. The mean estimate for the 1-h cooling time was 5.9 \( \pm \) 1.2 \( \mu \)T (CI 95 5–7 \( \mu \)T) (Table 1).

DISCUSSION

Paleomagnetic analysis clearly demonstrated that the remanent magnetization within the Allende sample was uniform and the matrix’s magnetic signal was dominated by pyrrhotite (Fe\(_{1-x}\)S [\(x = 0–0.17\)]). In all but two of the samples, high-coercivity component directions were clearly aligned, yielding a well-constrained mean direction with a 95% confidence cone (\(\sigma_95\)) of 6.5\( ^\circ\) (Table 1). Given the formation mechanism of the principal magnetic carrier phase (pyrrhotite)—a component of the micrometer to submicrometer matrix material that is interstitial to the millimeter-sized spherical chondrules—the unidirectional HC magnetic remanence must be postaccretional. This is in agreement with previous studies, which have also found a consistent unidirectional magnetization in the Allende

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Fig. 4. Continuous thermal demagnetization curve for sample a2h induced with a saturating isothermal remanent magnetization (SIRM) in a field of 1 T.

Fig. 5. Lower hemisphere projections of the principal (squares), major (triangles), and minor (circles) eigenvectors with 95% confidence ellipses, determined by measuring the anisotropy of AARM. The confidence limits for the principal and major axes are quite large. For comparison, the mean direction of the NRM HC component with 95% confidence ellipse is also plotted. D and I are the declination and inclination of the HC mean, respectively.
matrix (e.g., Banerjee and Hargraves 1972; Butler 1972; Nagata 1979; Wasilewski 1981; Weiss et al. 2010; Carporzen et al. 2011; Fu et al. 2014). Previous studies have also found on thermal demagnetization of Allende NRM, this HC unidirectional magnetization aligns with the “MT” (mid-temperature) component of the three-component thermal demagnetization spectra (Carporzen et al. 2011; Fu et al. 2014); the high-temperature (HT) component is only seen in certain chondrules.

The paleointensity data also support a coeval magnetization process throughout the sample. The Preissach paleointensity estimates are lower than the REM’ methods; from studies on terrestrial historical lavas, the Preissach method has been demonstrated to be more accurate than the REM family of methods (Muxworthy et al. 2011b). We therefore take a paleofield estimate of 5.9 ± 1.2 T, which has little intersample variation (Table 1) suggesting that they have recorded the same field. Compared to other paleofield estimates for Allende bulk/matrix material, this value is slightly lower than previous nonheating estimates: 12–18 μT (Wasilewski 1981) and ~22 μT (REM’) by Emmerton et al. (2011), but lower than the “AF estimate” of ~50–60 μT of Carporzen et al. (2011), a Thellier (heating) estimate of >100 μT (Banerjee and Hargraves 1972), and a single Preissach estimate of ~128 μT of Emmerton et al. (2011). The two differing Emmerton et al. (2011) estimates are for the same sample; usually REM methods yield higher paleointensity estimates than the Preissach method (Muxworthy et al. 2011b). Emmerton et al.’s (2011) Preissach paleointensity estimate was determined using an earlier version of the method (Muxworthy et al. 2011b); here we use the protocol outlined by Di Chiara (personal communication). Carporzen et al.’s (2011) study was not strictly nonheating as the estimates involved a thermal-calibration step. Generally, paleointensity estimates determined by heating protocols, e.g., Thellier-type approaches, yield higher estimates (Butler 1972; Carporzen et al. 2011). Due to known irreversible alteration of the Allende matrix material above 50 °C (Wasilewski 1981; Tarduno et al. 2016), these heating estimates should be treated with caution; Tarduno et al. (2016) found that the Allende matrix acquires remanence even on heating in zero-field.

Recent electron backscatter diffraction (EBSD) analysis (Watt et al. 2006; Bland et al. 2011) has found a pervasive uniaxial crystallographic fabric in Allende delineated by oriented matrix grains. Grain rotation occurred in response to impact shock, and an initially highly porous random aggregate of submicrometer fayalitic olivine grains was compacted to produce a uniaxial crystallographic matrix fabric (Watt et al. 2006; Bland et al. 2011). The AMR analysis of the magnetic fabric found the sample to be highly anisotropic (mean $P^2 = 2.2$ [Jelinek 1981]), displaying a strong planar anisotropy (foliation, $T = 0.74$ [Jelinek 1981]), within which there is a preferred direction. The mean high-coercivity remanence direction of the NRM lies at 95% confidence ellipse within the easy-plane (Fig. 5). Given the high anisotropy of the sample, it seems likely that the direction of the NRM is controlled/influenced to a degree by the intrinsic crystallographic fabric of the samples’ matrix (Watt et al. 2006; Bland et al. 2011). This in turn supports a common, impact compaction mechanism for both the crystallographic and the magnetic fabric, which would have induced ordering of the matrix pyrrhotite grains’ orientations, along with fayalitic olivine.

Given the planar nature of the fabric, we consider the most likely cause of the magnetic fabric to be an impact event. If impact generated, what information does this imply about the magnetization process? What are the implications for the remanent magnetization? Is the remanent magnetization controlled or affected by the impact? There are three scenarios:

1. the HC component of the NRM was formed at the same time as impact,
2. the HC component was formed preimpact, and was rotated into the plane,
3. the HC component was formed postimpact, and the magnetic remanence direction was strongly controlled by the existing fabric.

Impacts are thought to induce a remanent magnetization through one of two mechanisms (1) sufficient heating to induce thermomagnetic recording (Néel 1955) and (2) piezomagnetism (shock-magnetism) due to the interaction of the elastic and magnetic properties of a mineral (magnetoelastic interaction) (Nagata 1961); both mechanisms require the presence of an external field. However, shock-magnetization is thought to only magnetize the low-coercivity magnetic minerals (Cisowski and Fuller 1978; Louzada et al. 2010), i.e., “soft” magnetic minerals that are unlikely to be stable over billions of years, whereas thermoremanent magnetizations (TRM) have the potential to be stable over many billions of years (Néel 1955).

Within the paleomagnetic community, it is generally considered that for a TRM to be induced, high-shock pressures (>40 GPa) are required to produce sufficient heating (Weiss et al. 2010). This level of shock (>S4 in ordinary chondrites) would generate pervasive shock metamorphism throughout a meteorite. Allende is classified as stage S1 (shock pressures <5 GPa)—macroscopic shock textures are absent (Scott et al. 1992). Peak-shock pressures <4–5 GPa are thought to leave a meteorite unscathed, with no effects resulting
from a postshock temperature increase of ~20 K (Stöffler et al. 1991). In addition, although an impact may amplify an existing field, and a transient field may be produced by an impact (Doell et al. 1970; Hide 1972; Srnka 1977; Hood 1987; Crawford and Schultz 1988, 1993, 1999; Hood and Artemieva 2008), slow cooling from a high postshock temperature would not allow a magnetic phase to thermomagnetically record the transient (minutes) field generated by a large impact (Weiss et al. 2010). The assumption here is that even if an impact is large enough to produce a temperature increase sufficient for a magnetic phase to record a TRM, because a large volume of the target is affected, cooling will be slow. The consensus view therefore is that postshock heating in CCs is negligible—certainly too low to affect the paleomagnetic record in these rocks; and that while impacts may well generate or amplify fields (Weiss et al. 2010), those fields are too brief to be thermomagnetically recorded in the meteorite.

However, as discussed previously, this interpretation is founded on an empirical shock metamorphism calibration (Stöffler et al. 1991), based on shock recovery experiments in nonporous materials. Porous materials respond very differently, and as outlined in the introduction, it is likely that primordial chondritic parent bodies had significant porosity. Pore-space compaction attenuates shock energy and dramatically increases the amount of heating: a temperature increase sufficient for a magnetic phase to record a thermomagnetic remanence is achievable, even in a low-velocity collision. Impact modeling that accounts for the high porosity of primordial matrix indicates that chondrite matrix could be heated to temperatures well above the Curie temperature of pyrrhotite (~320 °C) in even low-velocity collisions (1–2 km s\(^{-1}\)), where bulk shock pressure does not exceed 4 GPa (Bland et al. 2014)—consistent with an S1 shock level. Thus, TRM is possible in typical primitive parent body collisions. Indeed, in evolving from highly porous primordial objects to the meteorites that we see today, it is inevitable. If compacted by impact, all chondrites would have been affected by this process. Therefore, a preimpact origin of the remanence (scenario 2) can be excluded, because even low-velocity impacts are likely to reset any pre-existing magnetic remanence.

The assumption that slow cooling from a high postshock temperature would not allow a magnetic phase to thermally record a transient field can also be deconstructed. It applies if the bulk postshock temperature of the whole object exceeds the Curie point of the magnetic phase. With notable exceptions (e.g., Beitz et al. 2013), impact modeling and experiments assume uniform material properties in the target. In these scenarios, an estimate of bulk postshock temperature has relevance in understanding heating at scales appropriate for interpreting meteorite data. But chondrites are not homogenous targets. More appropriately, they can be approximated as a target that has bimodal material properties—a fine-grained highly porous aggregate (matrix) juxtaposed against nonporous clasts (chondrules). In this scenario, bulk postshock temperature does not provide a useful guide to interpreting meteorite data. Bland et al. (2014) and Davison et al. (2016) showed that impact-induced compaction results in significant matrix heating, but that chondrules are largely unaffected. The result is a localized, transient temperature “spike” in matrix, as chondrules act as a heat sink. Davison et al. (2016) performed a simple finite difference calculation to solve the heat conduction equation and estimate the time scale for temperature equilibration. They found that this time scale is dependent on the final matrix porosity, but for impact scenarios consistent with Allende, the matrix and chondrules likely equilibrated on the order of tens of seconds. This behavior has significance in understanding the paleomagnetic record in meteorites. Specifically, in a scenario where matrix is heated higher than the Curie point of the magnetic phase, but matrix and chondrules together equilibrate to a bulk postshock temperature that is less than the Curie point, a matrix magnetic carrier phase would record a thermomagnetic remanence from any ambient field—stable or unstable, transient or long-lived. We have used mesoscale impact modeling to explore scenarios consistent with observations from Allende (constrained by estimated initial porosity, current bulk and matrix porosity, and the strength of the impact-induced matrix fabric). In our mesoscale modeling, we find that a 1 km s\(^{-1}\) planar impact scenario provides a good match to Allende porosity and fabric data. A number of studies converge on a peak metamorphic temperature for Allende of ~600 K (Rietmeijer and Mackinnon 1985; Weinbruch et al. 1994; Zanda et al. 1995; Bonal et al. 2007). Even assuming that the impact occurred long after peak metamorphism (in this scenario, we consider a starting temperature of 400 K), we still find that a large fraction of matrix is heated above the pyrrhotite Curie point, before being cooled rapidly below it (Fig. 6). This ability of chondrites to essentially record a “snapshot” of any ambient field during impact-induced compaction significantly increases the number of options for the origin of the field. Specifically, it opens up the possibility that we are observing paleomagnetic evidence of transient or unstable fields.

Therefore, if the remanent magnetization was acquired at the time of impact, for the magnetization to still exist, the magnetization must be a thermoremanence.
Basing their hypothesis on Muxworthy et al. (2011a) (a precursor to Bland et al. (2014)), Fu et al. (2014) also postulated this mechanism as the origin of the MT (=HC) component observed in the matrix and the chondrules. As some of the chondrules also exhibit a HT component, the MT remanence would, for these chondrules, be a partial rather than a full TRM. Fu et al. (2014) also considered scenario (3), i.e., the remanent magnetization is postimpact; in this case, the magnetization would most likely be chemical/crystallization remanent magnetization (CRM). This CRM would have been recorded during the formation of new pyrrhotite (aqueous metamorphism) in the presence of a magnetic field of unknown origin, but this is reliant on the magnetic fabric of the newly formed pyrrhotite grains being controlled by the existing crystallographic fabric, which is possible, though the texture is not always fully inherited (Craig and Vokes 1993; Barrie et al. 2010). Kojima and Tomeoka (1996) and Krot et al. (1998) identified crosscutting iron oxides and sulfide phases, suggesting postimpact formation; it is likely that these phases were formed in zero-field as the magnetite signal does not appear to carry a remanence (Watson 1983; Carporzen et al. 2011). These crosscutting iron oxides and sulfide phases also appear isotropic and will likely not contribute significantly to the magnetic fabric. Finally, in addition to the magnetic fabric, it should be noted that the relationship of iron oxide and sulfide veins (Kojima and Tomeoka 1996; Krot et al. 1998) to the larger Allende fabric measured later (via high-resolution EBSD analysis of matrix [Bland et al. 2011; Watt et al. 2006] and CT analysis of larger components [Tait et al. 2016]) has not been established. Allende matrix remains highly porous, ~40% (Bland et al. 2011). Whether compaction was sufficient for veins to visibly rotate within that aggregate is unknown. In short, more work is required to unambiguously determine
What is the origin of the recorded magnetic field? Studies have suggested that core dynamos within the CV and CM parent bodies are an explanation for the paleomagnetic record in meteorites (Weiss et al. 2010; Carporzen et al. 2011; Fu et al. 2014). We have presented evidence that potentially connects the paleomagnetic record in Allende to an impact that compacted Allende matrix and generated a pervasive matrix fabric. This does not specifically rule out a core dynamo, but it does open up a variety of new possibilities to explain the magnetization. External magnetic fields become a possibility. For external fields to be a viable alternative to a core dynamo requires that Allende must have been proximal to those fields. Based on our estimated bulk P(shock) for Allende, we can place some constraint on the position of Allende within the parent body with respect to a wide range of impact scenarios. To do this we employ the iSALE shock physics code (Collins et al. 2004; Wünnemann et al. 2006) to model the macroscale pair-wise collision of planetesimals (e.g., Davison et al. 2010, 2012). Bulk pore-space compaction was modeled using the $\nu$-$\alpha$ porous compaction model (Wünnemann et al. 2006; Collins et al. 2011), with both impactor and target given an initial bulk porosity of 50%. The simulations included self-gravity (using the algorithm described in Barnes and Hut 1986), so the full crater formation and collapse process could be simulated (Fig. 7). Lagrangian tracer particles tracked the peak pressure of material throughout the simulation. Peak pressures in the range $1.25$–$2$ GPa (appropriate for Allende, and highlighted in green in Fig. 7) are routinely encountered relatively close to the crater, in the breccia lens. The meteorite could have been exposed to an external field that impact-induced compaction allowed it to record.

There are two possibilities for an external field: an impact-generated field and a disk field. The magnitude of an impact-generated field can be estimated based on scaling relations derived from experimental data (Crawford and Schultz 1999, 2000). The events that we are concerned with are relatively large impacts into $\sim 100$ km diameter parent bodies, where large (transient) fields (orders of magnitude greater than the $11 \pm 4 \mu$T field that we observe in Allende) appear to be possible (the value of the magnetic field experienced will depend on the position of the material relative to the impact). There will be significant uncertainties in empirical scaling relations, and discharge mechanisms may exist, but this experimental work suggests that Allende could have been proximal to an impact-generated field far in excess of that required to explain the paleomagnetic data. The second possibility is disk fields. Although the paleomagnetic record in CMs is similar to CVs, the interpretation is different: Weiss et al. (2010), Carporzen et al. (2011), and Fu et al. (2014) assumed a core dynamo in the case of the CVs, while Cournede et al. (2015) highlighted the possibility that CM chondrite magnetization might have an external (nebula) origin. We agree, and extend that logic to the CVs. Our knowledge of the field strength and topology of disk fields is limited: Fu et al. (2014) reported that the Semarkona meteorite records a nebular field of $54 \pm 21 \mu$T (1–3 Myr), and Stephens et al. (2014) observed that a T Tauri star has a complex magnetic structure. Magnetohydrodynamic (MHD) simulations predict 1–100 $\mu$T fields in the midplane at asteroidal distances (Gammie 1996; Turner and Sano 2008; Bai and Stone 2013). However, as Cournede et al. (2015) noted, the disk has the potential to inherit a net vertical field from the cloud in which it forms, which may then be modified by MHD turbulence moderated by low
ionization. The latest MHD results indicate that this may generate relatively stable fields rather than the time-dependent ones found earlier: fields of order \( \sim 10 \mu T \) (Crutcher 2012) to 100 \( \mu T \) (Wardle 2007). The CM and CV parent bodies would have been exposed to these fields in the first 4 Myrs after CAI formation; recent Mn-Cr dating of secondary Ca-Fe silicates in CVs obtained ages of 3.2 Myr after CAI formation (MacPherson et al. 2017). These Ca-Fe silicates formed cogenetic with fayalite (and magnetite) during alteration on the CV3 parent body. While there is some evidence to suggest that the solar nebula field had decayed by \( \sim 3.8 \) Ma after CAI (Wang et al. 2017), there is no data suggesting that it was absent at earlier times. CVs could have been exposed to a disk field following transient heating of matrix following impact-induced compaction.

**CONCLUSIONS**

This study has shown that the Allende matrix has a strong planar magnetic fabric: there is an easy magnetic plane, which is likely to have formed during impact. The high-coercivity component of the NRM is aligned with this easy magnetic plane, suggesting that the remanent magnetization direction is strongly influenced by the fabric. This is in agreement with previous studies (Sugiura et al. 1985). The NRM is either a thermoremanent formed during the impact or a chemical/crystallization remanent magnetization formed subsequent to impact (though in the latter case this would require pyrrhotite growth to be controlled by the pre-existing fayalitic olivine fabric). Our modeling indicates that a low-intensity (\( \sim 1 \) km \( s^{-1} \)) impact into a simulated target would generate bulk and matrix porosity that are a match to Allende, as well as an impact-induced crystallographic matrix fabric consistent with observations (Bland et al. 2011). Importantly, this scenario would generate matrix heating sufficient to likely reset any previous remanent magnetization in the matrix, and because matrix heating is brief, it allows the magnetic carrier phase to record a transient or unstable field. We note that chondrites—bimodal mixtures of a highly porous matrix aggregate, and nominally zero-porosity chondrules—constitute an impact target material that is uniquely capable of recording these events. Impacts into homogeneously porous targets or low-porosity targets (e.g., planetary crusts [terrestrial impact craters]) generate less fine-scale heterogeneity in heating—bulk \( T_{\text{final}} \) is a useful proxy to \( T_{\text{final}} \) at fine scale. They cannot record transient fields. The paleointensity estimates could not identify the origin of the magnetic field, i.e., impact generated, planetesimal dynamo, or nebula, etc. But an impact-generated field or nebula field recorded during transient heating of matrix would provide an explanation for the mutually orientated nature of the primary magnetization, without having to invoke a paleofield and magnetization mechanism that is inconsistent with Allende’s well-constrained, low-temperature history and undifferentiated nature.

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**REFERENCES**

Acton G., Yin Q. Z., Verosub K. L., Jovane L., Roth A., Jacobsen B., and Ebel D. S. 2007. Micromagnetic coercivity distributions and interactions in chondrules with implications for paleointensities of the early solar system. *Journal of Geophysical Research* 112:B03S90.

Ahrens T. J. and Cole D. M. 1974. Shock compression and adiabatic release of lunar fines from Apollo 17. Proceedings, 5th Lunar and Planetary Science Conference. pp. 2333–2345.

Amelin Y., Connelly J., Zartman R. E., Chen J. H., Goepel C., and Neymark L. A. 2009. Modern U-Pb chronometry of meteorites: Advancing to higher time resolution reveals new problems. *Geochimica et Cosmochimica Acta* 73:5212–5223.

Amsden A. A., Ruppel H. M., and Hirt C. W. 1980. *SALE: A simplified ALE computer program for fluid flows at all speeds*. Los Alamos, New Mexico: Los Alamos National Laboratories. 101 p.

Bai X.-N. and Stone J. M. 2013. Wind-driven accretion in protoplanetary disks 1. Suppression of the magnetorotational instability and launching of the magnetocentrifugal wind. *The Astrophysical Journal* 769:1–21.

Banerjee S. K. and Hargraves R. B. 1972. Natural remanent magnetizations of carbonaceous chondrites and the magnetic field in the early solar system. *Earth and Planetary Science Letters* 17:110–119.

Barnes J. and Hut P. 1986. A hierarchical O(N-Log-N) force-calculation algorithm. *Nature* 324:446–449.

Barrie C. D., Boyle A. P., Cook N. J., and Prior D. J. 2010. Pyrite deformation textures in the massive sulphide ore deposits of the Norwegian Caledonides. *Tectonophysics* 483:269–286.

Beitz E., Guttler C., Nakamura A. M., Tsuchiya A., and Blum J. 2013. Experiments on the consolidation of chondrites and the formation of dense rims around chondrules. *Icarus* 225:558–569.

Benz W., Cameron A. G. W., and Melosh H. J. 1989. The origin of the Moon and the single-impact hypothesis III. *Icarus* 81:113–131.
Bland P. A., Howard L. E., Prior D. J., Wheeler J., Hough R. M., and Dyl K. A. 2011. Earliest rock fabric formed in the solar system preserved in a chondrule rim. *Nature Geoscience* 4:244–247.

Bland P. A., Collins G. S., Davison T. M., Abreu N. M., Ciesla F. J., Muxworthy A. R., and Moore J. 2014. Pressure-temperature evolution of primordial solar system solids during impact-induced compaction. *Nature Communications* 5:5451–5451.

Blum J. 2004. Grain growth and coagulation. *Astrophysics of Dust*, edited by Witt A. N., Clayton G. C., and Drain B. T. San Francisco, California: Astronomical Society of the Pacific. pp. 369–391.

Blum J. and Schrapler R. 2004. Structure and mechanical properties of high-porosity macroscopic agglomerates formed by random ballistic deposition. *Physical Review Letters* 93:115503-1–115503-4.

Bonali L., Bourot-Denise M., Quirico E., Montagnac G., and Lewin E. 2007. Organic matter and metamorphic history of CO chondrites. *Geochimica et Cosmochimica Acta* 71:1605–1623.

Butler R. F. 1972. Natural remanent magnetization and thermomagnetic properties of Allende meteorite. *Earth and Planetary Science Letters* 17:120–128.

Carpentier L., Weiss B. P., Elkins-Tanton L. T., Shuster D. L., Ebel D., and Gattacceca J. 2011. Magnetic evidence for a partially differentiated carbonaceous chondrite parent body. *Proceedings of the National Academy of Sciences* 108:6386–6389.

Cisowski S. M. and Fuller M. 1978. The effect of shock on the magnetism of terrestrial rocks. *Journal of Geophysical Research: Solid Earth* 83:3441–3458.

Clement B. M., Haggerty S., and Harris J. 2008. Magnetic inclusions in diamonds. *Earth and Planetary Science Letters* 267:333–340.

Collins G. S., Melosh H. J., and Ivanov B. A. 2004. Modeling damage and deformation in impact simulations. *Meteoritics & Planetary Science* 39:217–231.

Emmerton S., Muxworthy A. R., Hezel D. C., and Bland P. A. 2011. Magnetic characteristics of CV chondrules with paleointensity implications. *Journal of Geophysical Research* 116:E12007.

Fu R. R., Lima E. A., and Weiss B. P. 2014. No nebular magnetization in the Allende CV carbonaceous chondrite. *Earth and Planetary Science Letters* 404:54–66.

Funaki M. and Wasilewski P. J. 1999. A relation of magnetization and sulphidation in the parent body of Allende (CV3) carbonaceous chondrite. *Meteoritics & Planetary Science* 34:A39.

Funaki M. and Wasilewski P. J. 2000. Pentlandite and magnetic anisotropy in Allende CV3 chondrite (abstract #1148). 35th Lunar and Planetary Science Conference. CD-ROM.

Gammie C. F. 1996. Linear theory of magnetized, viscous, self-gravitating gas disks. *The Astrophysical Journal* 462:725–731.

Gattacceca J. and Rochette P. 2004. Toward a robust normalized magnetic paleointensity method applied to meteorites. *Earth and Planetary Science Letters* 227:377–393.

Gattacceca J., Rochette P., Denise M., Consolmagno G., and Folco L. 2005. An impact origin for the foliation of chondrites. *Earth and Planetary Science Letters* 234:351–368.

Hide R. 1972. Comments on the Moon's magnetism. *The Moon* 4:39.

Hood L. L. 1997. Magnetic-field and remanent magnetization effects of basin-forming impacts of basin-forming impacts on the moon. *Geophysical Research Letters* 14:844–847.

Hood L. L. and Artemieva N. A. 2008. Antipodal effects of lunar basin-forming impacts: Initial 3D simulations and comparisons with observations. *Icarus* 193:485–502.

Humayun M. and Weiss B. P. 2011. A common parent body for Eagle Station Pallasites and CV Chondrites. *Proceedings, 42nd Lunar and Planetary Science Conference*, p1507

Jackson M. 1991. Anisotropy of magnetic remanence—A brief review of mineralogical sources, physical origins, and geological applications, and comparison with susceptibility anisotropy. *Pure Applied Geophysics* 136:1–28.

Jelinek V. 1978. Statistical processing of anisotropy of magnetic susceptibility measures on groups of specimens. *Studia Geophysica et Geodetica* 22:50–62.

Jelinek V. 1981. Characterization of the magnetic fabric of rocks. *Tectonophysics* 79:T63–T67.
Kieffer S. W. 1971. Shock metamorphism of Coconino Sandstone at Meteor Crater, Arizona. Journal of Geophysical Research 76:5449–500.

Kirschvink J. L. 1980. The least-squares line and plane and the analysis of paleomagnetic data. Geophysical Journal of the Royal Astronomical Society 62:699–718.

Kojima T. and Tomeoka K. 1996. Indicators of aqueous alteration and thermal metamorphism on the CV parent body: Microtextures of a dark inclusion from Allende. Geochemistry et Cosmochimica Acta 60:2651–2666.

Krot A. N., Petaev M. I., Scott E. R. D., Choi B. G., Muxworthy A. R., Heslop D., Paterson G. A., and Michalk Muxworthy A. R. and Heslop D. 2011. A Preisach method for estimating absolute paleomagnetic data. Meteoritics & Planetary Science 46:1842–1862.

MacPherson G. J., Nagashima K., Krot A. N., Doyle P. M., and Ivanova M. A. 2017. $^{53}$Mn–$^{33}$Cl chronology of Ca–Fe silicates in CV3 chondrites. Geochemistry et Cosmochimica Acta 201:260–274.

Melosh H. J. 1989. Impact cratering: A geologic process. New York: Oxford University Press.

Muxworthy A. R. and Heslop D. 2011. A Preisch method for estimating absolute paleofield intensity under the constraint of using only isothermal measurements: 1. Theoretical framework. Journal of Geophysical Research 116:B04102.

Muxworthy A. R., Bland P. A., Collins G. S., Moore J. M., Davison T., and Ciesla F. J. 2011a. Heterogeneous shock in porous chondrites: Implications for Allende magnetization. San Francisco: AGU. p. GP11B-07. http://www.scienceirect.com/science/article/pii/S0012821X11040464

Muxworthy A. R., Heslop D., Paterson G. A., and Michalk D. 2011b. A Preisch method for estimating absolute paleofield intensity under the constraint of using only isothermal measurements: 2. Experimental testing. Journal of Geophysical Research 116:B04103. https://doi.org/10.1029/2010JB007844.

Nagata T. 1961. Rock magnetism. Tokyo: Maruzen. 350 p.

Nagata T. 1979. Meteorite magnetism and the early solar-system magnetic-field. Physics of the Earth and Planetary Interiors 20:324–341.

Néel L. 1955. Some theoretical aspects of rock magnetism. Advances in Physics 4:191–243.

Ormel C. W., Cuzzi J. N., and Tielens A. G. G. M. 2008. Co- accretion of chondrules and dust in the solar nebula. The Astrophysical Journal 679:1588–1610.

Rietmeijer F. J. M. and MacKinnon I. D. R. 1985. Poorly graphitized carbon as a new cosmothermometer for primitive extraterrestrial materials. Nature 315:733–736.

Roberts A. P., Pike C. R., and Verosub K. L. 2000. First-order reversal curve diagrams: A new tool for characterizing the magnetic properties of natural samples. Journal of Geophysical Research 105:28,461–28,475.

Roberts A. P., Heslop D., Zhao X., and Pike C. R. 2014. Understanding line magnetic particle systems through use of first-order reversal curve (FORC) diagrams. Reviews of Geophysics. https://doi.org/10.1002/2014RG000462.

Sahijpal S. and Gupta G. 2011. Did the carbonaceous chondrites evolve in the crustal regions of partially differentiated asteroids? Journal of Geophysical Research 116:E06004.

Sasso M. R., Macke R. J., Boesenberg J. S., Britt D. T., Rivers M. L., Ebel D. S., and Friedrich J. M. 2009. Incompletely compacted equilibrated ordinary chondrites. Meteoritics & Planetary Science 44:1743–1753.

Scott E. R. D., Keil K., and Stöffler D. 1992. Shock metamorphism of carbonaceous chondrites. Geochemistry et Cosmochimica Acta 65:4281–4293.

Sharp T. G. and de Carli P. S. 2006. Shock effects in meteorites. In Meteorites and the early solar system II, edited by Lauretta D. S. and McSween H. Y. Tucson, Arizona: The University of Arizona Press. pp. 653–677.

Srška L. J. 1977. Critical velocity phenomena and LTP. Physics of the Earth and Planetary Interiors 14:321–329.

Stephens I. W., Looney L. W., Kwon W., Fernandez-Lopez M., Hughes A. M., Mundy L. G., Crutcher R. M., Li Z.-Y., and Rao R. 2014. Spatially resolved magnetic field structure in the disk of a T Tauri star. Nature 514:597–599.

Stöffler D., Keil K., and Scott E. R. D. 1991. Shock metamorphism of ordinary chondrites. Geochemistry et Cosmochimica Acta 55:3845–3867.

Sugiura N. and Strangway D. W. 1985. NRM directions around a centimeter-sized dark inclusion in Allende. Journal of Geophysical Research 90(supplement):C729–C738.

Sugiura N., Matsui T., and Strangway D. W. 1985. On the natural remanent magnetisation in Allende Meteorite. Lunar and Planetary Institute 16:831–832.

Tait A. W., Fisher K. R., Srinivasan P., and Simon J. I. 2016. Evidence for impact induced pressure gradients on the Allende CV3 parent body: Consequences for fluid and volatile transport. Earth and Planetary Science Letters 454:213–224.

Tarduno J. A., O’Brien T. M., and Smirnov A. V. 2016. Does Earth core dynamo?. 47th Lunar and Planetary Science Conference, 47.2609

Tauxe L. 2010. Essentials of paleomagnetism. Berkley: University of California Press.

Turner N. J. and Sano T. 2008. Dead zone accretion flows in protostellar disks. The Astrophysical Journal Letters 679: L131–L134.

Vaughan D. J. and Craig J. R. 1978. Mineral chemistry of metal sulfides. Cambridge, UK: Cambridge University Press.

Wang H., Weiss B. P., Bai X.-N., Downey B. G., Wang J., Wang J., Suavet C., Fu R. R., and Zucolotto M. E. 2017. Lifetime of the solar nebula constrained by meteorite paleomagnetism. Science 355:623–627.

Wardle M. 2007. Magnetic fields in protoplanetary disks. Astrophysics and Space Science 311:35–45.

Wasilewski P. J. 1981. New magnetic results from Allende C3 (V). Physics of the Earth and Planetary Interiors 26:134–148.
Watson G. S. 1983. Large sample theory of the Langevin distribution. *Journal of Statistical Planning and Inference* 8:245–256.

Watt L. E., Bland P. A., Prior D. J., and Russell S. S. 2006. Fabric analysis of Allende matrix using EBSD. *Meteoritics & Planetary Science* 41:989–1001.

Weinbruch S., Armstrong J., and Palme H. 1994. Constraints on the thermal history of the Allende parent body as derived from olivine-spinel thermometry and Fe/Mg interdiffusion in olivine. *Geochimica et Cosmochimica Acta* 58:1019–1030.

Weiss B. P., Gattacceca J., Stanley S., Rochette P., and Christensen U. R. 2010. Paleomagnetic records of meteorites and early planetesimal differentiation. *Space Science Reviews* 152:341–390.

Wünemann K., Collins G. S., and Melosh H. J. 2006. A strain-based porosity model for use in hydrocode simulations of impacts and implications for transient crater growth in porous targets. *Icarus* 180:514–527.

Yu Y. J. 2006. How accurately can NRM/SIRM determine the ancient planetary magnetic field intensity? *Earth and Planetary Science Letters* 250:27–37.

Zanda B., Bourot-Denise M., and Hewins R. H. 1995. Condensate sulfide and its metamorphic transformations in primitive chondrites. *Meteoritics* 30:605.

Zel’Dovich Y. B. and Raizer Y. P. 1967. *Physics of shock waves and high-temperature hydrodynamic phenomena*. Mineola, New York: Dover Publications. 944 p.