Supplementary Material for

Evidence for a Dynamo in the Main Group Pallasite Parent Body

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Materials and Methods

Magnetic hysteresis data were collected using the University of Rochester Princeton Measurements Corporation Alternating Gradient Force Magnetometer. Values for the examples shown in Fig. 1 of the main text are as follows: $H_{cr}$, $H_c$ and $M_r/M_s$ are 154.6 Oe, 200.0 Oe and 0.3911 respectively for the Esquel specimen, and 111.1 Oe, 151.9 Oe and 0.3714, respectively, for the Imilac specimen. For all remanence measurements we select mm-sized gem-like olivine subsamples, lacking any surface discoloration that might be residual contamination from the surrounding pallasite metal (we note that our initial tests revealed that samples with visible inclusions from olivine crystal rims altered rapidly when heated). Obtaining suitable samples generally required cleaning crystals in distilled water. A weak acid (HCl) was used on some crystals to remove surface contamination. Remanence measurements were made with a 2G Enterprises 3-component 755R DC SQUID magnetometer and a 2G small (6.3 mm) bore 3-component DC SQUID magnetometer in the University of Rochester’s magnetically shielded room (ambient field $<$ 200 nT). CO$_2$ laser heating and cooling was conducted (in air) in additional magnetic shields to produce a magnetically null environment.

Olivine samples 2-3 millimeters in size were mounted on the end of quartz tubes with Omega cement (both of which are routinely measured to ensure the blank is in the $10^{-13}$ to $10^{-14}$ A m$^2$ range). The sample holder also served as the target for CO$_2$ laser heating (the 7 mm diameter laser beam applied at peak temperature for $\sim$1 minute ensures uniform heating of the crystal; heatings at each Thellier-Coe paleointensity step were for 3 minutes). The natural remanent magnetization of approximately 15% of the clean crystal subsamples measured were in the $10^{-9}$ to $10^{-10}$ A m$^2$ range; these are the focus of our studies as the magnetizations are well within the measuring range of the DC SQUID magnetometers throughout the demagnetization procedures. The success rate for crystals having these intensities (yielded interpretable paleointensity results) was $\sim$50%. This compares well with paleointensity success rates from Thellier-Coe experiments on whole-rock terrestrial basalts, which often average 20% (or less).

Thellier-Coe (24) paleointensity data consist of demagnetization of the NRM (field-off step), followed by the reheating of the sample at the same temperature in a known applied field (field-on step). We use orthogonal vector plots of the field-off steps to determine the optimal temperature range to calculate paleointensities. In this study, we typically use a lowermost Thellier-Coe unblocking temperature for paleointensity calculation that is slightly higher than the lowest unblocking temperature where we believe a primary magnetization is held (i.e. 360 °C). This approach is conservative, and aimed to avoid any influence of magnetizations held at lower unblocking temperatures. For consistency, we use this same temperature range in determining paleointensity from Total TRM data (see below), although we note that some minor alteration might be expected given the cumulative time at elevated temperature.
Heatings were minimized by collecting Thellier-Coe paleointensity data only in the temperature range where orthogonal vector plots show univectorial decay. An applied field of 60 µT was used for all Thellier-Coe measurements. After NRM demagnetization and collection of Thellier-Coe data, a Total TRM was applied. Using a CO$_2$ laser, samples were heated to 700 °C and then cooled in the presence of a field over a 10 minute time span. The Total TRM was subsequently stepwise demagnetized using the CO$_2$ laser. An applied field of 60 µT was used in the collection of all initial Total TRM data. After demagnetization of the first Total TRM, subsample Imilac E3 was given a second Total TRM in the presence of a 30 µT field (and subsequently demagnetized with a CO$_2$ laser) to check for any potential applied field dependence on paleointensity.

To test for consistency in magnetic directions, an oriented section 1-mm thick was prepared. Metal was etched away, leaving several mutually oriented gem-like olivine crystals, which we subsequently separated (maintaining orientation) and thermally demagnetized using the CO$_2$ laser.

SOM Text

Paleointensity selection criteria. Examples of accepted results are shown in Figure 2 of the main text. Two additional examples of accepted results are included here (fig. S1). Results of Thellier-Coe and Total TRM paleointensity experiments are reported in tables S1-2. Values are judged acceptable if Thellier-Coe paleointensity and Total TRM paleofield estimates are consistent within 15% (see table S2). The uncertainty in the individual Thellier-Coe and Total TRM paleointensity estimates must be \( \leq 15\% \).

Here we use demagnetization of a Total TRM to assess alteration because it can readily detect (and in our case exclude) whole-scale transformations with heating seen in some FeNi magnetic carriers in meteorites (10). Although our heating times using the CO$_2$ laser are very rapid compared to those of standard ovens used in paleomagnetism, we note that at the end of our experiments a specimen has still been exposed to elevated temperatures for a cumulative time exceeding 2 hours. We forgo pTRM checks (33) which, if applied, would have resulted in even longer cumulative times at elevated temperature. The Total TRM data also aid in the interpretation of magnetizations observed at high unblocking temperatures. For example, some Esquel olivine specimens acquire additional partial TRMs after the temperature at which the NRM appears to have been completely demagnetized. This is expressed as a flattening of NRM/TRM data (Fig. 2C), which in itself might suggest that a very low (or null) field is recorded at high unblocking temperatures. However, demagnetization of a Total TRM reveals only a minor TRM in this same temperature interval (Fig. 2D) suggesting that increases in partial TRM at high temperatures reflect either minor alteration and/or the influence of minor, and more complex, magnetic phases (see discussion in “Minor high unblocking temperature magnetizations” below).

Several factors contribute to the cause of unsuccessful experiments. The NRM intensity of some samples decreased rapidly on AF demagnetization to levels after which measurement with the SQUID magnetometers through an entire paleointensity run was no longer viable. The main cause of unsuccessful samples that did not display such AF demagnetization characteristics appears
to be thermally-induced alteration. This was manifested by either a scattered NRM demagnetization pattern (fig S2A,B) and/or a Total TRM curve that differed markedly from that of the NRM demagnetization (fig S2B,C).

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**Fig. S1.** Additional examples of successful paleointensity experiments on pallasite olivine. (A) Demagnetization of natural remanent magnetization (NRM) of Esquel olivine (black line). (B) Orthogonal vector plot of (A), red is inclination, blue is declination (orientation relative). (C) Thellier-Coe paleointensity data, NRM removed versus thermoremanent magnetization (TRM) gained using a 60 µT applied field suggests a paleofield of 109.6 µT. (D) Demagnetization of a laboratory Total TRM acquired in a 60 µT field (red curve in (A)) suggests a paleofield of 113.4 µT (calculated by comparing values at three temperature steps highlighted by grey boxes). (E-H) Paleointensity data as discussed above on Imilac olivine indicating paleofields of 57.9 µT (Thellier-Coe technique, 60 µT applied field) and 59.9 µT (Total TRM method, 60 µT applied field).
Fig. S2. Examples of paleointensity results that did not meet selection criteria. Intensity versus temperature plots show natural remanent magnetization (NRM) decay (A,B,C) (black) and Total Thermoremanent Magnetization decay (B,C) (red). Orthogonal vector plots are shown for NRM demagnetization (red is inclination, blue is declination of relative orientation).

Paleointensity results and averages. Two pallasite meteorites were sampled (Esquel and Imilac). Two thin slabs from each pallasite were available for study (denoted by 1, 2, respectively in the tables below). Several consistency tests were performed and the results of these tests were incorporated into hierarchial averages (tables S3-4) as follows. For Total TRM, paleofield results from the same crystal measured at different applied field values were averaged (“applied field average”). Paleofield results from different subsamples from a single olivine crystal were averaged (“crystal average”). Results from different crystals from a given meteorite sample were averaged (“meteorite sample estimate”). “Meteorite averages” were determined by averaging the two meteorite sample estimates available for each meteorite studied.
Table S1. Thellier-Coe paleointensity estimates.

<table>
<thead>
<tr>
<th>Subsample</th>
<th>$F_{ThC}$ (µT)</th>
<th>T (°C) [N]</th>
<th>$R^2$</th>
<th>$f$</th>
<th>$g$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Esquel 1 (green4)</td>
<td>132.4 ±5.7</td>
<td>400-500</td>
<td>0.92</td>
<td>0.155</td>
<td>0.278</td>
</tr>
<tr>
<td>Esquel 2 (19c)</td>
<td>110.7 ±5.2</td>
<td>400-450</td>
<td>0.98</td>
<td>0.129</td>
<td>0.493</td>
</tr>
<tr>
<td>Esquel 2 (3c)</td>
<td>116.0 ±5.4</td>
<td>410-485</td>
<td>0.98</td>
<td>0.185</td>
<td>0.788</td>
</tr>
<tr>
<td>Esquel 2 (4b)</td>
<td>109.6 ±7.0</td>
<td>410-500</td>
<td>0.99</td>
<td>0.184</td>
<td>0.724</td>
</tr>
<tr>
<td>Imilac 1 (F8)</td>
<td>74.4 ±6.7</td>
<td>400-500</td>
<td>0.92</td>
<td>0.058</td>
<td>0.392</td>
</tr>
<tr>
<td>Imilac 1 (E3)</td>
<td>64.9 ±4.5</td>
<td>400-500</td>
<td>0.98</td>
<td>0.059</td>
<td>0.742</td>
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<td>Imilac 1 (E7)</td>
<td>57.9 ±7.8</td>
<td>400-500</td>
<td>0.97</td>
<td>0.031</td>
<td>0.707</td>
</tr>
<tr>
<td>Imilac 2 (G9)*</td>
<td>82.1 ±6.3</td>
<td>425-520</td>
<td>0.98</td>
<td>0.038</td>
<td>0.739</td>
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<tr>
<td>Imilac 2 (G12)</td>
<td>79.3 ±7.2</td>
<td>400-520</td>
<td>0.94</td>
<td>0.081</td>
<td>0.726</td>
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</table>

Abbreviations: $F_{ThC}$, Thellier-Coe field value with 1σ uncertainty; $T$, temperature range of fit; $N$, number of temperature steps used in fit; $f, g$ are fraction of NRM fit and gap factor, respectively, from (33). * Sample omitted from averages because of high Total TRM paleointensity uncertainty (see table S2).

Table S2. Total TRM paleointensity estimates.

<table>
<thead>
<tr>
<th>Subsample</th>
<th>$F_{TTRM}$ (µT)</th>
<th>T (°C) [N]</th>
<th>$\Delta_{F_{TTRM}-F_{ThC}}$ %</th>
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<tr>
<td>Esquel 1 (green4)</td>
<td>134.3 ±6.1</td>
<td>400-500</td>
<td>1</td>
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<tr>
<td>Esquel 2 (19c)</td>
<td>118.8 ±5.7</td>
<td>400-450</td>
<td>7</td>
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<tr>
<td>Esquel 2 (3c)</td>
<td>115.9 ±6.8</td>
<td>410-485</td>
<td>&lt;1</td>
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<td>Esquel 2 (4b)</td>
<td>113.4 ±4.0</td>
<td>410-500</td>
<td>3</td>
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<tr>
<td>Imilac 1 (F8)</td>
<td>72.1 ±1.0</td>
<td>400-500</td>
<td>-3</td>
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<td>Imilac 1 (E3)†</td>
<td>65.9 ±4.4</td>
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<td>Imilac 1 (E3)‡</td>
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<td>Imilac 1 (E7)</td>
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<td>400-500</td>
<td>3</td>
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<td>3</td>
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<tr>
<td>Imilac 2 (G12)</td>
<td>77.7 ±2.2</td>
<td>400-520</td>
<td>-2</td>
</tr>
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</table>

Abbreviations: $F_{TTRM}$, Total TRM field value estimate with 1σ uncertainty; $T$, temperature range of fit; $N$, number of temperature steps used in fit; $\Delta_{F_{TTRM}-F_{ThC}}$, difference between Total TRM and Thellier-Coe paleointensity estimates, expressed as percent of the Thellier-Coe value. † 60 µT applied field; ‡ 30 µT applied field. * Sample omitted from averages because of high Total TRM paleointensity uncertainty.
Table S3. Thellier-Coe hierarchical paleointensity averages.

<table>
<thead>
<tr>
<th>Subsample</th>
<th>$F_{ThC}$ (µT)</th>
<th>Crystal average (µT)</th>
<th>Meteorite sample estimate (µT)</th>
<th>Meteorite average (µT)</th>
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<td>$132.4$</td>
<td>$122.3\pm14.4$ (N=2)</td>
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<td>Esquel 2</td>
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<td>19c</td>
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<td>$112.1\pm3.4$ (N=3)</td>
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</tr>
<tr>
<td>3c</td>
<td>$116.0\pm5.4$</td>
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<td></td>
<td></td>
</tr>
<tr>
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<tr>
<td>Imilac 1</td>
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</tr>
<tr>
<td>F8</td>
<td>$74.4\pm6.7$</td>
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<td>$67.9\pm9.2$ (N=2)</td>
<td>Imilac 73.6±8.1 (N=2)</td>
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<tr>
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<td>$79.3$</td>
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Abbreviations: $F_{ThC}$, Thellier-Coe field value. All averages shown with 1σ uncertainty.

Table S4. Total TRM hierarchical paleointensity averages.

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<th>Subsample</th>
<th>$F_{TRM}$ (µT)</th>
<th>Applied field average (µT)</th>
<th>Crystal average (µT)</th>
<th>Meteorite sample estimate (µT)</th>
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<td>$125.2\pm12.9$ (N=2)</td>
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<tr>
<td>4b</td>
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<td>$67.7\pm6.2$ (N=2)</td>
<td>Imilac 72.7±7.1 (N=2)</td>
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<tr>
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<td></td>
<td>$63.3\pm4.7$ (N=2)</td>
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<td></td>
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<td>G12</td>
<td>$77.7\pm2.2$</td>
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<td>$77.7$</td>
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Abbreviations: $F_{TRM}$, Total TRM field value estimate. All averages shown with 1σ uncertainty.
† 60 µT applied field; ‡ 30 µT applied field.
Minor high unblocking temperature magnetizations. Although the dominant natural remanent magnetization is removed by thermal demagnetization between 360 and 500 °C, consistent with a taenite carrier, we note there is a very small signal (1-5% of the NRM) at demagnetization temperatures >500 °C in some samples. On the basis of microprobe analyses (discussed below) and potential unblocking temperatures, we consider these small signals to be carried by a fine-grained mixture of taenite and kamacite. We further note that some samples show a small NRM and Total TRM remanence increase (and subsequent decrease) at thermal demagnetization temperatures >500 °C (cf Figure 2). This increase generally occurs over a restricted temperature range (∼100 °C), but its exact initiation temperature varies between samples. We interpret this as reflecting exchange interaction between fine-grained taenite and kamacite. Because these are very minor phases compared to the bulk magnetization, this interaction is not apparent in FORC diagrams. We also note that small amounts of tetrataenite could be recorded at these high unblocking temperatures. However, the reproducibility of the intensity increase seen in demagnetization of a Total TRM (see Figure 2e) indicates that tetrataenite cannot be solely responsible for these minor magnetizations because tetrataenite should not have survived heating to 700 °C (i.e. the temperature at which the Total TRM was applied).

Terrestrial weathering. Unblocking temperatures similar (but not identical) to those reported in our study have been reported by Uehara et al. (34) in weathered chondrite meteorites and interpreted to reflect maghemite and substituted magnetite formed during terrestrial weathering, resulting in a terrestrial magnetization overprinting an extraterrestrial signal. This was not the case for chondrites with no or little weathering. Maghemite generally inverts after heating above 250 °C (21), and this results in irreversible magnetic behavior; this was not observed in our thermal demagnetization experiments. Moreover, evidence for maghemite or a substituted magnetite phase was not found during our SEM or microprobe analyses (detailed below), whereas clear evidence for FeNi particles was identified. However, we emphasize that our analyses have been restricted to gem-like olivine particles. Our meteorite samples were selected to have minimal weathering. Although not studied here, we predict that weathered pallasite olivines do contain magnetic minerals formed during terrestrial weathering.

SEM and Microprobe analyses of FeNi particles. Scanning electron microscopy (SEM) analyses were conducted using a Zeiss SUPRA 40VP with EDAX spectrometer at the University of Rochester. SEM analyses reveal FeNi inclusions that are potential remanence recorders. These are similar to those reported in some prior studies (35-36) but differ from the tubular symplectic inclusions studied in the Fukang pallasite (37). We observed some Cr-rich inclusions, but these are not candidates for the major NRM carrier which demagnetizes between 360 and 500 °C. SEM analyses of an olivine inclusion that is a candidate remanence carrier from the Esquel meteorite is shown in fig. S3.
Fig. S3. SEM analyses of an inclusion in olivine of the Esquel pallasite meteorite. EDAX K and L shell shell composition maps are shown for Fe and Ni.

EDAX spectra show an absence of Si, Mg and O, indicating that the inclusion is distinct from the olivine matrix. Sulfur-rich regions (darker grey areas of the inclusion in the SEM image) separate concentrations of FeNi within the inclusion.

Compositions of inclusions were further explored using a JEOL 8900 electron microprobe at Cornell University with an accelerating voltage of 8 KeV to obtain ~0.5 micron resolution. Electron microprobe results reveal FeNi compositions within the inclusion (fig. S4). A pentlandite (Fe, Ni)₉S₈ standard from Manibridge, Ontario (weight percentages S: 33.01, Fe: 30.77, Co: 0.10, Ni: 36.12) was used for these analyses. Total weight percentages less than 100% in the analyses plotted reflect the presence of elements other than Fe and Ni (mostly S). The compositions of Ni-rich particles overlap with those of the ordered FeNi mineral tetrataenite. However, the dominant
changes in NRM intensity do not match the characteristic magnetic decay pattern related to the
~550 °C Curie temperature of tetrataenite (38) (see also fig. S5). In addition, as noted above, the
reproducibility exhibited by the Total TRM data (Figure 2, tables S2, S4) are inconsistent with
tetrataenite.

Fig. S4. Electron microprobe results for Esquel olivine (same inclusion shown in Fig. S4). Brightest areas (B,F,G,H) are high Ni, surrounded by Fe sulfide (A). Slightly less bright areas are low Ni, FeNi metal (C,D). Area (E) likely samples region of several adjacent grains, but nonetheless demonstrates S between region of low and high Ni.
Fig. S5. Predicted tetrataenite NRM demagnetization behavior versus observed demagnetization of NRM and total TRM of an Esquel pallasite olivine subsample (cf. Figure 2). Note the NRM value of the hypothetical tetrataenite component is arbitrary, and the decay qualitative, drawn after the behavior reported by Wasilewski (38).

We interpret the dominant magnetic recorder in inclusions within the pallasite olivine as metastable, disordered taenite. We note that this conclusion differs from the result of a thermomagnetic analysis of the “silicate” (presumable olivine) part of a pallasite (Yamato-74044) reported in an early study by Nagata (11) which is consistent with kamacite. We note, however, that our analyses have been restricted to gem-like olivines lacking large inclusions. It is possible that contamination from the metal surrounding the olivine (especially near the olivine rim) was responsible for the signal reported by Nagata (11).

Given the slow cooling recorded by the massive FeNi surrounding the olivine (28), we might expect taenite to have ordered. However, the high-Ni magnetic carriers in olivine have a very different origin from the tetrataenite rims in the metallic matrix as the latter formed by low temperature equilibration between FeNi phases whereas the former are minor components in FeS inclusions that are predominantly composed of FeS. We envisage that the Fe-Ni-S inclusions were injected as a melt into cracks [perhaps at ∼900 °C, (39)] which were sealed up on cooling. We speculate that minor Fe exchange between the FeNi and FeS phases during slow cooling may inhibit ordering in very small particles such as those that characterize the olivine magnetic inclusions.

**Pressure effects on magnetization.** Hydrostatic pressure can potentially affect magnetizations (21). Turcotte and Schubert [(40), eq. 2-71] give the pressure as a function of radial position r inside a homogeneous body of density ρ:

\[
P = \frac{2}{3} \pi \rho^2 G (R^2 - r^2)
\]

where \( R \) is the radius of the body and \( G \) is the gravitational constant. Taking our estimates of the size of the pallasite parent body, \( R=200 \text{ km} \), \( \rho=3 \text{ g/cc} \), we obtain pressures of 9.6 MPa at \( r = 180 \text{ km} \).
and 22 MPa at \( r = 150 \) km. Although detailed pressure experiments are not available on sub-micron sized FeNi particles at these estimated pressures, we note that our estimates of pressures are much lower than the value at which substantial changes are observed in magnetite and titanomagnetite [e.g. (41-42)] and FeNi (43). Shock can induce much larger pressures. However, beyond the event that caused olivine-metal mixing and break-up of the parent body, pallasites do not appear to have been affected by the massive shocks that characterize other meteorites. For example, pallasite metal does not show the characteristic hatched pattern seen in shocked IIIAB iron meteorites (44), and dislocation densities are low in both olivine and metal [e.g. (9, 45)].

Martelli and Newton (46) studied shock associated with a hypervelocity projectile (15 km/s) cratering a terrestrial basalt and proposed that a shock magnetization comparable to a TRM could be produced. In this case, the shock magnetization was thought to be related to a plasma formed by the impact. Because the pallasites we have studied lack massive shock features, we believe they were not near the surface at the time of the impact that disrupted the parent body. In this case, the more relevant far-field remagnetization process is shock-induced crystallographic transformation. Dickinson and Wasilewski (47) found that this process normally induces a magnetization much less than that recorded by our TRM experiments.

**Absolute ages of metal-olivine mixing.** Lugmair and Shukolyukov (26) report \(^{53}\text{Mn},^{53}\text{Cr}\) systematics on main group pallasites Omolon and Springwater, suggesting ages of 4558.0 ± 1.0 Ma and ~4557 Ma, respectively. These ages are consistent with those reported from Re-Os systematics by Shen et al. (48) (4.60 ± 0.5 Ga), who interpret the time scale for pallasite fractional crystalization as spanning 10-20 Myr. These ages provide an upper bound (i.e. oldest age) on the time of metal-olivine mixing.

In an early K-Ar study, Megrue (49) reported ages of 4.3 Ga for 3 pallasites (Krasnojarsk, Marjalahti, and Springwater). A much younger \(^{40}\text{Ar},^{39}\text{Ar}\) age of 0.86 Ga has been reported for the non-main group pallasite Eagle Station by Niemeyer (50), and interpreted as recording an impact of that age. But that study was conducted on weathered olivine [(50), p. 1007] and the age probably reflects terrestrial processes. Similarly Megrue (49), noted that an apparent young Kr-Ar age of 0.44 Ga for the Admire pallasite could reflect “diffusion of radiogenic Ar40” or “potassium contamination of the sample by terrestrial weathering”.

Bondar and Perelygin (27) reported model fission track ages of 4.37 Ga ± 0.01 Ga for the Marjalahti pallasite, 4.19 ± 0.02 Ga for the Omolon pallasite, 4.18 ± 0.03 Ga for the Bragin pallasite and 4.21 ± 0.02 for the Krasnoyarsk pallasite. Given that the temperature for the complete removal of tracks is 450-500 °C (27), these ages could be close to the time of acquisition of the magnetizations we have recorded. However, such an assignment depends critically on the accuracy of the fission track age model. Nevertheless, these ages provide a lower bound (i.e. youngest age) on the time of metal-olivine mixing.

**Thermal modeling.** Given the uncertainties in even such basic parameters as its size, we adopt
a very simple model to track the thermal evolution of the pallasite parent body. This model resembles those of previous investigations [e.g. (29, 51-53)]; one important complication, however, is that it includes an insulating, near-surface megaregolith (54-55). Although, especially for the larger bodies, mantle convection or advection via melt might be important in the early stages, we simply assume spherically-symmetric conductive transfer of heat:

\[
\frac{\partial T}{\partial t} = \frac{1}{r^2} \frac{\partial}{\partial r} \left( \kappa r^2 \frac{\partial T}{\partial r} \right)
\]  

(2)

where \( T(r) \) is temperature, \( r \) is radial position, and \( \kappa \) is thermal diffusivity. We neglect internal heat production because of the \( \sim 100 \) Myr timescale of greatest interest to us. Since this timescale is long compared to the decay of \(^{26}\)Al or \(^{60}\)Fe, the effect of this early heating is incorporated into our initial conditions. Conversely, over the timescale of interest the decay of long-lived radiogenic elements like \(^{40}\)K will not qualitatively change our conclusions.

The temperature at the surface (\( r = R \)) is kept at a fixed value \( T_s \). The temperature gradient at \( r = 0 \) is zero. We assume an isothermal initial condition (\( T = 1600 \) K) in which the mantle is solid but the core is liquid. We solve equation (2) numerically using a centered finite-difference scheme.

The near-surface megaregolith layer is assumed to have a thermal diffusivity one-tenth that of the solid rock (54). Based on Fig. 12 of Warren (55), we assume megaregolith thicknesses of 5, 8 and 24 km for bodies of radii 100, 200 and 600 km, respectively.

The megaregolith is important not only as an insulator, but also because it is an unlikely source of the pallasites (which would have experienced shock and mixing with other silicate phases, neither of which are observed). We consider large (600 km) objects as unlikely sources of pallasites because the cooling rate data would require them to have originated within the megaregolith layer (see main text). We also note that it is possible to ultimately disrupt the parent body by impact without widespread heating (56).

If core solidification has not yet finished, we only solve equation (2) within the mantle and take the temperature at the core-mantle boundary (\( r = R_c \)) as our boundary condition. The core is assumed isothermal (due to vigorous convection) and its bulk temperature is updated based on the heat extracted into the mantle (\( 4\pi R_c^2 k_m \frac{\partial T}{\partial r} \bigg|_{r=R_c} \Delta t \)) during that timestep \( \Delta t \), where \( k_m \) is the mantle thermal conductivity. We take \( R_c = R/2 \).

We assume that the core solidifies at a single temperature \( T_l \), here taken to be 1200 K. If \( T = T_l \) during a particular timestep, the core temperature then remains pinned at that temperature until the total latent heat of the core (\( \frac{4}{3} \pi R_c^3 \rho_c L_c \)) has been extracted into the mantle via conduction. Here \( L_c \) is the specific latent heat of the core material and \( \rho_c \) its density.

Once the core has completely solidified, we then proceed to solve equation (2) as before, but now the bottom boundary condition is to set the temperature gradient to zero at \( r=0 \). Note that the thermal diffusivities \( \kappa_m \) and \( \kappa_c \) for mantle and core respectively are quite different.

We use a grid spacing \( \Delta r = R/100 \) and adopt a constant timestep governed by the Courant criterion: \( \Delta t = 0.3 \Delta r^2 / \kappa_c \), where we use the core diffusivity because it is more restrictive. 

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verified that our numerical model reproduces the analytical solutions for a uniform sphere cooling from an isothermal initial state (57) when the core radius is zero. Parameter values adopted are given in Table S5.

This model ignores many details, such as the fact that core solidification is likely a multi-phase process which involves a progressive reduction in solidus temperature as solidification proceeds. Nonetheless, the conclusions reached in the main text are unlikely to be significantly affected by such uncertainties. This is because the key results - cooling rate and liquid core lifetime - are both primarily dependent on the size of the parent body, while a low blocking temperature implies relatively shallow depths, irrespective of the details of the calculations.

In the main text we focus on the 2-9 K/Myr cooling data for main group pallasites obtained by Yang et al. (28). We do not consider one higher rate reported in that study (18 K/Myr ± 9 K/Myr) because of its high uncertainty; it also appears to be an outlier in cooling rate, and in cloudy zone particle size versus tetrataenite bandwidth (Fig. 7 of Yang et al., (28)). Ito and Ganguly (58) calculate a cooling rate of 20-40 K/Myr at 1000°C on the basis of a thermochronological model applied to $^{53}$Mn-$^{53}$Cr age data on olivine from the Omolon pallasite (26). For our 200 km radius model object, this cooling rate is obtained over a depth range of 6-20 km, which is compatible with the results shown in Fig. 3 of the main text.

Because we envisage the metallic injections as being relatively narrow, dike-like features, they will cool rapidly to attain the background temperature of the surrounding material. Their subsequent cooling history (including passage through the Curie temperature) will then simply be that of the surrounding material.

We also considered the possible role of impact heating during the injection event, but concluded that it was likely to have negligible effect on the overall cooling, as follows. Early in the history of the asteroid belt, impact velocities were probably comparable to escape velocities. The escape velocity is roughly 200 m/s ($R/100$ km) where $R$ is the radius of the body. So for bodies a few hundred km in radius, impact velocities will be lower than the sound speed of intact rock (a few km/s), i.e. the impacts are subsonic. As a result, the impact heating (and associated shock pressures) will be small. If all the kinetic energy of a 200 m/s projectile were contained within one projectile radius, the temperature increase would only be about 20 K and the actual temperature change would be even less, because the heat is in reality more broadly distributed.

### Table S5. Thermal modeling parameters, based on Table 1 of Sahijpal et al. (52).

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Value</th>
<th>Units</th>
<th>Quantity</th>
<th>Value</th>
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<td>$T_l$</td>
<td>1213</td>
<td>K</td>
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<td>m$^2$ s$^{-1}$</td>
<td>$\kappa_c$</td>
<td>$5 \times 10^{-6}$</td>
<td>m$^2$ s$^{-1}$</td>
</tr>
<tr>
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<td>MJ kg$^{-1}$</td>
<td>$k_m$</td>
<td>3</td>
<td>Wm$^{-1}$K$^{-1}$</td>
</tr>
</tbody>
</table>
Late impact heating and remagnetization. Many HED meteorites have $^{39}$Ar-$^{40}$Ar ages that suggest impact-induced heating \((59-60)\) to temperatures capable of partially (or completely) resetting a magnetization that might have been acquired during initial cooling. Some of these impact heating events occur as late as \(\sim 3.4\) Ga, at a time when dynamo action might have ceased on the HED parent body (Vesta) due to core solidification. If the pallasite meteorites we have studied were reheated in a similar manner, after dynamo cessation on the pallasite parent body, it would affect our conclusions. Here, we first consider how our conclusions would be affected, and then explore whether impact reheating is consistent with available constraints.

If the pallasites were reheated after dynamo action ceased, an ambient magnetic field still must have been present to account for the magnetizations we have recorded. A potential source of this ambient field is the entire parent body mantle and crust, which if magnetized during the dynamo epoch would create a global remanent field. The relevant magnetic carrier of the magnetization would be kamacite, since its high Curie point (approximately 765 °C, \((21)\)) is consistent with the idea that it could retain a magnetization even if heated to temperatures sufficient to reset the magnetizations we have reported from the pallasites.

In this case, the constraints on the pallasite parent body would come from the cooling rates of the pallasite metal and the inferred magnetization in the parent body mantle and crust (i.e. not directly from the magnetization of the pallasite olivine, which in this scenario is reset by impact heating and records only an echo of the dynamo). Given these new constraints, a 600 km radius body can still be excluded because the pallasites would have to reside in the regolith. However, we would have less resolution on the position of the pallasites within the parent body and on the parent body size. Specifically, the pallasites could have formed deeper in the parent body, within 10% of the core-mantle boundary, and parent bodies ranging in size from 100 to 200 km radius could satisfy the constraints.

Paleomagnetic directional and paleointensity data from the pallasite olivine are linear between 360 °C and 500 °C (Fig 2). It is unlikely that the ambient magnetic field during a hypothetical late heating event induced by impact, and the field during initial cooling would be exactly the same. Hence, following Thellier’s laws of pTRM additivity \((21)\), if the hypothetical late heating event reached temperatures greater than 360 °C, but less than 500 °C, a break in slope would be seen in the paleointensity and/or directional data. Because this is not observed, we conclude that any late stage heating must have reached a temperature close to 500 °C to be compatible with the magnetic data.

As noted above, a low velocity impact is insufficient to generate a temperature change great enough to affect the magnetization. High-velocity impacts create transient high pressures; as the material subsequently unloads along an adiabat, heating results with higher peak pressures causing more heating \((61)\). To quantify this effect we used the same approach and parameters as in Barnhart et al. \((62)\), Appendix A2. We calculate the peak pressure and temperature increase $\Delta T$ experienced directly below the impactor. The results are plotted in Fig S6. As expected, there is an almost linear relationship between peak pressure and temperature increase. A temperature increase to 500
°C is required to reset the magnetization. This implies temperature increases ranging from 300 K for an impact shortly after core solidification to greater than 400 K for impacts >1 billion years later. This heating range implies shock pressures ranging from 4.5 GPa to greater than 5.5 GPa (∼45-55 kbar).

Studies of olivine dislocation density in main group pallasites limit shock pressure to a few tens of bars (9). These shock levels are incompatible with the much higher shock (i.e. GPa’s) that would accompany impacts needed to produce heating sufficient to reset the magnetizations we have recorded. These studies were conducted only on the Admire, Brenham and Dora pallasites. But metal macro- and microstructure is also sensitive to shock and shock-induced heating. For example, shock up to 1 GPa can cause Neumann band (mechanical twins) in kamacite, and cloudy taenite intergrowth can be completely lost upon short term heating (years) at 500 °C (63). Metal microstructure has been examined in detail for the Esquel and Imilac pallasites (28) and no evidence for shock has been reported.

Because relatively high shock accompanies impact heating (Fig. S6), and there is no evidence of shock at these levels in the pallasites following the olivine-metal mixing event, we conclude that resetting of the pallasite magnetization by impacts is unlikely. We therefore predict that any future Ar-Ar or fission track dating efforts on the Esquel and Imilac pallasites should yield ages at least as old as the 4.4-4.2 Ga ages (27) obtained from other main group pallasites.

Figure S6. Shock pressure and temperature increase as a function of impact velocity $v_i$, calculated using the approach of Barnhart et al. (62), Appendix A2). We assumed target and projectile densities of 2.6 g/cc. Other parameters are as given in Barnhart et al. (62).
**Model for parent body evolution.** Our model for the origin of pallasites and the nature and evolution of their parent body is based on paleomagnetic constraints, cooling rate data (28), thermal modeling, and previously published geochemical analyses. We envision the impact of a differentiated body with a partially solidified core into a larger differentiated body having a liquid core. Liquid FeNi from the impactor intrudes into target dunite mantle in the form of dikes, which result in the formation of the characteristic pallasite compositions. This source of FeNi metal satisfies Ir constraints (3). The mantle and crust of the impactor are lost and/or incorporated into the shallowest levels of the target. A similar impact scenario for the generalized case of planetesimal accretion is presented by Asphaug et al. [(64) Figures 6-7, panels d-f of that paper]. It is also possible that the impactor had a very thin mantle/core, especially if it had seen prior impacts. There are more complex scenarios that might meet the available constraints, but our central conclusions are that the main group pallasites cooled in a body with a core dynamo and that the metal in the pallasites did not come from that core. Our preferred model (fig. S7) is arguably the simplest explanation of these findings.

**Fig. S7 (next page).** Our model for the origin of pallasites and the nature and evolution of their parent body is based on paleomagnetic constraints and thermal modeling. (A) Liquid FeNi from an impactor is injected into the mantle of the parent body. (B) Enlargement of boxed region in (A) shows injection of metal in dike-like intrusions in dunite mantle, which result in the pallasite texture (C) During the dike intrusion, metal is injected into fractures in the olivine (D); these fractures subsequently heal upon cooling. Although we illustrate one impact, several similar impacts might have occurred, and they together might explain the scatter in Ir values seen in main group pallasites [e.g. (65)]. (E) Heat flux and temperature versus time. At a reference of 30 km depth, the lower range of taenite Curie temperatures (estimated from the observed paleomagnetic unblocking temperatures) will be reached at $\sim$137 million years after initial cooling. Core heat flux is sufficient at this time to drive a dynamo if compositional convection occurs (14). (F) After cooling below the lower Curie temperatures of taenite, collision of the pallasite parent body with a similarly-sized protoplanet occurred, possibly in the terrestrial planet-forming zone [e.g. (32)], destroying the pallasite parent body. (G) Pallasite meteorites were later scattered by interaction with a protoplanet.
Fig S7
References and Notes


23. Materials and methods are available as supplementary materials on *Science* Online.


