

1 **Magnesium isotope evidence that accretional vapour loss shapes planetary** 2 **compositions**

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18 **Introductory paragraph**

19 It has long been recognised that Earth and other differentiated planetary bodies are chemically
20 fractionated compared to primitive, chondritic meteorites and by inference the primordial disk from
21 which they formed. An important question has been whether the notable volatile depletions of
22 planetary bodies are a consequence of accretion¹, or inherited from prior nebular fractionation². The
23 isotopic compositions of the main constituents of planetary bodies can contribute to this debate³⁻⁶.
24 Using a new analytical approach to address key issues of accuracy inherent in conventional methods,
25 we show that all differentiated bodies have isotopically heavier magnesium compositions than
26 chondritic meteorites. We argue that possible magnesium isotope fractionation during condensation of
27 the solar nebula, core formation and silicate differentiation cannot explain these observations.
28 However, isotopic fractionation between liquid and vapour followed by vapour escape during
29 accretionary growth of planetesimals generates appropriate residual compositions. Our modelling
30 implies that the isotopic compositions of Mg, Si and Fe and the relative abundances of the major

31 elements of Earth, and other planetary bodies, are a natural consequence of substantial (~40% by
32 mass) vapour loss from growing planetesimals by this mechanism.

33

34 **Main text**

35 Magnesium is a fundamental building block of the terrestrial planets, constituting ~15% of Earth's
36 mass. Compared to the 'solar' composition of the primordial disk, it is well established that the Earth
37 is depleted in Mg by ~20% relative to more cosmochemically refractory elements (e.g. Al), consistent
38 with increasing terrestrial depletions of elements with higher volatility (e.g. ref [7]). Mass-dependent
39 isotopic fractionations of other major elements, Si and Fe, have been used to argue for compositional
40 modification by accretional vapour loss^{3,6}. Yet contrasting conclusions have been derived from similar
41 observations in studies that point to fractionations occurring for these elements during planetary
42 differentiation^{8,9}. In this respect, the isotopic composition of Mg may be better suited to interrogate
43 chemical fractionation in the Solar System⁵. However, detecting mass-dependent isotopic differences
44 between primitive chondrites and differentiated bodies poses significant analytical challenges because
45 only small (<<1‰/amu) fractionations typically occur during the high temperature processes (>1300
46 K) that shape planetary evolution. Indeed, various comparisons of terrestrial and chondritic Mg
47 isotope ratios have reached opposing conclusions as to whether or not the Earth is chondritic^{5,10-17}.
48 Much of the divergent opinion appears to stem from difficulties in achieving sufficiently accurate
49 measurements¹⁸.

50 In Figure 1a we illustrate that a number of different studies report chondritic $^{25}\text{Mg}/^{24}\text{Mg} \sim 0.03$ -
51 0.05% lower than the Earth^{5,10-12}, but in only one case do the authors actually argue that the Earth is
52 non-chondritic⁵. Reliable resolution of such small isotopic differences requires tight control on
53 analytical artefacts. Of particular concern is that samples and standards behave slightly differently
54 during analysis, despite prior purification. Unlike the traditional sample-standard bracketing approach,
55 the method of double-spiking explicitly corrects such behaviour, but it is not standard for elements
56 with only three stable isotopes, like magnesium. We have, therefore, developed a new approach of
57 "critical mixture double spiking" to overcome this problem¹⁹ (see Methods for details). Based on
58 propagation of conservative estimates of systematic error, this method has a limiting accuracy

59 <0.005‰/amu¹⁹. Repeat measurements of solution standards and geological reference materials
60 indicate that we can achieve reproducibilities of ±0.010‰ (2se) for octuple measurements.

61 We have measured the Mg isotope compositions of a range of terrestrial rocks and primitive and
62 differentiated meteorites using critical mixture double spiking (Table 1). As shown in Figure 1b,c, our
63 new data substantiate that chondrites have ²⁵Mg/²⁴Mg ~0.02‰ lower than the differentiated Earth,
64 Mars, eucrite and angrite parent bodies.

65 Visual (Figure 1b) and statistical (see Methods) inspection shows some data scatter. For terrestrial
66 samples this is mostly due to two ocean island basalts (OIB), whose sources likely contain recycled
67 components, originally fractionated at Earth's surface. However, we focus on analyses of mantle
68 peridotites to estimate the composition of bulk Earth more robustly. Mg isotope variability among
69 chondrites is foremost within the carbonaceous group. This scatter is readily explained by variable
70 parent body aqueous alteration, as their ²⁵Mg/²⁴Mg vary systematically with oxygen isotope
71 compositions²⁰ and petrographic class (Extended Data Figure 1) in the sense anticipated from
72 terrestrial weathering²¹. Removing these anomalous samples, we document a statistically significant
73 difference in Mg isotope compositions between chondrites and the variably-sized, differentiated
74 bodies of Earth, Mars and the eucrite parent body ($p_{ANOVA} 1.5 \times 10^{-6}$). If enstatite chondrites are
75 considered separately from other chondrites ($p_{t-test} 0.017$), their mean ²⁵Mg/²⁴Mg is also significantly
76 lighter than differentiated bodies (~0.013‰; $p_{ANOVA} 7.8 \times 10^{-4}$). In contrast, the differentiated bodies are
77 statistically indistinguishable from each other ($p_{ANOVA} 0.16$).

78 The lack of systematic differences in ²⁵Mg/²⁴Mg between melt-depleted harzburgites, fertile
79 lherzolites and mid-ocean ridge basalts reaffirm, at higher precision, previous observations¹² that Mg
80 isotopes do not fractionate discernibly during (ultra)mafic silicate differentiation (Figure 1, Table 1
81 and Extended Data Figure 2). Silicate differentiation is, therefore, not responsible for the relatively
82 heavy Mg isotope compositions of our samples of differentiated planetary bodies.

83 Although the difference in ²⁵Mg/²⁴Mg between Earth and primitive meteorites is small, bulk
84 perturbation of such an abundant element requires the operation of planetary scale processes. Core
85 formation cannot account for this observation, because unlike Fe and Si, Mg only becomes siderophile
86 at temperatures too high for significant isotopic fractionation²². Vapour fractionation typically

87 produces larger isotopic differences than magmatic processes. Such fractionation may have occurred
88 during initial condensation of solids from the cooling proto-solar nebula, as has been invoked to
89 explain variability between meteoritic and planetary silicon isotope compositions⁴. Yet, such a model
90 predicts the low Mg/Si enstatite chondrites should have the lowest ²⁵Mg/²⁴Mg of the primitive
91 meteorites. This is inconsistent with our observations, in which enstatite chondrites have the
92 isotopically heaviest Mg of the anhydrous chondrites and the Earth, moreover, is distinct from
93 chondrites.

94 Instead, we propose that differentiated planetary bodies obtained their relatively heavy Mg isotope
95 compositions by vapour-melt fractionation following impacts during accretionary growth of
96 planetesimals. Loss of vapour, which was subsequently swept into the Sun or blown beyond the zone
97 of accretion, would leave residual planetary bodies isotopically heavy. Vapour loss is most effective
98 for bodies <~0.1 Earth masses (M_{\oplus}) as escape velocities of larger bodies are mostly too high to be
99 exceeded by silicate vapour molecules (see Supplementary Information). We posit that vapour loss
100 signatures were abundantly imparted to small bodies in the inner Solar System, as evidenced by our
101 analyses of silicate achondrites (Figure 1b). Larger bodies like Earth dominantly accreted from such
102 vapour depleted planetesimals, thereby inheriting their signatures.

103 Although small bodies can readily lose vapour, low mutual impact velocities limit the amount of
104 vapour produced directly during these early stages of accretion. To investigate this quantitatively, we
105 have post-processed the results of previously published, high-resolution N-body simulations²³ (see
106 Methods for details) using a parameterisation of impact-induced vaporisation⁴. In this model,
107 cumulative vapour produced increases from <1 to ~20% of a body's final mass for bodies from 10^{-4} to
108 $10^{-1} M_{\oplus}$ (Figure 2a). As an alternative approach we have calculated vapour mass loss by direct
109 outflow of silicate vapour from the surface of impact-generated magma ponds/oceans (see
110 Supplementary Information for details). Although magma oceans may have been produced by
111 radiogenic heating (e.g. ²⁶Al decay), we conservatively focussed on converting impact velocities of the
112 same N-body simulations to energy available for melting/vapour outflow. We obtain cumulative

113 vapour mass losses of ~4-36% for bodies with 10^{-4} to $10^{-1} M_{\oplus}$ final masses (Figure 2b). Given the
114 higher mass losses from smaller bodies in the second, magmatic scenario we focus on this process.

115 The outflow velocities of vapour above the magma ponds/oceans are sufficiently high that this
116 mechanism is not mass discriminative. This implies that isotopic fractionation only occurs during
117 production of vapour at the surface of the magma pond/ocean. We treat this as an equilibrium process
118 between a molten surface and transient atmosphere, as vapour-liquid chemical equilibrium is attained
119 rapidly²⁴. Furthermore, kinetic isotope fractionation in this scenario can be ruled out empirically from
120 consideration of marked effects on K isotope ratios that are unobserved in nature²⁵ (see Methods).

121 We have modelled Mg, Si and Fe equilibrium isotope fractionation between a liquid and vapour
122 with compositions calculated using thermodynamic data for a nine-component system of major and
123 minor elements of variable volatility (see Methods for details). To obtain a residual planetesimal with
124 a Mg isotope composition ~0.02‰ higher than our average chondritic value, our model predicts
125 ~0.30‰ and ~0.013‰ super-chondritic $^{30}\text{Si}/^{28}\text{Si}$ and $^{56}\text{Fe}/^{54}\text{Fe}$ (Figure 3a). These values are largely
126 compatible with the Earth (Figure 3) and several other bodies (Extended Data Figure 3), despite some
127 currently rather poorly constrained input parameters (see Methods); refinement of these values may
128 resolve these minor mismatches. Starting more specifically with an enstatite chondrite composition, as
129 thought most isotopically appropriate for Earth (e.g. ref [26]), ~47% of the body is vaporised and lost,
130 which removes ~14% of the initial Mg mass together with ~65% Si, ~48% Fe and >99.9% of Na and
131 K (Figure 3c). Similar total mass losses were deduced previously for Earth²⁷ from the volatilities of
132 major constituents based on a comparison to vaporisation experiments. The total mass losses inferred
133 from these chemical constraints (Figure 3) are reassuringly similar to those calculated from our
134 physical model of vapour loss (Figure 2b).

135 The predicted elemental losses modify the composition of enstatite chondrites into one similar to
136 the Earth's (Figure 3c,d), without requiring substantial, preferential collisional erosion of a putative
137 crust formed by small degrees melts of an undifferentiated body²⁸. Our modelled vapour-loss
138 modification resolves the conundrum that Earth's isotopic anomalies suggest a close genetic link to
139 enstatite chondrites despite vastly different elemental compositions²⁶. Close matches to Earth's
140 composition are also obtained, however, when starting with e.g. a CI chondrite composition (Extended

141 Data Figure 4). While potassium depletion relative to chondrites in eucrite (>94%) and angrite (>99%)
142 parent bodies²⁵ is comparable to our vapour depletion model results, it requires re-accretion of ~20%
143 chondritic material to match terrestrial alkali abundances. This material can be derived from the
144 distribution tail of bodies that escape melting and vapour loss.

145 Our vapour loss model quantitatively accounts for non-chondritic abundances and isotopic
146 compositions of major elements as a natural consequence of planetary accretion and provides a viable
147 mechanism for previous suggestions of the importance of evaporative losses in planetary
148 evolution^{3,6,27,29}. This process inevitably results in dramatic loss of elements more volatile than Mg.
149 Planetary compositions thus intrinsically reflect their history of collisional formation.

150

151 Main references

- 152 1 Ringwood, A. E. Chemical Evolution of the Terrestrial Planets. *Geochim. Cosmochim. Acta*
153 **30**, 41-104, doi:10.1016/0016-7037(66)90090-1 (1966).
- 154 2 Anders, E. Chemical Processes in the Early Solar System, as Inferred from Meteorites.
155 *Accounts Chem Res* **1**, 289-298, doi:10.1021/ar50010a001 (1968).
- 156 3 Pringle, E. A., Moynier, F., Savage, P. S., Badro, J. & Barrat, J. A. Silicon isotopes in angrites
157 and volatile loss in planetesimals. *Proc. Natl. Acad. Sci., USA* **111**, 17029-17032,
158 doi:10.1073/pnas.1418889111 (2014).
- 159 4 Dauphas, N., Poitrasson, F., Burkhardt, C., Kobayashi, H. & Kurosawa, K. Planetary and
160 meteoritic Mg/Si and $\delta^{30}\text{Si}$ variations inherited from solar nebula chemistry. *Earth Planet. Sci.*
161 *Lett.* **427**, 236-248, doi:10.1016/j.epsl.2015.07.008 (2015).
- 162 5 Young, E. D., Tonui, E., Manning, C. E., Schauble, E. & Macris, C. A. Spinel-olivine
163 magnesium isotope thermometry in the mantle and implications for the Mg isotopic
164 composition of Earth. *Earth Planet. Sci. Lett.* **288**, 524-533, doi:10.1016/j.epsl.2009.10.014
165 (2009).
- 166 6 Poitrasson, F., Halliday, A. N., Lee, D. C., Levasseur, S. & Teutsch, N. Iron isotope
167 differences between Earth, Moon, Mars and Vesta as possible records of contrasted accretion
168 mechanisms. *Earth Planet. Sci. Lett.* **223**, 253-266, doi:10.1016/j.epsl.2004.04.032 (2004).
- 169 7 Palme, H. & O'Neill, H. S. C. in *The Mantle and Core Vol. 2 Treatise on Geochemistry* (ed R.
170 W. Carlson) (Elsevier-Pergamon, 2003).
- 171 8 Weyer, S. & Ionov, D. A. Partial melting and melt percolation in the mantle: The message
172 from Fe isotopes. *Earth Planet. Sci. Lett.* **259**, 119-133, doi:10.1016/j.epsl.2007.04.033
173 (2007).
- 174 9 Georg, R. B., Halliday, A. N., Schauble, E. A. & Reynolds, B. C. Silicon in the Earth's core.
175 *Nature* **447**, 1102-1106, doi:10.1038/Nature05927 (2007).
- 176 10 Bourdon, B., Tipper, E. T., Fitoussi, C. & Stracke, A. Chondritic Mg isotope composition of
177 the Earth. *Geochim. Cosmochim. Acta* **74**, 5069-5083, doi:10.1016/j.gca.2010.06.008 (2010).
- 178 11 Pogge von Strandmann, P. A. E. *et al.* Variations of Li and Mg isotope ratios in bulk
179 chondrites and mantle xenoliths. *Geochim. Cosmochim. Acta* **75**, 5247-5268,
180 doi:10.1016/j.gca.2011.06.026 (2011).
- 181 12 Teng, F. Z. *et al.* Magnesium isotopic composition of the Earth and chondrites. *Geochim.*
182 *Cosmochim. Acta* **74**, 4150-4166, doi:10.1016/j.gca.2010.04.019 (2010).
- 183 13 Wiechert, U. & Halliday, A. N. Non-chondritic magnesium and the origins of the inner
184 terrestrial planets. *Earth Planet. Sci. Lett.* **256**, 360-371, doi:10.1016/j.epsl.2007.01.007
185 (2007).

- 186 14 Bizzarro, M. *et al.* High-precision Mg-isotope measurements of terrestrial and extraterrestrial
187 material by HR-MC-ICPMS—implications for the relative and absolute Mg-isotope
188 composition of the bulk silicate Earth. *J. Anal. At. Spectrom.* **26**, 565-577,
189 doi:10.1039/C0ja00190b (2011).
- 190 15 Handler, M. R., Baker, J. A., Schiller, M., Bennett, V. C. & Yaxley, G. M. Magnesium stable
191 isotope composition of Earth's upper mantle. *Earth Planet. Sci. Lett.* **282**, 306-313,
192 doi:10.1016/j.epsl.2009.03.031 (2009).
- 193 16 Chakrabarti, R. & Jacobsen, S. B. The isotopic composition of magnesium in the inner Solar
194 System. *Earth Planet. Sci. Lett.* **293**, 349-358, doi:10.1016/j.epsl.2010.03.001 (2010).
- 195 17 Yang, W., Teng, F. Z. & Zhang, H. F. Chondritic magnesium isotopic composition of the
196 terrestrial mantle: A case study of peridotite xenoliths from the North China craton. *Earth
197 Planet. Sci. Lett.* **288**, 475-482, doi:10.1016/j.epsl.2009.10.009 (2009).
- 198 18 Teng, F. Z. *et al.* Interlaboratory comparison of magnesium isotopic compositions of 12 felsic
199 to ultramafic igneous rock standards analyzed by MC-ICPMS. *Geochem. Geophys. Geosy.* **16**,
200 3197-3209, doi:10.1002/2015gc005939 (2015).
- 201 19 Coath, C. D., Elliott, T. & Hin, R. C. Double-spike inversion for three-isotope systems. *Chem.
202 Geol.* **451**, 78-89, doi:10.1016/j.chemgeo.2016.12.025 (2017).
- 203 20 Clayton, R. N. & Mayeda, T. K. The oxygen isotope record in Murchison and other
204 carbonaceous chondrites. *Earth Planet. Sci. Lett.* **67**, 151-161, doi:10.1016/0012-
205 821x(84)90110-9 (1984).
- 206 21 Tipper, E. T. *et al.* The magnesium isotope budget of the modern ocean: Constraints from
207 riverine magnesium isotope ratios. *Earth Planet. Sci. Lett.* **250**, 241-253,
208 doi:10.1016/j.epsl.2006.07.037 (2006).
- 209 22 O'Rourke, J. G. & Stevenson, D. J. Powering Earth's dynamo with magnesium precipitation
210 from the core. *Nature* **529**, 387-389, doi:10.1038/nature16495 (2016).
- 211 23 Carter, P. J., Leinhardt, Z. M., Elliott, T., Walter, M. J. & Stewart, S. T. Compositional
212 Evolution during Rocky Protoplanet Accretion. *Astrophys. J.* **813**, doi:10.1088/0004-
213 637x/813/1/72 (2015).
- 214 24 Fegley, B. & Schaefer, L. in *The atmosphere - History Vol. 6 Treatise on Geochemistry* (ed J.
215 Farquhar) Ch. 6.3, 406 (Elsevier, 2013).
- 216 25 Humayun, M. & Clayton, R. N. Potassium isotope cosmochemistry: Genetic implications of
217 volatile element depletion. *Geochim. Cosmochim. Acta* **59**, 2131-2148, doi:10.1016/0016-
218 7037(95)00132-8 (1995).
- 219 26 Javoy, M. *et al.* The chemical composition of the Earth: Enstatite chondrite models. *Earth
220 Planet. Sci. Lett.* **293**, 259-268, doi:10.1016/j.epsl.2010.02.033 (2010).
- 221 27 Young, E. D. Assessing the implications of K isotope cosmochemistry for evaporation in the
222 preplanetary solar nebula. *Earth Planet. Sci. Lett.* **183**, 321-333, doi:10.1016/S0012-
223 821x(00)00276-4 (2000).
- 224 28 Boujibar, A., Andrault, D., Bolfan-Casanova, N., Bouhifd, M. A. & Monteux, J.
225 Cosmochemical fractionation by collisional erosion during the Earth's accretion. *Nature
226 Communications* **6**, doi:10.1038/ncomms9295 (2015).
- 227 29 Pringle, E. A. & Moynier, F. Rubidium isotopic composition of the Earth, meteorites, and the
228 Moon: Evidence for the origin of volatile loss during planetary accretion. *Earth Planet. Sci.
229 Lett.* **473**, 62-70, doi:10.1016/j.epsl.2017.05.033 (2017).
- 230 30 Walsh, K. J., Morbidelli, A., Raymond, S. N., O'Brien, D. P. & Mandell, A. M. Populating the
231 asteroid belt from two parent source regions due to the migration of giant planets—"The Grand
232 Tack". *Meteorit. Planet. Sci.* **47**, 1941-1947, doi:10.1111/j.1945-5100.2012.01418.x (2012).
- 233 31 Wasson, J. T. & Kallemeyn, G. W. Compositions of Chondrites. *Philos T Roy Soc A* **325**, 535-
234 544 (1988).
- 235

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244

245 **Author contributions**

246 All data presented were measured by RH. RH and CDC performed vapour-liquid modelling. PC
247 was responsible for calculations relating to N-body simulations, FN modelled the direct outflow
248 vapour loss mechanism. RH and TE wrote the manuscript. CDC, YJL, PPvS and MW were involved
249 in measurements in the initial stages of this study. All authors read and commented on the manuscript.

250

251 **Author Information**

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256 **Figure captions**

257 Figure 1. Magnesium isotope compositions. a) Probability density plots of Mg isotope
258 compositions from previous standard-sample bracketing work, highlighting the results of individual
259 studies^{5,10-12} that presented numerous analyses of both terrestrial and chondritic samples using the
260 same methodology. These data show systematic subtle differences (0.02-0.05‰) between the Earth
261 and primitive meteorites. Typically authors refrained from interpreting such small differences. b)
262 Samples from this study (measured by critical mixture double spiking) ordered according to sample-
263 type. Lines and shaded bars indicate means and 2se. Samples displayed with pale symbols are
264 excluded from means (see main text). c) Earth and chondrite analyses from b) shown as probability
265 plot to compare with a).

266

267 Figure 2. Median cumulative vapour fractions produced as a function of final planetary mass (in
268 Earth masses M_{\oplus}) determined from high-resolution N-body simulations of planetary accretion²³. The
269 N-body simulations encompassed two scenarios: a calm disk without gas drag (“calm”) and a disk that
270 is disturbed by a Grand Tack motion of Jupiter³⁰ (“GT”). a) Vapour loss fractions calculated for
271 impact vaporisation parameterised to impact velocity⁴. b) Vapour loss fractions produced by direct
272 outflow above exposed magma ponds/oceans. In a Grand Tack scenario, Jupiter’s motions cause
273 higher eccentricities and hence higher impact velocities for such small bodies, which explains their
274 high vapour fractions.

275

276 Figure 3. Comparison between modelled compositions of a vapour depleted liquid and observed
277 terrestrial compositions. a) Changes in isotope compositions (‰/amu) against total relative vapour
278 loss (F_{Total} , in mole fractions) calculated in our model. b) Observed terrestrial Mg (this study), and Si
279 (from [4] and references therein) isotope compositions relative to enstatite chondrites (EH). Errors are
280 2se. c) Loss (mole fraction) of a given element (X), f_X , versus F_{Total} . d) Molar element/Ca of the
281 terrestrial mantle⁷, normalised to EH³¹. Shaded bands give error bounds for F_{Total} inferred from Mg
282 isotope data and the intersection of different curves with this field indicates the range of terrestrial
283 depletions predicted for our vapour loss model. Comparison of these values to those calculated for the
284 Earth relative to an enstatite chondrite starting composition (b and d) is generally good, despite
285 uncertainties in model input parameters (see Methods) and additional influences on observed values
286 from nebular and core formation processes^{4,9}. Left pointing arrows show the effect of post-volatile loss
287 accretion of 20% chondrite (EH).

288

289

290 **Methods**

291 *Chemical and analytical protocols, and data statistics*

292 Powdered samples were digested in cleaned Savillex PFA beakers with 3:1 ultra-pure ~15.5M
293 HNO₃ and ~28M HF. Prior to drying, ~0.5 ml 7M HClO₄ was added to prevent precipitation of
294 insoluble fluorides. At this stage, insoluble chromites were visible in some peridotites. They were
295 separated from the supernatant and further digested in high-pressure acid digestion vessels (Anton
296 Paar Asher) before being re-combined with the rest of the digested sample. Similarly, graphite
297 precipitates were removed from enstatite chondrites by transferring the samples into pre-cleaned BN
298 crucibles that were placed in a muffle furnace at 600 °C, making use of the phase transition of graphite
299 to CO₂ under these conditions. Following complete digestion, any remaining organic components were
300 attacked with ~15.5M HNO₃ and 30% H₂O₂. Terrestrial samples generally weighed 2-15 mg, while
301 larger meteorite samples were dissolved (~7-150 mg, majority 60-100 mg; see Extended Data Table 1)
302 to obtain representative samples.

303 Magnesium was separated from the sample matrix with a Biorad AG50W-X12 cationic exchange
304 resin and 2M HNO₃ eluent as described previously¹¹. Yields were monitored and were >99.85% for all
305 analysed samples. Following separation, samples were dissolved in 0.3M HNO₃ and critically double
306 spiked (see below) for analyses on a Thermo Finnigan Neptune (s/n 1020) multi-collector inductively
307 coupled plasma mass spectrometer (MC-ICPMS). A Savillex PFA nebuliser (30-40 μl min⁻¹ uptake)
308 aspirated the sample solution into an Elemental Scientific Inc. Apex HF introduction system prior to
309 ionisation in the plasma. A combination of Jet sampler and H skimmer cones was used and Mg masses
310 24, 25 and 26 were collected in cups L4, centre and H4, respectively. The Neptune was further
311 operated in medium resolution mode (M/ΔM ≥ 4000, 5-95% peak height definition) with an off-peak
312 centre mass (²⁵Mg mass 24.979 instead of 24.986) to resolve any minor possible ¹²C¹⁴N⁺ interferences
313 on ²⁶Mg⁺. However, these interferences should be insignificant as they were observed to be only ~15
314 ppm of the ²⁶Mg⁺ signal. This medium resolution mode also improved signal-to-noise ratios, reducing
315 total background signals to <2×10⁻¹³ A. A 10¹¹ Ω feedback resistor was used in the amplifier
316 connected to the centre cup, while 10¹⁰ Ω resistors were used on L4 and H4. Total Mg intensities of
317 ~2×10⁻⁹ A were measured at total Mg concentrations of ~2.5 μg ml⁻¹ for 168 s integration (20 cycles of
318 8.4 s), consuming ~185 ng sample Mg per analysis in double-spiked samples.

319 A measurement sequence typically consisted of analysis of a mixture of double spike and DSM-3
320 reference standard followed by three sample – double-spike mixtures, repeating until all samples were
321 measured before starting this series again to make duplicate analyses of samples. Normally, double-
322 spike – sample mixture measurements were repeated four times per sequence, in which a total of 9-17
323 samples were measured. Sequences were repeated in different analysis sessions over the course of the
324 study to yield at least 8 repeated analyses of each double-spike – sample mixture. All sample and
325 standard analyses were preceded by a 240 s wash-out and an on-peak background measurement (168 s
326 integration). The mean intensities of the latter were subtracted from the intensities of the sample or

327 standard analysis that followed it. Additionally, three analyses of unspiked DSM-3 (the concentration
328 matched to the double-spike – sample mixtures) were performed before, after and in the middle of the
329 sequence of double-spike – sample (or standard) mixtures. A 15 minute wash-out was performed
330 between these unspiked and spiked analyses because the Mg isotope composition of the background
331 greatly changes as a consequence of switching between spiked and unspiked analyses, which makes
332 critical mixture double spiking more prone to added variability due to background corrections. In our
333 analyses, background corrections usually affected the reported isotope compositions by <0.0005‰,
334 which is well below the uncertainties of 0.008-0.010‰. Sequences typically took ~10-20 hours.

335 Interferences by hydrides (i.e. $^{24}\text{MgH}^+$ on $^{25}\text{Mg}^+$ and $^{25}\text{MgH}^+$ on $^{26}\text{Mg}^+$), which are difficult to fully
336 resolve, as well background Mg intensities were reduced by using a -1 kV extraction voltage instead
337 of the default -2 kV. We observed that such a change in the extraction voltage reduces background
338 $^{24}\text{MgH}^+$ and $^{25}\text{MgH}^+$ intensities by ~10 times, i.e. ~500 cps $^{24}\text{MgH}^+$ in medium resolution.
339 Simultaneously, the change in extraction voltage lowers background Mg intensities by ~4-5 times,
340 while sample sensitivity is only reduced by <20%. Applying an extraction weaker than -1 kV further
341 decreases backgrounds, but increasingly lowers sensitivity. We interpret these effects as evidence that
342 the background signals are dominantly produced by liberation (with/without hydride) of Mg deposited
343 at the back of the skimmer cone. This hypothesis is substantiated by the fact that progressively
344 increasing Mg background intensities can be lowered by brief cleaning of the back of the skimmer
345 cone with a droplet of dilute HNO_3 . Moreover, background Mg isotope compositions are extremely
346 light after mass bias correction with the same instrumental fractionation factor as for sample Mg,
347 suggesting additional mass fractionation. Finally, we observed that hydride intensities increase over
348 background only ~75 fold when Mg intensities are increased ~25'000 fold by introducing sample Mg.
349 We thus estimate that the largest relative hydride intensity contribution to Mg ($^{24}\text{MgH}^+ / ^{25}\text{Mg}^+$) is ~13
350 ppm. Since we externally normalise samples to intensity matched DSM-3 standards (see *Critical*
351 *mixture double spiking* below), these hydride contributions are reduced to such small levels that they
352 do not affect our reported isotope compositions.

353 Data in Table 1 are reported as relative differences of isotope ratios to the reference solution
354 standard DSM-3³² in delta notation following IUPAC recommendations³³:

355 $\delta^{25/24}\text{Mg} = \left(\left(\frac{^{25}\text{Mg}}{^{24}\text{Mg}} \right)_{\text{Sample}} / \left(\frac{^{25}\text{Mg}}{^{24}\text{Mg}} \right)_{\text{DSM-3}} - 1 \right)$. The data are presented in per mil and
356 represent means of repeated analyses. Reported uncertainties are standard error of the mean ($2s_e$)
357 calculated as $2s/\sqrt{n}$ in which n is the number of repeated analyses for the corresponding sample.
358 Instead of using the standard deviation (s) of a geological reference material (e.g. BHVO-2, $2s =$
359 0.029‰ , $n = 60$ from 3 digestions), we determined s by a homoscedastic approach, pooling over all
360 standards and samples³⁴. That is, after determining the sums of the squared deviations of n repeated
361 measurements from their m sample (and standard) means, these sums were themselves summed before

362 dividing by the degrees of freedom (k): $s^2 = \frac{1}{k} \sum_{i=1}^m \sum_{j=1}^{n_i} (x_{ij} - \bar{x}_i)^2$ with $k = \left(\sum_{i=1}^m n_i \right) - m$. We obtained

363 $2s = 0.029\%$ ($k = 693$) and have found justification in using this scheme because the n repeats of m
364 samples and standards passed the null hypothesis of a Bartlett's test, i.e. the m samples and standards
365 come from normal distributions with the same variance.

366 Group means (e.g. Earth) and their $2se$ uncertainties are calculated directly from the relevant
367 sample data in Table 1. Significances of group mean differences were determined with a one-way
368 Analysis of Variance (one-way ANOVA) or a two sample Student's t-test, again using the relevant
369 data in Table 1. Within group differences were investigated with one-way ANOVA using sample
370 repeat measurements.

371

372 *Critical mixture double spiking*

373 The analysis routine described above is broadly similar to previous studies performed by standard-
374 sample bracketing. Our contribution to addressing previously observed inter-laboratory
375 inconsistencies¹⁸ resides in the novelty of adding a double-spike prior to mass-spectrometric analysis.
376 As detailed elsewhere¹⁹, this technique of "critical mixture double spiking" corrects for mass bias
377 effects of residual matrix, eliminating the most plausible cause for the inter-laboratory inconsistencies.
378 It helps ensure that analyses are not only precise, but also accurate to within the quoted precision. In
379 practice, we dissolved ²⁵MgO and ²⁶MgO spikes (purchased from Oak Ridge National Laboratory,
380 batch numbers 217201 and 158690, respectively) and mixed them to produce a double spike with
381 ^{24,25,26}Mg abundances of 0.30%, 6.27%, 93.43%, approximating those for an accuracy optimised
382 scenario (0%, 8.4%, 91.6%, respectively¹⁹). We calibrated this double-spike to reference solution
383 standard DSM-3 following our suggested protocol¹⁹.

384 After ion exchange chemistry, we aliquoted a weight of separated Mg sufficient for at least 10 runs
385 and mixed this with a weighed amount of double-spike estimated to yield a molar spike proportion of
386 0.5908 in the mixture. This is the critical mixing proportion for our double-spike composition. Prior to
387 analysis, double-spike – sample mixtures were equilibrated at 100 °C for 2-3 h in closed Savillex PFA
388 autosampler vials. Subsequently, all samples were analysed once by MC-ICPMS in a short sequence
389 to determine their deviation from the critical spike proportion. At this stage, the mixtures were rarely
390 within our acceptable bound of $\pm 0.5\%$ molar proportion from critical. Based on the documented
391 sample and double-spike weights and the observed double-spike proportion, either sample or double-
392 spike was added to the mixtures, followed by another short analysis sequence. At this stage most
393 mixtures were found to be within 0.5% from critical.

394 An estimate of the instrumental mass bias is required for critical mixture double spiking¹⁹. We
395 obtained this estimate from the unspiked DSM-3 analyses in the analysis sequence by dividing the
396 measured ^{25/24}Mg by the reference ^{25/24}Mg of 0.12663³⁵. Although this can also be done on ^{26/24}Mg, we
397 find that ^{25/24}Mg is a more reproducible indicator of mass bias because it is less affected by incomplete

398 wash-out due to the smaller contrast between the unspiked and spiked ratios. These instrumental mass
399 bias factors were linearly or quadratically interpolated based on the evolution of DSM-3 – double-
400 spike mixtures to obtain a close mass bias approximation for the sample – double-spike mixtures.
401 Once the mass bias estimate was obtained, the analyses were run through a double-spike inversion
402 routine, followed by a normalisation of sample isotope compositions to DSM-3 isotope compositions
403 (obtained from double-spike – DSM-3 mixtures) to correct for residual non-exponential mass
404 fractionation in the instrument.

405 Our observations are that the mass bias rarely varies by $>0.5\%/amu$ during a 20 h sequence and
406 that this variation tends to be characterized by a smooth evolution. Therefore, we conservatively
407 estimate that the interpolated mass bias is accurate to $\leq 0.25\%/amu$. Combined with our criterion of
408 double-spike proportions being $\leq 0.5\%$ from critical, we calculate that our Mg isotope compositions
409 should be accurate to better than 0.005% on $^{25}\text{Mg}/^{24}\text{Mg}^{19}$. Note that this accuracy estimate is distinct
410 from the reported uncertainty, which is determined from the scatter of replicates.

411 We have also analysed USGS geological reference material BHVO-2 with a deviation from a
412 critical spike proportion of -1% (overspiked) and $+1\%$ (underspiked). The means of these samples
413 are $-0.092\pm 0.007\%$ and $-0.090\pm 0.007\%$, respectively. They are hence not significantly different at
414 this level of precision, giving further confidence that our $\pm 0.5\%$ mixing bound is sufficiently tight for
415 accurate results.

416 Previously published data measured on key reference samples scatter beyond their 2σ
417 uncertainties¹⁸. Assuming sample homogeneity, this scatter highlights the distinction between
418 accuracy and precision, which cannot be assessed separately in the traditional standard-sample
419 bracketing procedure used by those studies. Whilst some of the data may thus appear precise, they can
420 still be inaccurate. In contrast, we have assessed that our new critical double spiking technique should
421 be accurate within our quoted precision. Hence, application of our technique should reduce the scatter
422 in potential future studies from that observed in Extended Data Figure 5 to within the error bounds of
423 the isotope ratios found in this study, assuming sample homogeneity. Nonetheless, much of the
424 literature data has $\delta^{25/24}\text{Mg}$ consistent with our critically double spiked measurements at a $\pm 0.25\%$
425 level (Extended Data Figure 5).

426

427 *N-body collisional accretion simulations*

428 Our N-body collisional accretion simulations have previously been published²³ and the reader is
429 referred to the original publication for model details. These simulations were ideally suited for our
430 investigations because they included growth of planetary bodies from $\sim 10^{-4}$ to $\sim 10^{-1} M_{\oplus}$ with
431 imperfect accretion. We used two different modes of the high resolution simulations²³, which include
432 i) a calm disk, and ii) the Grand Tack³⁰. Whilst we did not modify the simulations in any manner, we
433 extracted additional information (notably impact parameters) to compute accretional vapour loss by
434 two different mechanisms.

435 *1. Impact vaporisation*

436 This information was obtained following published vapour fraction equations (assuming molten
 437 basalt)⁴. The equations enabled determining vapour mass (M_{vap}) as a fraction of the impactor mass
 438 (M_{imp}) in each collision. The vapour mass fraction in each collision (f_{vap}) was calculated from the
 439 impact velocity and angle (v_{imp} and θ), which we have obtained from the N-body simulations
 440 (correcting for the expanded radii that were used in these simulations to increase the collision rate):

441 $f_{vap} = M_{vap} (v_{imp} \sin \theta) / M_{imp}$. The vapour fraction in each collision was subsequently integrated

442 over all collisions a body experienced, taking into account that each planetesimal has its own
 443 collisional history, to obtain the fraction of material that experienced impact induced volatilisation

444 relative to the total mass of a body: $F = (1/M) \int_0^M f_{vap} (v_{imp} \sin \theta) dm$. This equation is modified from

445 the original⁴ by replacing the $v_{esc}(m)$ term, which was a proxy for impact velocity, by the impact
 446 velocities (v_{imp}) obtained in our N-body simulations. This replacement resulted in higher volatilisation
 447 fractions in our study compared to Dauphas et al.⁴, particularly for the Grand Tack simulation in which
 448 higher impact velocities occur due to dynamical excitation by Jupiter's migration. We obtained Figure
 449 2 by binning the results for F in groups of masses (relative to Earth's mass M_{\oplus}) in each of the two
 450 simulation modes.

451

452 *2. Vapour mass loss by direct outflow above exposed magma ponds*

453 *Summary:* The overall picture is as follows. We use the impact information from the N-body
 454 collisional accretion simulations and let each impact generate a hemispheric pool of melt. Vapour
 455 forms above the melt pool, with a pressure determined by the melt pool temperature. The melt vapour
 456 escapes at a rate determined by its temperature and the gravity of the body. The melt pool is assumed
 457 to be convecting vigorously, keeping it isothermal, and it cools from the surface due to evaporation
 458 and radiative heat transfer. As the melt cools, the vapour pressure drops and the vapour mass loss rate
 459 decreases until it effectively stops.

460 This overall behaviour is described by an energy equation as follows:

461
$$mC_p \dot{T} = \dot{m} \left(\frac{GM}{R_s} + L_v \right) + A\sigma T_{eff}^4 \quad (1)$$

462 where the left-hand side denotes the energy source (secular cooling of the melt pool) and the right-
 463 hand side denotes the energy sinks (evaporative cooling, gravitational potential energy and radiative
 464 heat transfer). Here m is the mass of the melt pool, C_p is its specific heat capacity, \dot{T} is the cooling
 465 rate of the melt pool, \dot{m} is the vapour mass loss rate, M is the mass of the body, R_s is its radius, G is
 466 the gravitational coefficient, L_v is the latent heat of vaporization, σ is Stefan's constant, T_{eff} is the
 467 effective radiating temperature of the melt pool and A is its surface area. We take $C_p = 1200 \text{ J kg}^{-1} \text{ K}^{-1}$,
 468 $L_v = 5 \text{ MJ kg}^{-1}$ and the magma density and body bulk density to both be 3000 kg m^{-3} .

469 Below we describe the different energy sources and sinks, and our calculation methods, in more
 470 detail. We have deliberately tried to keep our approach relatively simple, but even so there are a large
 471 number of parameters, some of which are highly uncertain. That being the case, a more complicated
 472 model seems premature; our main aim is to demonstrate that vapour loss from melt ponds is a
 473 plausible mechanism for generating vapour mass loss, and thereby isotopic fractionation.

474 *Mass loss rate:* Atmospheres which have a gas thermal velocity comparable to the escape velocity of
 475 the body are vulnerable to loss by direct outflow³⁶. For Earth-mass bodies the thermal velocity
 476 required is implausibly high (except perhaps for close-in “roaster” planets that may occur in
 477 exoplanetary systems³⁷) but for small bodies this requirement is less restrictive. The resulting outflow
 478 can be described by the equations of mass and momentum conservation:

$$479 \quad \frac{1}{\rho} \frac{d\rho}{dr} + \frac{1}{v} \frac{dv}{dr} + \frac{2}{r} = 0 \quad (2)$$

$$480 \quad \frac{1}{\rho} \frac{dP}{dr} + v \frac{dv}{dr} = -\frac{GM}{r^2} \quad (3)$$

481 where r is the radial coordinate, $P(T)$, $\rho(T)$ and v are the pressure, density and radial velocity of the
 482 gas and we are assuming spherical geometry. For an ideal gas we have $R'T = P/\rho = v_s^2$ where R' is
 483 the gas constant divided by the molar mass, T is the gas temperature and v_s is the sound speed. To
 484 close the system, we also need an energy equation. The simplest, which we will adopt here, is to
 485 assume an isothermal atmosphere with the temperature set by the melt pool surface temperature T . A
 486 more realistic description would be to assume a vapour-saturated atmosphere; Lehmer et al.³⁸ have
 487 shown that this latter system is actually well-approximated by the isothermal description.

488

489 Under these assumptions, the gas will accelerate outwards until it reaches the “sonic point” at
 490 which $v=v_s$. At this critical radius r_c the sonic velocity is half the local escape velocity, with

$$491 \quad r_c = \frac{GM}{2R'T} = R_s \left(\frac{v_{esc}}{2v_s} \right)^2$$

492 and v_{esc} is the escape velocity at the body surface ($r=R_s$). The atmospheric mass loss rate is then
 493 given by $4\pi r_c^2 \rho(r_c) v_s$. Given the surface density, equations (2) and (3) may then be integrated
 494 outwards to calculate $\rho(r_c)$ and thus the mass loss rate. For small bodies, r_c may be equal to or less than
 495 R_s , in which case the mass loss rate is $4\pi R_s^2 \rho(R_s) v_s$. In our case the gas density at the surface $\rho(r_s)$
 496 is set by the vapour pressure of the melt pool. Because only the melt pool is losing vapour, we replace
 497 $4\pi R_s^2$ with A , the melt pool area, in our mass loss calculations. We assume a molar mass of 0.04
 498 kg/mol, appropriate for MgO or SiO.

499 Driving mass loss requires energy because of two effects (equation 1): conversion of liquid to
 500 vapour takes latent heat, and removal of mass from the surface to large distances involves potential
 501 energy changes. We consider both these effects in more detail below.

502 *Secular cooling and radiative heat transfer:* Secular cooling of the melt pool provides the energy to
 503 drive mass loss. The melt pool is assumed isothermal (temperature T) due to vigorous convection. This
 504 temperature sets the gas vapour pressure at the surface. Here we take the vapour pressure to be given
 505 by the following empirical expression obtained from our thermodynamic liquid-vapour model (see
 506 below):

$$507 \quad \ln P = -4.0041 (\ln T)^3 + 88.788 (\ln T)^2 - 639.3 \ln T + 1480.23$$

508 where P is in bars.

509 An important aspect of our model is that the effective radiating temperature of the melt pool T_{eff} is
 510 lower than the temperature T which sets the gas vapour pressure and defines the internal temperature
 511 of the melt. This effect is observed at terrestrial lava lakes, where the surface temperature is typically
 512 many hundreds of K cooler³⁹ than the measured subsurface magma temperature and that inferred from
 513 the gas equilibrium chemistry⁴⁰. A similar effect is found at Io, where the majority of a lava lake's
 514 surface is at much lower temperature than the expected magma temperature⁴¹.

515 The reason for this effect is that there is a thin conductive skin on top of the convecting magma,
 516 which is at a lower temperature than the material beneath. To approximately describe this
 517 phenomenon, we set the heat flux conducted across this layer equal to the heat flux radiated from the
 518 top surface:

$$519 \quad F = \frac{k(T - T_{eff})}{\delta} = \sigma T_{eff}^4$$

520 where k is the thermal conductivity and δ is the skin thickness. This allows us to solve for T_{eff} given
 521 T and δ . To calculate δ we assume the convecting melt pool can be described by isoviscous
 522 convection⁴²

$$523 \quad \delta \approx 2 \left(\frac{\kappa \eta}{\rho g \alpha [T - T_{eff}]} \right)^{1/3}$$

524 At high temperatures, the dominant energy sink in equation (1) is evaporative cooling, because
 525 radiative heat loss increases as T^4 , while mass loss (which depends on the vapour pressure) increases
 526 exponentially with T . Thus, the overall mass loss calculations are not particularly sensitive to the
 527 details of the radiative cooling calculations.

528 Here we take $\alpha = 3 \times 10^{-5} \text{ K}^{-1}$, $\eta = 10^3 \text{ Pa s}$. The thermal conductivity k is taken to be $3 \text{ W m}^{-1} \text{ K}^{-1}$ which
 529 then specifies κ . For $T = 1400 \text{ K}$ and a 1000 km radius body we find $T_{eff} = 853 \text{ K}$, which is roughly
 530 consistent with terrestrial lava lake measurements³⁹.

531 *Melt pool volume and temperature:* To calculate the melt pool volume and temperature we assume
 532 that the volume of material shocked to the maximum temperature (the isobaric core) is hemispherical
 533 and has a radius linearly proportional to the impactor radius: $R_{ic} = f R_i$, with $f = 3$ following
 534 literature^{4,43}. Assuming that some fraction of the kinetic energy of the impactor is consumed within
 535 this volume, the temperature change associated with impact is then given by

$$\Delta T = \frac{1}{C_p} \left[\left(\frac{R_i}{R_{ic}} \right)^3 \beta v_{imp}^2 - L_m \right] \quad (4)$$

where L_m is the latent heat of fusion, v_{imp} is the impact velocity (given by the N-body simulations) and β is a factor to account for energy delivered beyond the isobaric core. Here we take $\beta=0.5$ to provide a conservative estimate of the initial melt temperature. Inspection of this equation shows that a minimum velocity of about 4.6 km/s is required to initiate significant mass loss.

We assume that the pre-impact target material is at the solidus, here taken to be 1400 K. The justification for doing so is that heating by ^{26}Al will melt bodies that formed early enough (to be conservative, we ignore at this stage that such melting induced by ^{26}Al heating could have led to vapour mass loss). Melt advection will rapidly cool the target to 1400 K, but subsequent cooling (by conduction or solid-state convection) will be much slower. We also impose a maximum melt temperature of 3500 K. This is approximately the temperature at which total silicate vaporization is achieved and the physics of mass loss will change. Imposing this cut-off will underestimate the total amount of mass loss.

Although the radius of the isobaric core is somewhat uncertain, varying f does not have a very large effect on our results, because of the trade-offs involved. A smaller isobaric core reduces the surface area and cooling timescale, but it increases the melt temperature (and thus the vapour pressure and the initial mass loss rate). By contrast, the difference between regular and Grand Tack accretion is dramatic. This is because the Grand Tack results in more dynamical excitation, and thus more energetic collisions.

Extended Data Figure 6 plots the median impact velocity normalized to the escape velocity as a function of body radius. The biggest difference between a calm disk (without gas drag) and Grand Tack accretion is that the normalized impact velocities for large bodies are much bigger in the latter. Since no vapour loss happens below a particular cut-off velocity (equation 4), it is clear that Grand Tack simulations will generate much more vapour loss. Increasing β to 1 causes only modest increases in the predicted mass loss for conventional accretion, because impact velocities are still general below the cut-off value.

Implementation: Each body undergoes mass growth by collision; each collision may also lead to some mass loss via vapour outflow. For each impact, we calculate the initial melt pool mass and temperature as outlined in the section *Melt pool volume and temperature*. The instantaneous mass loss rate is calculated using as described in the section *Mass loss rate*. Given the mass loss rate, the change in melt pool temperature is calculated according to equation (1). The temperature is updated accordingly and the cycle then repeats. The total mass lost is tracked.

A disadvantage of our approach is that the pre-history of impactors striking a target body is not recorded. Since these impactors have likely undergone mass loss of their own, our approach may therefore underestimate the total fractional mass loss of the final body. To mitigate this problem, for the Grand Tack results we perform a two-step calculation. In the first step we calculate the total mass

572 loss for each body neglecting inheritance effects. We use these results to determine the median mass
 573 loss over all bodies within a particular final mass range. The median prior fractional mass loss ϕ
 574 determined in this manner for the GT bodies is given by the following empirical formula:

$$575 \quad \phi = 0.4 \times 10^{-5} R_s^{1.65} e^{-0.65 \times 10^{-5} R_s^{1.65}}$$

576 where R_s is in km. We then re-run the calculations, this time assuming that each impactor has
 577 experienced prior mass loss based on this median value. The results of both calculations are shown in
 578 Extended Data Figure 7.

579 For the conventional accretion simulations without gas present in the disk (“calm disk”) vapour
 580 loss is sufficiently small that we neglected this inheritance effect. Extended Data Figure 7 plots the
 581 median fractional mass loss for all bodies in the Grand Tack accretion simulations. The simulations for
 582 a calm disk result in very limited mass loss, and only for the largest (see Figure 2b in main text). By
 583 contrast, mass loss in the Grand Tack scenario is much more extensive.

584

585 *Vapour-liquid Mg isotope fractionation models*

586 We first consider a kinetic isotope fractionation scenario. For this, we modelled the elemental loss
 587 of Mg required to obtain $\sim 0.02\%$ heavier Mg isotope compositions by a Rayleigh distillation process
 588 with an experimentally determined kinetic isotope fractionation factor⁴⁴ (referred to as $\alpha^{25/24} Mg_{g/liq}^{kin}$)
 589 of 0.9860. We obtained a relative mass loss of Mg by vaporization (f_{Mg}^{vap}) of 0.14% from

$$590 \quad f_{Mg}^{vap} = 1 - e^{\delta^{25/24} Mg / (\alpha^{25/24} Mg_{g/liq}^{kin} - 1)}$$

591 This is a very small mass loss, but we can expect much larger
 592 concomitant K mass losses, owing to its higher volatility. Although it is hard to quantify such K loss,
 593 we can estimate it from an equilibrium liquid-vapour model (see below), given that kinetic evaporative
 594 fluxes (J) from a molten surface are proportional to their saturation vapour pressures as well as the
 595 elemental mass⁴⁴. In the simplest scenario, i.e. assuming a single vaporisation event, we find 0.14% of
 596 Mg is vaporised at ~ 2700 K, at which temperature $\sim 28\%$ of K occurs as vapour (all as atomic K).
 597 Using a published⁴⁵ $\alpha^{40/39} K_{g/liq}^{kin}$ of 0.9892 we obtain that $^{41}K/^{39}K$ fractionates by $\sim 7\%$. This compares
 598 to $<2.7 \pm 1\%$ observed fractionation in Solar System material²⁵, underscoring the implausibility of
 kinetic fractionation.

599 We also examine an equilibrium liquid-vapour fractionation model. Vapour production results in
 600 isotopically light vapour and heavy liquid compositions, thus evolving a vapour depleted bulk
 601 planetesimal to isotopically heavier compositions. The composition of silicate vapour in equilibrium
 602 with a molten planetesimal was calculated with a thermodynamic model involving equilibrium
 603 reactions between silicate liquid and vapour species for nine elements: O, Na, Mg, Al, Si, K, Ca, Ti
 604 and Fe. The model is modified from the MAGMA code by Fegley and co-workers^{46,47}. We followed
 605 the approach of these authors to non-ideal mixing in the silicate liquid (i.e. assuming ideal mixing of
 606 complex oxide and silicate pseudo-components), used their tabulated thermodynamic data, and, like

607 them, we assumed ideal mixing in the gas. Our approach differs from theirs in that we let go of their
608 constraint that oxygen-to-metal ratios in the gas must be identical to the original liquid. Instead, we
609 used the ideal gas law to convert partial pressures of gas species to number of moles in order to
610 enforce conservation of mass by simultaneously solving mass balance equations for the number of
611 moles of each element. Hence, we have 66 equilibrium equations and nine mass balance equations to
612 solve for activities of 75 species. The disadvantage of our approach is that the ideal gas law adds one
613 freely adjustable parameter in the form of the volume the gas can expand into. We have chosen to fix
614 this volume such that vapour fractions between ~ 0.001 and ~ 0.15 were obtained between 2500 K and
615 3500 K. Such fractions are of similar order of magnitude to the average vapour fraction produced in
616 single collisions in the N-body simulations (note that an order of magnitude variation in the chosen
617 volume leads to ~ 5 fold variation in the vapour fractions but these changes do not affect our
618 conclusions because their effect is counterbalanced by the number of “collisions” that are run to obtain
619 the observed Mg isotope composition; see below). To convert from activities of the liquid species to
620 moles, which we need for the mass balance equations, we need one more variable, i.e. the total moles
621 of all liquid species, bringing the number of unknowns to 76. However, there is the activity
622 normalisation constraint: the activities (i.e. mole fractions) of liquid species must sum to 1. Hence, we
623 have 76 constraints to solve for 76 variables. The code is easily expandable with extra elements and
624 equilibrium reactions and is freely available from the authors upon request.

625 In our equilibrium liquid-vapour fractionation model, we used published reduced partition
626 functions (β) to approximate equilibrium Mg, Si and Fe isotope fractionation factors between silicate
627 liquid and atomic silicate vapour. Experimental calibrations of equilibrium isotope fractionation
628 factors between vapour and liquid do not exist to date. As β are currently unavailable for silicate
629 liquid, we approximated them with those for forsterite crystals for Mg⁴⁸ and Si⁴⁹ and fayalite crystals
630 for Fe⁵⁰. We used further literature sources for β for other vapour species MgO_(g)⁴⁸, FeO_(g)⁵¹ and
631 SiO_(g)^{49,52}. Relative abundances of species in mixed vapours, e.g. Mg(g) and MgO(g), were obtained
632 from the thermodynamic model described above. A two-component mass balance relationship
633 combining isotope compositions of the vapour (subscript *g*) and the remaining ‘liquid’ (subscript *liq*)
634 were used to determine the offset relative to the original bulk composition (set to 0 for reference here)
635 prior to loss of a vapour fraction (f_g): $\delta^{25/24}Mg_{liq} = -f_g \delta^{25/24}Mg_g / (1 - f_g)$.

636 We used this equilibrium vapour-liquid fractionation model to simulate vapour loss as various
637 temperature events (i.e. planetary collisions). We initiated the model with a chondritic elemental
638 composition (see main text) and isotope compositions of 0, as mentioned above. We removed 95% of
639 the chondritic Fe assuming it segregated into a core which is not affected by vaporisation. We let a
640 single “collision” affect 20% of the body and assumed that the body undergoes full chemical
641 homogenisation between two events. After the first “collision”, the calculated isotopic and elemental
642 compositions of the residual liquid were then used as initial values for the next “collision”, and so
643 forth. The temperature used to calculate equilibrium vapour-liquid compositions and isotopic

644 fractionation in any “collision” event was obtained by random selection of a temperature uniformly
645 distributed between 2500 K and 3500 K.

646

647

648 **Additional references**

- 649 32 Galy, A. *et al.* Magnesium isotope heterogeneity of the isotopic standard SRM980 and new
650 reference materials for magnesium-isotope-ratio measurements. *J. Anal. At. Spectrom.* **18**,
651 1352-1356, doi:10.1039/b309273a (2003).
- 652 33 Coplen, T. B. Guidelines and recommended terms for expression of stable-isotope-ratio and
653 gas-ratio measurement results. *Rapid Commun Mass Sp* **25**, 2538-2560, doi:10.1002/rcm.5129
654 (2011).
- 655 34 Regelous, M., Elliott, T. & Coath, C. D. Nickel isotope heterogeneity in the early Solar
656 System. *Earth Planet. Sci. Lett.* **272**, 330-338, doi:10.1016/j.epsl.2008.05.001 (2008).
- 657 35 Catanzaro, E. J., Murphy, T. J., Garner, E. L. & Shields, W. R. Absolute isotopic abundance
658 ratios and atomic weight of magnesium. *Journal of Research of the National Bureau of*
659 *Standards Section - A Physics and Chemistry* **70A**, 453-458 (1966).
- 660 36 Pierrehumbert, R. T. *Principles of planetary climate*. (Cambridge University Press, 2010).
- 661 37 Perez-Becker, D. & Chiang, E. Catastrophic evaporation of rocky planets. *Mon Not R Astron*
662 *Soc* **433**, 2294-2309, doi:10.1093/mnras/stt895 (2013).
- 663 38 Lehmer, O. R., Catling, D. C. & Zahnle, K. J. The longevity of water ice on Ganymedes and
664 Europas around migrated giant planets. *The Astrophysical Journal* **839**, 32-40,
665 doi:10.3847/1538-4357/aa67ea (2017).
- 666 39 Patrick, M. R., Orr, T., Swanson, D. A. & Lev, E. Shallow and deep controls on lava lake
667 surface motion at Kīlauea Volcano. *J. Volcanology Geothermal Res.* **328**, 247-261,
668 doi:10.1016/j.jvolgeores.2016.11.010 (2016).
- 669 40 Moussallam, Y. *et al.* Hydrogen emissions from Erebus volcano, Antarctica. *B. Volcanol.* **74**,
670 2109-2120, doi:10.1007/s00445-012-0649-2 (2012).
- 671 41 Davies, A. G. *Volcanism on Io: A comparison with Earth*. (Cambridge University Press,
672 2007).
- 673 42 Solomatov, V. S. Scaling of temperature- and stress-dependent viscosity convection. *Physics*
674 *of Fluids* **7**, 266-274, doi:10.1063/1.868624 (1995).
- 675 43 Melosh, H. J. *Impact cratering: A geologic process*. (Oxford University Press, 1989).
- 676 44 Richter, F. M., Janney, P. E., Mendybaev, R. A., Davis, A. M. & Wadhwa, M. Elemental and
677 isotopic fractionation of Type B CAI-like liquids by evaporation. *Geochim. Cosmochim. Acta*
678 **71**, 5544-5564, doi:10.1016/j.gca.2007.09.005 (2007).
- 679 45 Richter, F. M., Mendybaev, R. A., Christensen, J. N., Ebel, D. & Gaffney, A. Laboratory
680 experiments bearing on the origin and evolution of olivine-rich chondrules. *Meteorit. Planet.*
681 *Sci* **46**, 1152-1178, doi:10.1111/j.1945-5100.2011.01220.x (2011).
- 682 46 Fegley, B. & Cameron, A. G. W. A vaporization model for iron/silicate fractionation in the
683 Mercury protoplanet. *Earth Planet. Sci. Lett.* **82**, 207-222, doi:10.1016/0012-821x(87)90196-8
684 (1987).
- 685 47 Schaefer, L. & Fegley, B. A thermodynamic model of high temperature lava vaporization on
686 Io. *Icarus* **169**, 216-241, doi:10.1016/j.icarus.2003.08.023 (2004).
- 687 48 Schauble, E. A. First-principles estimates of equilibrium magnesium isotope fractionation in
688 silicate, oxide, carbonate and hexaaquamagnesium(2+) crystals. *Geochim. Cosmochim. Acta*
689 **75**, 844-869, doi:10.1016/j.gca.2010.09.044 (2011).
- 690 49 Huang, F., Wu, Z. Q., Huang, S. C. & Wu, F. First-principles calculations of equilibrium
691 silicon isotope fractionation among mantle minerals. *Geochim. Cosmochim. Acta* **140**, 509-
692 520, doi:DOI 10.1016/j.gca.2014.05.035 (2014).
- 693 50 Polyakov, V. B. & Mineev, S. D. The use of Mössbauer spectroscopy in stable isotope
694 geochemistry. *Geochim. Cosmochim. Acta* **64**, 849-865, doi:10.1016/S0016-7037(99)00329-4
695 (2000).

696 51 Pahlevan, K. *Chemical and isotopic consequences of lunar formation via giant impact* PhD
697 thesis, California Institute of Technology, (2010).

698 52 Javoy, M., Balan, E., Méheut, M., Blanchard, M. & Lazzeri, M. First-principles investigation
699 of equilibrium isotopic fractionation of O- and Si-isotopes between refractory solids and gases
700 in the solar nebula. *Earth Planet. Sci. Lett.* **319**, 118-127, doi:10.1016/j.epsl.2011.12.029
701 (2012).

702 53 Clayton, R. N. & Mayeda, T. K. Oxygen isotope studies of carbonaceous chondrites.
703 *Geochim. Cosmochim. Acta* **63**, 2089-2104, doi:10.1016/S0016-7037(99)00090-3 (1999).

704 54 Van Schmus, W. R. & Wood, J. A. A chemical-petrologic classification for the chondritic
705 meteorites. *Geochim. Cosmochim. Acta* **31**, 747-765, doi:10.1016/S0016-7037(67)80030-9
706 (1967).

707 55 Sossi, P. A., Nebel, O. & Foden, J. Iron isotope systematics in planetary reservoirs. *Earth*
708 *Planet. Sci. Lett.* **452**, 295-308, doi:10.1016/j.epsl.2016.07.032 (2016).

709 56 O'Neill, H. S. C. & Palme, H. Collisional erosion and the non-chondritic composition of the
710 terrestrial planets. *Phil. Trans. Roy. Soc. A* **366**, 4205-4238, doi:10.1098/rsta.2008.0111
711 (2008).

712 57 Badro, J. *et al.* Effect of light elements on the sound velocities in solid iron: Implications for
713 the composition of Earth's core. *Earth Planet. Sci. Lett.* **254**, 233-238,
714 doi:10.1016/j.epsl.2006.11.025 (2007).

715 58 Hin, R. C., Fitoussi, C., Schmidt, M. W. & Bourdon, B. Experimental determination of the Si
716 isotope fractionation factor between liquid metal and liquid silicate. *Earth Planet. Sci. Lett.*
717 **387**, 55-66, doi:10.1016/j.epsl.2013.11.016 (2014).

718 59 Armytage, R. M. G., Georg, R. B., Savage, P. S., Williams, H. M. & Halliday, A. N. Silicon
719 isotopes in meteorites and planetary core formation. *Geochim. Cosmochim. Acta* **75**, 3662-
720 3676, doi:10.1016/j.gca.2011.03.044 (2011).

721 60 Schiller, M., Handler, M. R. & Baker, J. A. High-precision Mg isotopic systematics of bulk
722 chondrites. *Earth Planet. Sci. Lett.* **297**, 165-173, doi:10.1016/j.epsl.2010.06.017 (2010).

723 61 Schiller, M., Baker, J. A. & Bizzarro, M. ²⁶Al-²⁶Mg dating of asteroidal magmatism in the
724 young Solar System. *Geochim. Cosmochim. Acta* **74**, 4844-4864,
725 doi:10.1016/j.gca.2010.05.011 (2010).

726 62 Larsen, K. K. *et al.* Evidence for Magnesium Isotope Heterogeneity in the Solar
727 Protoplanetary Disk. *Astrophys J Lett* **735**, doi:10.1088/2041-8205/735/2/L37 (2011).

728

729

730

731 **Data availability**

732 Source data for Table 1 and all Figures not generated by modelling data have been provided with
733 the paper. Data generated by modelling code can be made available from the corresponding author
734 upon reasonable request. Magnesium isotope compositions generated in this study are available in the
735 EarthChem repository.

736

737

738 **Extended Data Table and Figure captions**

739

740 Extended Data Table 1. Sample sources and weights of digested sample from which aliquots were
741 taken for Mg isotope analysis in this study. NHM is Natural History Museum, London, UK. NASA is
742 National Aeronautics and Space Administration. ¹Aliquot from a homogenised powder of several
743 grams.

744

745 Extended Data Figure 1. Magnesium isotope compositions of carbonaceous chondrites plotted
746 against their average literature oxygen isotope compositions⁵³. These mass-dependent oxygen isotope
747 measurements reflect parent body hydrothermal alteration²⁰, so the correlation ($R^2 = 0.78$) between Mg
748 and O isotopes (as well as with petrographic group⁵⁴; indicated in brackets under sample names)
749 implies that the Mg isotope compositions of some carbonaceous chondrites have been altered by
750 hydrothermal processes. The most altered samples, to the upper right of this diagram, are excluded
751 from our chondrite Mg isotope means.

752

753 Extended Data Figure 2. Magnesium isotope compositions of peridotites plotted against whole rock
754 MgO (panel a) and Al₂O₃ (panel b) contents. The absence of correlations of Mg isotope compositions
755 with MgO or Al₂O₃ indicates absence of discernible Mg isotope fractionation during partial melting.

756

757 Extended Data Figure 3. Comparison between modelled compositions of a vapour depleted liquid
758 and observed planetary compositions. As Figure 3 in main text, but additionally including observed
759 isotope compositions for Mars, and eucrite and angrite parent bodies as well as including elemental
760 and isotopic Fe observations (panels b and d; Fe isotope data from [55] and references therein, all
761 other references as in Figure 3). Comparison of observed Fe contents and isotope ratios are
762 complicated by core formation because most Fe enters the core. In our model we assume that the iron
763 in the core has not been affected by vaporisation, inferred to occur later. For instance, the effect of
764 ~48% Fe loss (panel c) on the current bulk silicate Earth Fe content is dependent on the fraction of Fe
765 that entered the core prior to collisional vaporisation and the oxygen fugacity evolution of the growing
766 Earth. For reference, the datum labelled Fe** in panel d is therefore the Fe/Ca of the bulk Earth
767 (calculated from [56]) instead of the Fe/Ca of the bulk silicate Earth. Similarly, Si can also enter the

768 core, although its quantity is likely <3 wt%⁵⁷. Right pointing arrows in b) and d) indicate the effect of
769 3 wt% Si in the core (3000 K assumed for metal-silicate Si isotope fractionation factor⁵⁸).

770

771 Extended Data Figure 4. Comparison between modelled compositions of a vapour depleted liquid
772 and observed planetary compositions. Similar to Extended Data Figure 3, but for model runs with a CI
773 chondrite initial composition. Observed Mg and Fe isotope compositions (panel b) are presented
774 relative chondrite mean, while Si isotope observations are relative to a mean of carbonaceous and
775 ordinary chondrites⁵⁹, because those chondrites have undistinguishable Si isotope compositions, yet
776 are distinctly different from enstatite chondrites (see [4] and references therein).

777

778 Extended Data Figure 5. Magnesium isotope compositions of reference samples analysed in
779 multiple studies. The shaded areas show the mean and 2se of the isotope compositions observed in this
780 study. Note that the plotted composition of Murchison for Bourdon et al.¹⁰ is a mean of the two
781 replicates presented in their Table 1. The value for BHVO of Chakrabarti and Jacobsen¹⁶ is BHVO-1,
782 all others are BHVO-2.

783

784 Extended Data Figure 6. Variation in velocity of individual impacts (normalized by target body
785 escape velocity) as a function of target body radius. Central line denotes median value, shaded box
786 encompasses region spanning 25th-75th percentiles, upper lines denote 90th percentile. Bulk density
787 assumed to be 3000 kg m⁻³.

788

789 Extended Data Figure 7. Fractional mass loss in Grand Tack simulation as a function of final body
790 radius for direct vapour outflow model to illustrate results both with (white boxes, as Figure 2b) and
791 without (shaded boxes) the inclusion of inheritance effects (see Methods). Boxes denote the median
792 value, bars denote the 25th and 75th quartiles.

793

795 Table 1. Magnesium isotope compositions of chondrites, terrestrial (ultra-)mafic, and achondrites.

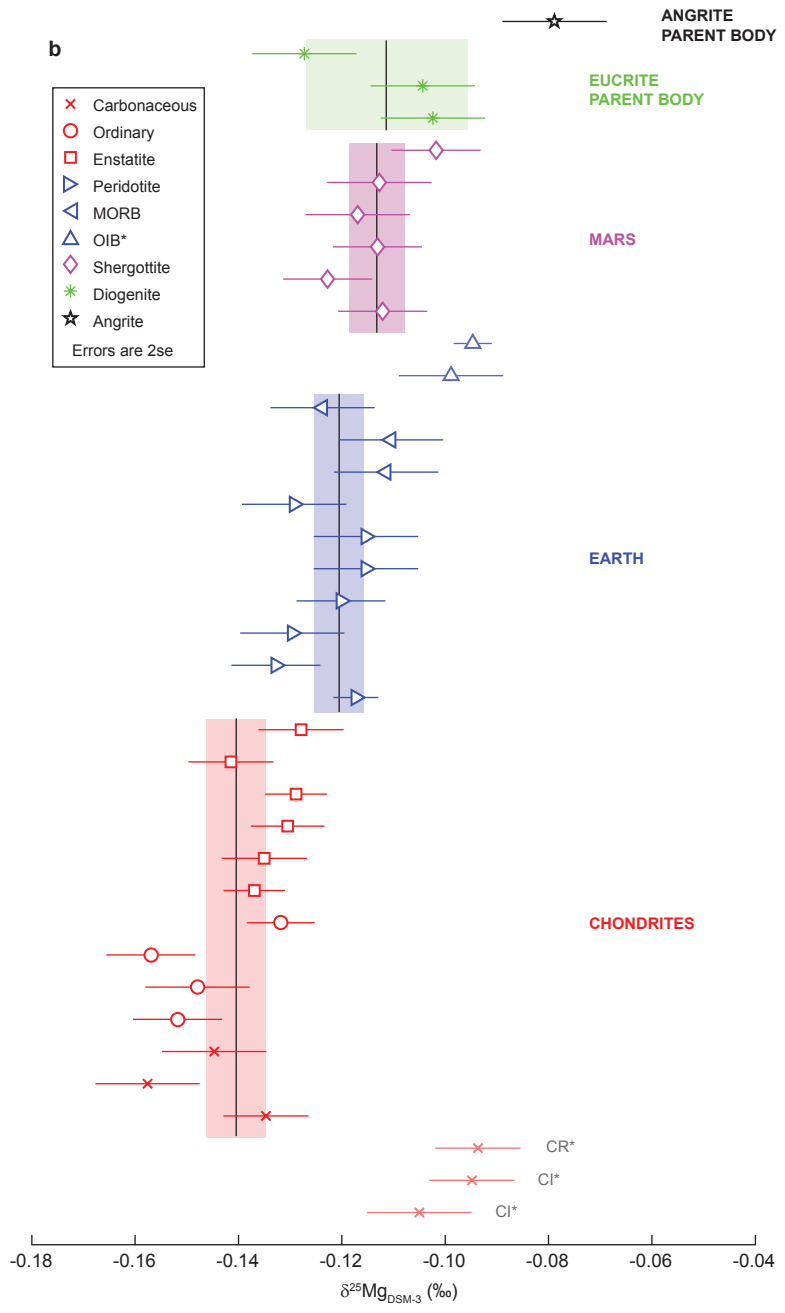
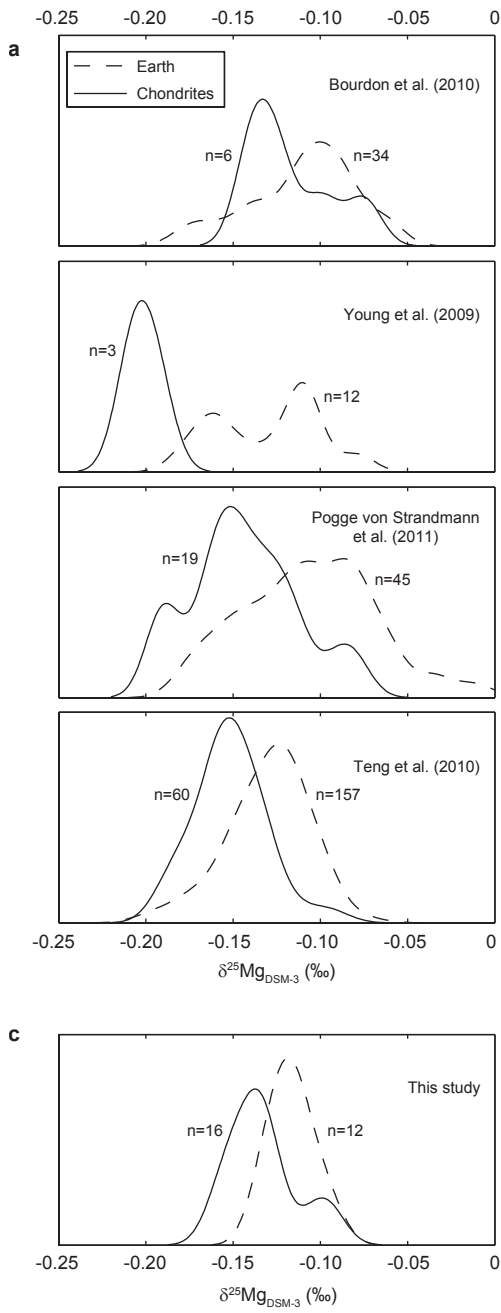
Sample	Type	$\delta^{25}\text{Mg}$ (‰)	2se	n ¹
Orgueil	C11	-0.105	0.010	8
Ivuna	C11	-0.095	0.008	12
Al Rais	CR2	-0.094	0.008	12
Renazzo	CR2	-0.135	0.008	12
Kainsaz	CO3	-0.158	0.010	8
Murchison	CM2	-0.145	0.010	8
Parnallee	LL3	-0.152	0.009	11
Dhurmsala	LL6	-0.148	0.010	8
Ceniceros	H3	-0.157	0.009	11
Zag	H3	-0.132	0.007	19,2
Khairpur	EL6	-0.137	0.006	23,2
Hvittis	EL6	-0.135	0.008	12
Yilmia	EL6	-0.130	0.007	16
St Mark's	EH5	-0.129	0.006	23,2
Abee	EH4	-0.142	0.008	12
Indarch	EH4	-0.128	0.008	12
<i>Mean Chondrites</i> ²		<i>-0.140</i>	<i>0.006</i>	
JP-1	Dunite	-0.117	0.004	43,5
BZ116	Sp Hz	-0.133	0.009	11
Vi 313-102	Gt Lz	-0.130	0.010	8
BZ143	Sp Lz	-0.120	0.009	11
BZ-29	Sp Lz	-0.115	0.010	8
Vi 314-56	Sp Lz	-0.115	0.010	8
BZ251	Plg Lz	-0.129	0.010	8
2370-1	MORB	-0.111	0.010	8
DS12-29	MORB	-0.110	0.010	8
9/30a(1)	MORB	-0.124	0.010	8
BIR-1	OIB	-0.099	0.010	8
BHVO-2	OIB	-0.095	0.004	60,3
<i>Mean Earth</i> ³		<i>-0.121</i>	<i>0.005</i>	
ALHA 77005	Lz Sherg	-0.112	0.009	11
DaG 476	Bas Sherg	-0.123	0.009	11
EETA 79001	Bas Sherg	-0.113	0.009	11
RBT 04262	Ol Sherg	-0.117	0.010	8
Zagami	Bas Sherg	-0.113	0.010	8
LAR 06319	Ol Sherg	-0.102	0.009	11
<i>Mean Mars</i>		<i>-0.113</i>	<i>0.006</i>	
Johnstown	Diogenite	-0.102	0.010	8
Shalka	Diogenite	-0.104	0.010	8
Tatahouine	Diogenite	-0.127	0.010	8
<i>Mean Diogenites</i>		<i>-0.111</i>	<i>0.016</i>	
d'Orbigny	Angrite	-0.079	0.010	8

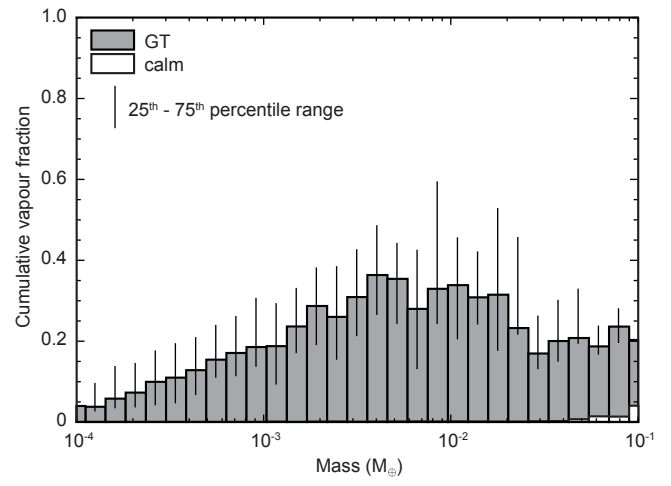
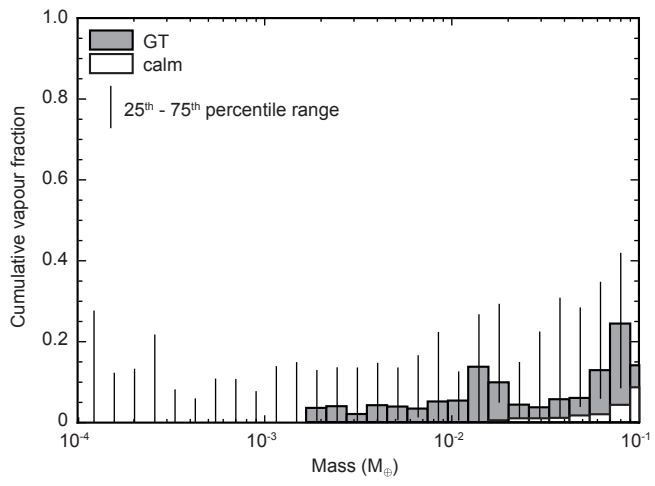
¹ Total number of analyses. If a comma is present, the number behind the comma refers to the number of digestions. Each digestion was passed through chemistry once.

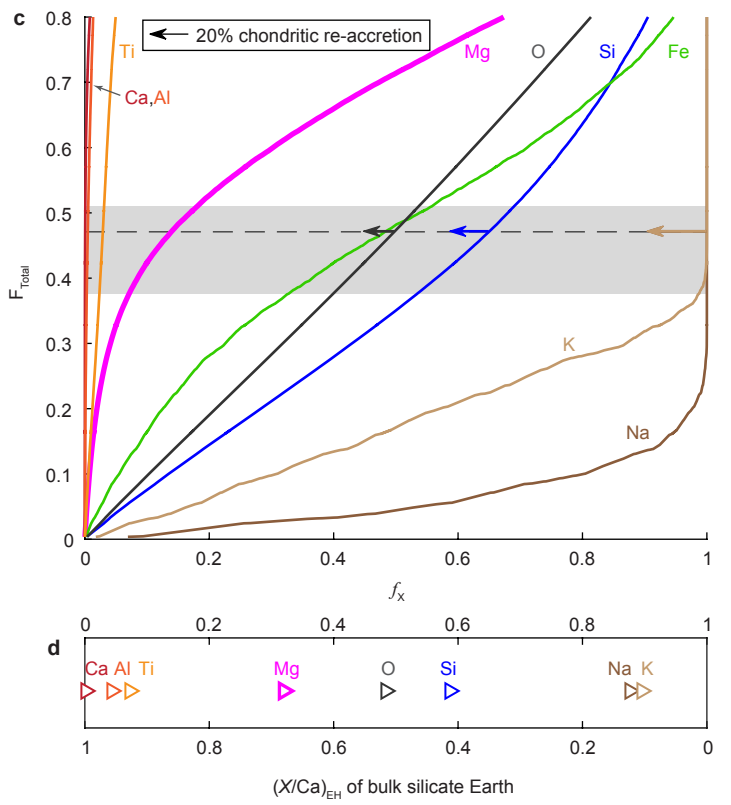
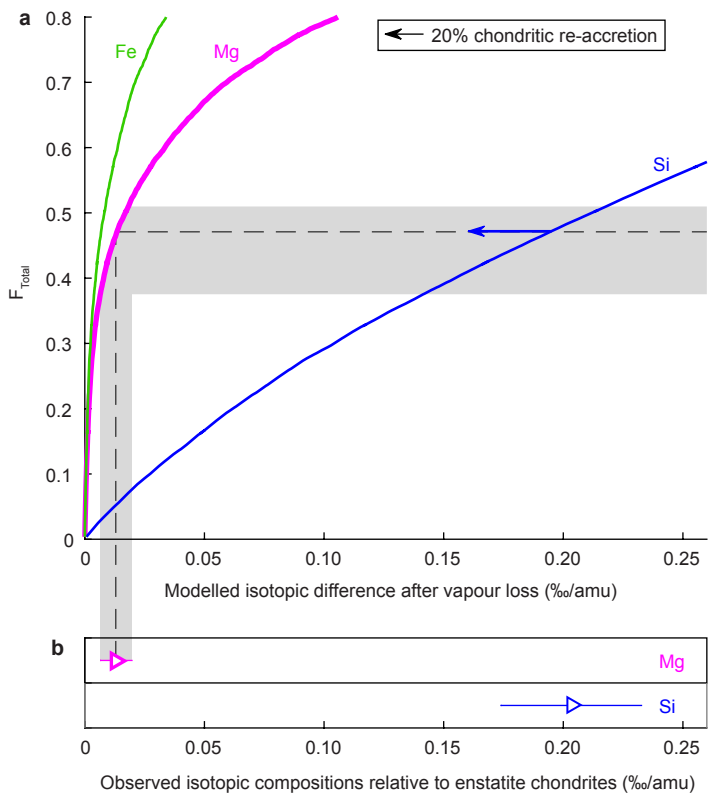
² The two C11 chondrites (Ivuna, Orgueil) and one CR2 (Al Rais) were excluded from the mean.

³ The two Ocean Island Basalts (BHVO-2 and BIR-1) were excluded from the mean.

Sp Hz = Spinel Harzburgite, Gt Lz = Garnet Lherzolite, Plg = Plagioclase, MORB = Mid-Ocean Ridge Basalt, OIB = Ocean Island basalt, Bas Sherg = Basaltic Shergottite, Ol = Olivine-phyric.







<i>Meteorites</i>	Source	Sample weight digested (mg)	Identification number
Orgueil	NMH	98.14	BM 36104
Ivuna	NHM	11.6	BM 1991, M5
Al Rais	NHM	4.6	BM 1971, 289
Renazzo	NHM	4.4	BM 41105
Kainsaz	NMH	69.34	BM 1988,M24
Murchison	NMH	86.92	BM 1970,6
Parnallee	NMH	6.37	BM 34792
Dhurmsala	NHM	12.69	BM 96262
Ceniceros	NMH	46.93	BM 1989,M31
Zag	Meteorite dealer	7.18 ¹	-
Khairpur	NMH	69.39	BM 51366
Hvittis	NHM	45.8	BM 86754
Yilmia	NHM	29.8	BM 1972, 132
St Mark's	NMH	60.37	BM 1990, 339
Abee	NHM	66.7	BM 1997, M7
Indarch	NHM	97.0	BM 1921, 23
ALHA 77005	NASA	71.99	ALHA77 005, 234
DaG 476	Meteorite dealer	91.78	-
EETA 79001	NASA	121.19	EETA79 001, 665
RBT 04262	NASA	89.09	RBT 04 262, 61
Zagami	Meteorite dealer	61.00	-
LAR 06319	NASA	95.58	LAR 06 319, 59
Johnstown	NHM	133.31	BM 1959,828
Shalka	NHM	155.58	BM 33761
Tatahouine	NHM	112.89	BM1931,490
d'Orbigny	Meteorite dealer	15.28 ¹	-
<i>Terrestrial</i>			Location
BZ116	ref 51	7.89 ¹	Bozu section, Horoman peridotite, Japan
Vi 313-102	ref 52	23.97 ¹	Vitim volcanic field, Russia
BZ143	ref 51	4.23 ¹	Bozu section, Horoman peridotite, Japan
BZ-29	ref 53	12.97 ¹	Zabargad Island, Egypt
Vi 314-56	ref 52	10.39 ¹	Vitim volcanic field, Russia
BZ251	ref 51	11.93 ¹	Bozu section, Horoman peridotite, Japan
2370-1	ref 54	3.39	9-10°N East Pacific Rise
DS12-29	ref 55	3.92	26°S Mid-Atlantic Ridge
9/30a(1)	ref 56	2.07	57°E South-West Indian Ridge

¹Aliquot from a homogenised powder of several grams

NHM = Natural History Museum

NASA = National Aeronautics and Space Administration

