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A Mercury-like component of early Earth yields U in the core and high mantle ^{142}Nd

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Recent ^{142}Nd isotope data indicate that the silicate Earth has an Sm/Nd ratio greater than the supposed chondritic building blocks of the planet. This elevated Sm/Nd has been ascribed either to a “hidden” reservoir in the Earth^{1,2} or to loss of an early-formed terrestrial crust by impact ablation³. Since removal of crust by ablation would also remove the heat producing elements, K, U and Th, the difficulty of balancing terrestrial heat production with heat flow would be severe³. The alternative, a “hidden” reservoir is generally assigned to the silicate lower mantle. We have previously shown, however, that the core is a likely reservoir for some lithophile elements such as Nb⁴. We therefore addressed the question of whether core formation could have fractionated Nd from Sm and also acted as a sink for heat producing elements. We show here that addition of a reduced Mercury (or enstatite-chondrite)-like body rich in sulfur to the early Earth would generate a superchondritic Sm/Nd in the mantle and $^{142}\text{Nd}/^{144}\text{Nd}$ anomaly of ~14ppm relative to chondrite. Additionally, the S-rich core would partition U strongly and Th slightly, providing a substantial part of the “missing” heat source for the geodynamo.

Terrestrial rocks were recently found to have higher ratios of radiogenic ^{142}Nd to nonradiogenic ^{144}Nd than the chondritic meteorites generally supposed to be representative of the material from which Earth accreted^{1,2}. ^{142}Nd was produced during the early history of the solar system from decay of the extinct radionuclide ^{146}Sm ($t_{1/2}=68 \text{ Ma}^5$) and the presence of a positive ^{142}Nd anomaly of ~20 ppm

$$\left(= \frac{10^6 \left[\left(\frac{^{142}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{earth}} - \left(\frac{^{142}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{chondrite}} \right]}{\left(\frac{^{142}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{earth}}} \right) \text{ or } \sim 9 \text{ ppm}^6$$
 in silicate Earth would require an Sm/Nd ratio higher than chondritic^{1,2}. This high Sm/Nd ratio was established early in Earth history while ^{146}Sm was “alive”. A plausible mechanism for generating high Sm/Nd in Earth’s mantle is partial melting and melt extraction to form a crust. Because Nd is less compatible in mantle silicates than Sm⁷ partial melts have relatively low Sm/Nd and the solid residue high Sm/Nd. A low Sm/Nd crust could be completely removed from the mantle

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system by subduction to an inaccessible region of the deep mantle¹ or removed from Earth by impact ablation³. The problem with the former hypothesis is the lack of evidence for a hidden silicate reservoir, while the latter suffers from the requirement that much of Earth's heat production, in the form of radioactive U, Th and K would be removed together with the low Sm/Nd crust. Assuming chondritic abundances of U and Th and a K/U ratio of ~12000 for silicate Earth⁸ the heat production in the Earth is only about 0.6 times current heat loss⁹. Reducing the heat sources further by ablation loss would make it even more difficult to reconcile production with heat loss. An additional question in the context of heat production is the energy source for the Earth's magnetic field¹⁰. Arising from convection in the core, Earth has had a magnetic field for at least 3.5 Ga. The crystallisation of the inner core is an important source of energy for the geodynamo¹¹ but most attempts to construct histories of core cooling indicate that the inner core cannot be much older than 1-1.5Ga^{10,11} unless a source of radioactive heating is present. Numerous studies have focussed on ⁴⁰K as a potential core heat source since K, in common with all moderately volatile elements⁸ is depleted in silicate Earth relative to chondritic abundance. Furthermore, high pressure experiments^{12,13} indicate that K enters sulfide under oxidising conditions and S is believed to be a major component of the core's complement of ~10% of elements of low atomic number¹⁴. It appears, however that the maximum possible K content of the core is insufficient to generate more than a small fraction of the 2-5 TW required to generate reasonable core thermal histories^{11,13}. The alternative explanation, that U and/or Th provide the energy for core convection has some support from early experiments on sulfide-silicate partitioning¹⁵ but more recent results indicate very little partitioning of U into S-bearing metals even under extreme conditions¹⁶.

We approached the problem of U, Th, Nd and Sm in Earth's mantle and core from the standpoint of recent work on partitioning between sulfide melts and silicate melts¹⁷. Kiseeva and Wood (2013) found that the sulfide-silicate partition coefficient for any element *i*,

$D_i = \frac{[i]_{sulf}}{[i]_{sil}}$ is dependent on the FeO content of the silicate melt such that for FeS-rich sulfides:

$$\log D_i = A - n \log [FeO] \quad (1)$$

where *A* is a constant and *n* is a constant dependent on the valency of element *i*. Therefore, under strongly reducing conditions, where the FeO content of the silicate melt is very low (<1% for example) one would expect *D_i* values to be much higher than under the conditions of MORB crystallisation where the FeO content of the melt is about 8-10%. This hypothesis is consistent with the data of Murrell and Burnett¹⁵ who observed strong partitioning of U into sulfide liquid at low *f_{O2}*. Given terrestrial accretion models calling for prolonged periods of growth under reduced conditions^{18,19}, the demonstration that Mercury is a highly-reduced sulfur-rich planet with a liquid core^{20,21} and the association of rare earth elements, U and Th with sulphides in enstatite chondrite meteorites²², we investigated partitioning of U, Th, Sm, Nd and several other lithophile elements into liquid iron sulfide under reducing conditions.

Experiments were performed at 1.5 GPa and temperatures between 1400 and 1650°C using starting materials which were approximately 50:50 mixtures of silicate and FeS doped with a range of lithophile trace elements including U, Th, La, Nd, Sm, Eu, Yb, Ce, and Zr (see Methods). The silicate was a basalt-like composition in the system CaO-MgO-Al₂O₃-SiO₂ with variable FeO. Analysis was by electron microprobe and Laser Ablation ICP-MS (Methods). Table 1 presents a summary of sulfide-silicate partitioning results (see Extended Data Section for complete analyses).

Fig 1a shows data from a series of experiments performed at 1400°C. As can be seen, the partition coefficients of U, Nd and Sm are strong functions of the FeO content of the silicate melt, increasing dramatically, as predicted, as the FeO content decreases below 1wt%. The negative slope of $\log D_i$ as $f(\log[\text{FeO}]_{\text{sil}})$ reverses at high FeO_{sil} , however, because the sulfide dissolves progressively more oxygen as the FeO content of silicate increases and these lithophile elements follow oxygen into the sulfide. We found similar behaviour in two more series of experiments at higher temperature (Table 1). Other lithophile elements, notably Ti, Nb and Ta (BJW, unpublished data) behave similarly. Importantly we find $D_{\text{U}} > D_{\text{Nd}} > D_{\text{Sm}}$ for partitioning into sulfide in all experiments. At very low FeO contents all D_i become >1 (Fig 1a). Furthermore (Fig 1b) D_{Nd} is always significantly greater than D_{Sm} with $D_{\text{Nd}}/D_{\text{Sm}}$ approaching 1.5 in some cases.

The implications of Fig 1 are that segregation of sulfide (or S-rich metal) from reduced FeO-poor silicate will lead to enrichment of the metallic phase in U and in Nd relative to Sm when compared to the silicate. Addition of such material to the core and mantle respectively of a growing planet would provide a core heat source and a mantle with superchondritic Sm/Nd, Yb/Sm and Yb/La. Although potentially detectable in terms of a mantle ¹⁴²Nd anomaly, the fractionation of heavy from light REE (Table 1) in the primitive mantle of the body (bulk silicate earth in this case) would have little effect on its overall REE pattern (Extended Figure 3). Similarly, there would be no significant Eu anomaly despite the fact that Eu is probably in the 2⁺ oxidation state (unlike the other 3⁺ lanthanides) under these conditions (Extended Figure 3). If such a body represented Earth early in its history then the mantle would have a positive ¹⁴²Nd anomaly relative to chondrite (as observed) and much of the energy deficit identified for core convection¹⁰ would be supplied by U (and Th). We find that $D_{\text{Th}}/D_{\text{U}}$ is about 0.1, indicating that U would be accompanied by Th in the S-rich core. Addition of more oxidised material later in accretion would lead to the higher current FeO content of the mantle (8.1%)⁸, but could not erase the superchondritic Sm/Nd ratio of the mantle and U content of the core unless there were complete core-mantle re-equilibration.

Figure 2 illustrates the impact of adding a highly reduced body rich in sulfide to the growing Earth. The Th/U ratio of silicate Earth would be higher than chondritic (3.8-3.9^{8,23}) which provides an important constraint on how much U can be present in Earth's core. Based on the Pb-isotopic compositions of Archean galenas²⁴ and of 3.5 Ga komatiites²⁵ the Th/U ratio of the Archean mantle has been estimated to be 4.3. Tatsumoto²⁶ argued, based on the Pb isotopic compositions of basalts, for an early differentiation of the mantle which resulted in a Th/U of 4.2-4.5 in the mantle source regions. Since that time the Th/U ratio of

the mantle has decreased probably due to preferential recycling of the more soluble uranium²⁷.

Figure 2 shows 4 models of U content of the core and ¹⁴²Nd anomaly of the mantle (relative to bulk Earth) based on our partitioning data. We choose a reduced body of 0.15 mass fraction sulfide corresponding to the S content of primitive CI chondrites⁸ and use values of D_{Sm} (Sm_{sulf}/Sm_{sil}) close to the observed maximum of 0.8-2.2 noting that D_{Sm} values increase with decreasing temperature and that segregation of sulfide from a crystal-melt mush instead of melt alone would increase them further because of the incompatibility of Sm, Nd, U in crystals. As can be seen (Fig 2a), adding 20% of such a body to Earth would lead to 4-5 ppb of U in the core, Th/U of silicate Earth of 4.17 and a ¹⁴²Nd anomaly in the mantle relative to bulk Earth of ~7ppm. Increasing the reduced body mass to 45% (Fig 2b) leads to about 8 ppb U in the core, Th/U of silicate Earth of 4.5 and mantle ¹⁴²Nd anomaly of 13.9 ppm relative to bulk Earth. We have performed a sensitivity analysis (Extended Figure 2) and find that, if Th/U of silicate Earth is 4.5, the maximum U content of the core is 8 ppb with a Th content of ~8 ppb. These figures increase to ~10 ppb if Th/U of silicate Earth is 4.7. The ¹⁴²Nd anomaly is 13.9 ppm in the former case and ~17 ppm in the latter. The estimated U and Th contents of the core would lead to 2-2.4 TW, sufficient to power the geodynamo¹¹ even without the potential 0.4-0.8TW from ⁴⁰K decay¹³. We can reduce the size of the reduced body by increasing its sulphur content (Fig 2c,d) and increase D_U/D_{Sm} but the overall effects on the core and mantle ¹⁴²Nd remain close to those summarised above if Th/U of BSE is constrained to be 4.5 or 4.7. Note that the scenarios shown in Fig 2 refer to a terrestrial core containing between 3.2 and 8.1 wt% S. The concentration of cosmochemically abundant volatile S in the core is unknown, but recent suggestions range from a cosmochemical estimate of 1.7 wt%¹⁴ to ~6 wt%²⁸ from liquid metal density measurements and 14.7 wt%²⁹ from high temperature-high pressure equation of state measurements. The range shown in Fig 2 is, therefore appropriate for the current state of knowledge.

We conclude that a period of growth of the accreting Earth under reduced, S-rich conditions would generate a significant (~14ppm) ¹⁴²Nd anomaly in silicate Earth, in agreement with observations and add sufficient U and Th to the core to generate 2-2.4 TW required to drive the geodynamo.

Methods

Experimental Methods

Starting materials for high pressure experiments consisted of mixtures of ~50wt% (Fe,Ni)S and ~50% of a synthetic silicate approximating the 1.5 GPa eutectic composition in the Anorthite-Diopside-Forsterite system³⁰. The sulfide component was analytical grade FeS doped with 1-3% NiS. Trace elements were added as a mix consisting of Zr, La, Ce, Nd, Sm, Eu, Yb, Th and U as oxides. After adding the trace element mix such that each element was present at 1000-2000 ppm, the silicate and sulfide starting materials were mixed in 50:50 proportions and ground under acetone for 20 minutes before being dried at 110°C prior to the experiment. Starting compositions were loaded into 3 mm O.D. and 1mm I.D. graphite capsules.

Experiments were conducted in a ½-inch diameter piston-cylinder apparatus using external cylinders either of BaCO₃-Silica glass (at 1500 and 1650 °C) or CaF₂ (at 1400°C) and an 8 mm O.D. graphite furnace with 1 mm wall. The unsealed capsule was separated from the graphite furnace by an interior MgO sleeve, with a 0.5 mm thick alumina disk on top to prevent puncture by the thermocouple. Temperatures were controlled and monitored using a tungsten-rhenium thermocouple (W5%Re/W26%Re), and temperature maintained within $\pm 1^\circ\text{C}$. Experimental conditions were 1400°C, 1500°C and 1650°C at 1.5 GPa and with experiment durations between 1 and 4.5 hrs. These times are sufficient to approach equilibrium in small graphite capsules¹⁷. Experiments were quenched by turning off the power supply. After quench, the capsule was extracted from the furnace, mounted in Acrylic and polished for further analyses with electron microprobe and Laser Ablation ICPMS. All experimental charges contained sulfide blebs embedded in a silicate glass matrix.

Microanalysis

Samples were analyzed on the JEOL 8600 electron microprobe in the Archaeology Department at the University of Oxford. WDS analyses of major element compositions of silicate glasses and sulfides were performed at 15 kV with a beam current of 20 nA and a 10µm defocused beam (Extended Data Table 1,2). At least 20 analyzes were taken of the silicate and sulfide in each experiment. Count times for major elements (Si, Al, Ca, Mg, Fe in silicate, Fe in sulfide) were 30 seconds on the peak and 15 seconds background. Minor elements (S, Ni, O) were analyzed for 60 seconds peak and 30 seconds background. We have previously noted Ni loss from similar experiments¹⁷ and the principal reason for adding Ni was to provide an additional check on Laser Ablation ICP-MS analyses of the trace elements of interest (see below). A range of natural and synthetic standards was used for calibration. Standards for silicate were wollastonite (Si, Ca), jadeite (Al), periclase (Mg) and Hematite (Fe). Standards for sulfides were Ni metal (Ni), galena (S) and Hematite (Fe, O). Oxygen in the sulfides was determined using the K α peak and LDE crystal.

We determined Uranium and Samarium contents of 3 product sulfides as a further check on the LA-ICP-MS analyses. In this case we measured the M α peak for U and L α peak for Sm using standards of UO₂ and SmPO₄ respectively and a PET crystal. Operating conditions were 15 kV, 40 nA, and a 10µm beam. Count time for Uranium was 120 seconds on peak and 60 seconds background. Samarium was analyzed for 150 seconds on the peak and 75 seconds background.

Trace elements in silicates and sulfides were measured by Laser Ablation ICPMS employing a NexION 300 quadrupole mass spectrometer coupled to a New Wave Research UP213 Nd:YAG laser at the University of Oxford. A laser repetition rate of 10Hz and 25-50 µm spot size were used for silicate glasses and sulphides (Extended Data Table 3,4) with an energy density of $\sim 12 \text{ J.cm}^{-2}$. Operating in time-resolved mode, we employed 20 seconds of background acquisition, followed by ablation for 60 seconds. Between analyses we employed a 60-90 second “wash-out” time. The following masses were counted: ²⁴Mg, ²⁷Al, ²⁹Si, ⁵⁷Fe, ⁶⁰Ni, ⁴³Ca, ⁹⁰Zr, ¹³⁹La, ¹⁴⁰Ce, ¹⁴²Nd, ¹⁵²Sm, ¹⁵³Eu, ¹⁷⁴Yb, ²³²Th, ²³⁹U. Our external standard was NIST610 glass and we typically collected 3 spectra of this at the beginning and end of each sequence of 10-15 unknowns. BCR-2G standard

was used as a secondary standard to check the accuracy of the calibration. Ablation yields were corrected by referencing to the known concentrations of Si and Ca (silicate glass) and Fe (sulfides), which had been determined by microprobe. Data reduction was performed off-line using the Glitter 4.4.3 software package, which enabled us to identify occasional sulfide inclusions in the silicate analyses. Since the Fe content of the NIST610 standard is only 460 ppm, the background is high and the matrices are very different, cross-checks on the sulfide analyses were required. Therefore Ni was measured with the electron microprobe and LA-ICPMS. In agreement with Kiseeva and Wood¹⁷, we observed no systematic offset between electron microprobe and LA-ICP-MS analyses for Ni (Extended data Tables 2,4). Additionally, as discussed above, the U and Sm contents of the sulfides were measured by electron microprobe in three experimental charges (1415, 464, 477). 20-43 electron probe analyses were performed on each sample. The highest U and Sm concentrations were measured in experiment 1415 with Laser ICPMS (U=2958 ppm, Sm=719 ppm). Comparative measurements with electron probe yielded values of U=3280±490 ppm and Sm=707±110 ppm, (uncertainty=2 standard errors) and therefore show excellent agreement. Two samples with lower U and Sm concentrations were also analysed. ICPMS measurements for 464 yielded U=952ppm and Sm=327ppm, while sample 477 gave U=927 ppm and Sm=300ppm. Electron microprobe concentrations of U=1164±224ppm and Sm=319±87 ppm (464) and U=991±69ppm and Sm=277±38 ppm (477) are also in excellent agreement with the LA-ICP-MS measurements. We conclude that our LA-ICP-MS results have no detectable systematic offset due to matrix effects or calibration errors.

Extended Data

Extended Fig 1. Partition coefficients for U, Nd and Sm with changing log FeO content in silicate melt (wt%)

(a) represents results for D values of experiments performed 1.5 GPa and 1500°C. (b) D value results at 1.5GPa and 1650°C .

Extended Fig 2. Core content of U (ppb) and mantle ^{142}Nd anomaly (ppm) at $D_U/D_{Sm} = 3$
Figure illustrates the effect on the Nd and U content using the same parameters (D_{Nd}/D_{Sm} at 1.4 and D_{Th}/D_U at 0.1) as in figure 2 but with a higher D_U/D_{Sm} ratio. (a), (b) show the calculated effect of adding to Earth 20% M_E or 45% M_E of reduced body containing 0.15 mass fraction sulphide. (c) and (d) illustrate the same scenario except the reduced body contains 0.22 mass fraction sulfide.

Extended Fig 3. Rare earth element fractionation at 3.2% and 8.1% sulfur in the core

This figure shows the calculated rare earth element pattern in Bulk Silicate Earth (BSE) for the two extreme cases of Fig 1a (3.2% S) and Extended Fig 2d (8.1% S). Black diamonds represent REE concentrations relative to chondritic abundances and normalized to Yb = 1, at 3.2% sulfur in the core (20% reduced mass impactor and 0.15 mass fraction sulfide). White diamonds illustrate the REE fractionation at elevated sulfur content (8.1% sulfur in the core, 35% reduced mass impactor and 0.22 mass fraction sulfide). We assumed $D_{\text{Sm}}(\text{sulfide/silicate}) = 1$ and D_i/D_{Sm} ratios for other elements from experiment 464. Both scenarios result in very small depletions of light REE relative to heavy REE in BSE. The trend is broadly consistent with that seen in the depleted MORB-mantle composition (blue diamonds) but much smaller. The effect on the REE pattern of BSE would, as can be seen, be undetectable. Blue diamonds illustrate the measured ratio of depleted MORB mantle (Salters and Stracke, 2004) to bulk silicate earth (Palme and O'Neill, 2003), assuming chondritic abundances of refractory lithophile elements in the latter. Error bars are from propagated error calculation and correspond to 1 standard deviation.

Extended Table 1
Major element composition of silicate glass

P = 1.5 GPa, T = 1400°C										
Run	421		428		427		426		429	
n	25	σ	20	σ	23	σ	19	σ	31	σ
SiO ₂	56.19	0.28	57.69	0.25	59.63	0.80	57.93	0.01	48.14	0.26
Al ₂ O ₃	20.50	0.14	20.40	0.11	20.21	0.11	19.60	0.09	18.17	0.16

P = 1.5 GPa, T = 1400°C										
Run	421		428		427		426		429	
CaO	11.48	0.08	13.02	0.09	10.97	0.07	10.48	0.16	9.39	0.08
MgO	6.66	0.06	6.53	0.06	6.50	0.11	6.22	0.08	5.61	0.05
FeO	3.15	0.07	1.21	0.04	0.56	0.03	0.50	0.23	16.36	0.13
S	0.11	0.01	0.18	0.01	0.34	0.02	6.11	0.24	0.24	0.02
Σ trace	0.81	0.02	0.90	0.02	0.78	0.04	0.30	0.03	0.65	0.05
Total	98.90		99.93		98.99		101.14		98.56	

P = 1.5 GPa, T = 1650°C									
Run	461		462		477		464		
n	43	σ	40	σ	31	σ	27	σ	
SiO ₂	53.28	0.84	58.38	0.22	58.51	0.80	58.60	0.31	
Al ₂ O ₃	21.64	0.62	20.32	0.12	19.00	0.26	19.17	0.23	
CaO	12.02	0.08	10.78	0.09	9.89	0.12	9.15	0.08	
MgO	8.13	0.29	7.01	0.13	5.99	0.10	6.20	0.06	
FeO	1.98	0.11	0.61	0.09	0.51	0.20	0.48	0.08	
S	0.42	0.04	1.39	0.08	5.93	0.12	7.18	0.12	
Σ trace	0.79	0.03	0.75	0.01	0.44	0.01	0.40	0.01	
Total	98.26		99.24		100.27		101.18		

P = 1.5 GPa, T = 1500°C									
Run	1414		1415		1416		1417		
n	26	σ	22	σ	26	σ	31	σ	
SiO ₂	48.86	0.34	50.66	0.34	41.12	0.43	39.28	0.36	
Al ₂ O ₃	16.58	0.16	13.87	0.16	15.95	0.16	15.00	0.17	
CaO	15.14	0.09	13.16	0.07	14.91	0.14	14.12	0.16	
MgO	15.99	0.12	14.11	0.07	15.51	0.12	15.05	0.17	
FeO	0.61	0.11	0.41	0.06	7.51	0.28	11.59	0.5	
S	0.79	0.03	6.98	0.08	0.25	0.13	0.24	0.24	
Σ trace	2.01	0.04	1.55	0.04	2.28	0.03	2.16	0.10	
Total	99.98		100.74		97.53		97.44		

Values in wt%; σ = calculated from error propagation; n = number of measurements; Σ trace= Sum of trace elements measured with LA-ICPMS

Extended Table 2
Major element composition of sulfides

P = 1.5 GPa, T = 1400°C										
Run	421		428		427		426		429	
n	21	σ	27	σ	23	σ	25	σ	23	σ
O	0.92	0.78	0.36	0.05	0.21	0.04	0.20	0.15	3.06	0.21
S	36.85	1.26	36.00	0.39	36.86	0.33	35.63	1.46	33.24	0.25
Fe	60.97	0.37	62.15	0.52	62.34	0.37	62.51	1.08	62.52	0.18
Ni	0.23	0.43	0.35	0.06	0.29	0.05	0.36	0.10	0.32	0.02

P = 1.5 GPa, T = 1400°C										
Run	421		428		427		426		429	
Sm	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.
U	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.
Σ trace	0.005	0.0005	0.008	0.001	0.040	0.004	0.556	0.139	0.004	0.002
Total	98.98		98.77		99.74		99.26		99.14	

P = 1.5 GPa, T = 1650°C									
Run	461		462		477		464		
n	22	σ	25	σ	21	σ	22	σ	
O	0.37	0.20	0.21	0.15	0.21	0.12	0.16	0.08	
S	36.68	0.02	35.55	0.78	32.92	1.40	33.82	0.53	
Fe	61.66	0.32	62.90	0.74	64.45	1.29	62.94	0.64	
Ni	0.37	0.33	0.33	0.09	0.25	0.07	0.14	0.02	
Sm	n.m.	n.m.	n.m.	n.m.	0.028	0.004	0.032	0.009	
U	n.m.	n.m.	n.m.	n.m.	0.099	0.007	0.116	0.022	
Σ trace	0.019	0.002	0.105	0.009	0.225	0.016	0.206	0.073	
Total	99.11		99.09		98.18		97.41		

P = 1.5 GPa, T = 1500°C									
Run	1414		1415		1416		1417		
n	26	σ	40	σ	35	σ	35	σ	
O	0.24	0.18	0.51	0.32	1.98	0.77	3.18	0.83	
S	33.89	2.63	32.18	2.06	35.46	0.79	33.40	0.85	
Fe	63.37	2.10	64.05	2.08	60.50	0.29	61.26	1.37	
Ni	1.09	0.51	0.85	0.25	1.16	0.03	1.13	0.05	
Sm	n.m.	n.m.	0.071	0.011	n.m.	n.m.	n.m.	n.m.	
U	n.m.	n.m.	0.328	0.049	n.m.	n.m.	n.m.	n.m.	
Σ trace	0.100	0.011	0.380	0.048	0.014	0.002	0.015	0.00	
Total	98.69		98.37		99.11		98.99		

Values in wt%; σ = calculated from error propagation; n = numbers of measurements; n.m. = not measured

Extended Table 3
Trace element concentration in silicates

P = 1.5 GPa, T = 1400°C										
Run	421		428		427		426		429	
n	5	σ	5	σ	5	σ	5	σ	5	σ
Zr90	1226	29	1153	43	1171	49	206	55	956	72
La139	641	21	574	8	607	31	247	25	511	39
Ce140	4.0	0.1	4	0.1	3.7	0.4	2	0	3.4	0.4
Nd142	494	15	450	6	474	27	167	25	397	30
Sm152	796	25	728	9	775	39	304	30	645	50
Eu153	709	13	670	11	636	37	452	24	578	47

P = 1.5 GPa, T = 1400°C										
Run	421		428		427		426		429	
Yb174	1002	31	903	6	1009	42	590	25	799	57
Th232	1030	21	967	14	984	51	465	37	813	57
U238	961	21	929	22	893	43	117	4	782	52
Ni60	n.m.	n.m.	n.m.	n.m.	34	12	108	3	7	1

P = 1.5 GPa, T = 1650°C									
Run	461		462		477		464		
n	5	σ	5	σ	5	σ	5	σ	
Zr90	1174	26	1191	19	413	4	311	8	
La139	632	24	568	2	387	8	367	11	
Ce140	4.7	0.4	3.8	0.2	3	0.1	2.5	0.1	
Nd142	489	18	441	6	278	6	325	7	
Sm152	780	30	715	7	477	7	435	6	
Eu153	661	31	624	7	557	6	499	18	
Yb174	985	29	1028	22	739	9	683	20	
Th232	980	24	1030	17	712	12	640	16	
U238	922	35	689	12	159	1	135	3	
Ni60	n.m.	n.m.	20	11	9	8	4	0	

P = 1.5 GPa, T = 1500°C									
Run	1414		1415		1416		1417		
n	5	σ	5	σ	5	σ	5	σ	
Zr90	3048	52	2287	53	3455	35	3133	38	
La139	1558	25	1190	29	1752	27	1746	20	
Ce140	3.0	0.2	2.0	0.2	3.1	0.1	3.0	0.2	
Nd142	1222	23	917	18	1393	18	1372	21	
Sm152	2062	40	1583	32	2290	24	2231	40	
Eu153	1642	31	1422	31	1985	28	1957	21	
Yb174	2466	54	2312	53	2838	27	2580	29	
Th232	2551	51	2454	59	2887	32	2636	51	
U238	2439	49	932	27	2688	38	2606	45	
Ni60	17	10	10	5	13	1	21	3	

values in ppm; n = number of measurements, n.m. = not measured.

Extended Table 4
Trace element concentration in sulfides

P = 1.5 GPa, T = 1400°C										
Run	421		428		427		426		429	
n	5	σ	5	σ	5	σ	5	σ	5	σ
Zr90	0.5	0.0	1.2	0.2	6	1	1228	58	2.2	1.0
La139	4.1	0.4	10	1	47	7	573	12	2.8	1.7

P = 1.5 GPa, T = 1400°C										
Run	421		428		427		426		429	
Ce140	0.1	0.1	0.1	0.03	0.4	0.1	4	0	0.0	0.0
Nd142	3.2	0.5	8	1	38	3	472	16	2.1	1.2
Sm152	3.6	0.5	9	1	48	4	683	13	3.2	2.1
Eu153	19	2	47	8	94	11	140	1	5.9	2.5
Yb174	0.7	0.3	1.8	0.1	8	1	218	12	2.7	2.9
Th232	0.4	0.1	0.7	0.1	3.1	0.3	339	6	3.0	2.8
U238	11	1	19	1	101	6	1792	94	14	3
Ni60	2837	64	3596	138	3320	88	3764	159	3153	51

P = 1.5 GPa, T = 1650°C									
Run	461		462		477		464		
n	5		σ	5	σ	5	σ	5	σ
Zr90	3.5		0.4	77	3	909	66	840	194
La139	17		1	80	7	201	17	151	119
Ce140	0.16		0.03	0.7	0.1	2	0.2	1.8	0.9
Nd142	14		1	76	7	192	14	275	28
Sm152	18		2	110	8	300	18	327	31
Eu153	62		3	109	11	131	12	78	58
Yb174	4.8		0.4	43	3	179	4	142	82
Th232	2.5		0.3	26	3	184	15	156	121
U238	41		4	380	36	927	43	952	74
Ni60	3643		90	3130	133	2573	181	1538	103

P = 1.5 GPa, T = 1500°C									
Run	1414		1415		1416		1417		
n	5	σ	5	σ	5	σ	5	σ	
Zr90	12	3	1639	223	2.8	0.4	9	4	
La139	106	8	564	65	16	2	14	1	
Ce140	0.2	0.1	1.3	0.2	0.04	0.01	0.06	0.02	
Nd142	77	5	491	42	12	1	11	1	
Sm152	100	7	719	42	14	1	15	2	
Eu153	471	51	569	63	59	5	35	3	
Yb174	16	2	225	54	3.9	0.2	8	5	
Th232	7	2	315	31	3.1	0.3	8	2	
U238	216	36	2958	196	44	2	53	4	
Ni60	10728	381	8480	386	9234	149	11066	72	

values in ppm; n = number of measurements.

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Fig 1. Sulfide-silicate partitioning data

(a) Partition coefficients $D_i \left(= \frac{[i]_{sulfide}}{[i]_{silicate}} \right)$ for U, Nd and Sm at 1.5 GPa/1400°C plotted versus the log of the FeO content of the silicate melt in wt%. (b) The ratio of D_{Nd}/D_{Sm} plotted versus $\log \text{FeO}$. Error bars in both cases are ± 2 std errors and, if absent, are smaller than symbol size.

Fig 2. Core content of U (ppb) and mantle ^{142}Nd anomaly (ppm)

(a) Calculated effect of adding to Earth 20% M_E of reduced body containing 0.15 mass fraction sulfide. The sulfide is added to the core and the silicate to the mantle. Sulfide-silicate D_U/D_{Sm} fixed at 2, D_{Nd}/D_{Sm} at 1.4 and D_{Th}/D_U at 0.1 (Table 1). (b) Same as (a) except mass of reduced body is 45% of Earth mass. (c) and (d) as for (a) and (b) except reduced body contains 0.22 mass fraction sulfide. The reduced body and remainder of Earth each contain 14ppb U and 53.5 ppb Th consistent with chondritic abundances. Sulfide extraction was assumed to take place at the origin of the solar system.

Table1

Summary of sulfide/silicate partition coefficients

Run	P (GPa)	T (°C)	LogFeO _(sil) wt(%)	D _{Sm}	D _{Nd} /D _{Sm}	D _U /D _{Sm}	D _{Th} /D _U	D _{Eu} /D _{Sm}	D _{La} /D _{Sm}	D _{Yb} /D _{Sm}
421	1.5	1400	0.50	0.005	1.42	2.47	0.036	5.85	1.38	0.16
σ				0.001	0.29	0.37	0.008	0.93	0.23	0.07
428	1.5	1400	0.08	0.013	1.35	1.56	0.038	5.50	1.35	0.16
σ				0.001	0.19	0.17	0.005	1.06	0.20	0.02
427	1.5	1400	-0.25	0.062	1.30	1.81	0.028	2.36	1.23	0.13
σ				0.006	0.19	0.24	0.004	0.40	0.22	0.02
426	1.5	1400	-0.30	2.247	1.25	6.81	0.048	0.14	1.03	0.16
σ				0.333	0.30	1.43	0.010	0.05	0.15	0.09
429	1.5	1400	1.21	0.005	1.04	3.68	0.200	2.04	1.10	0.67
σ				0.0001	0.12	0.89	0.041	0.42	0.47	0.19
461	1.5	1650	0.30	0.023	1.22	1.92	0.058	4.09	1.18	0.21
σ				0.003	0.20	0.32	0.009	0.58	0.18	0.03
462	1.5	1650	-0.21	0.154	1.12	3.58	0.046	1.13	0.92	0.27
σ				0.011	0.13	0.43	0.007	0.14	0.10	0.03
477	1.5	1650	-0.29	0.629	1.10	9.26	0.044	0.37	0.83	0.39
σ				0.038	0.11	0.73	0.043	0.38	0.80	0.39
464	1.5	1650	-0.32	0.751	1.13	9.41	0.035	0.21	0.55	0.28
σ				0.073	0.16	1.18	0.020	0.16	0.44	0.16
1414	1.5	1500	-0.21	0.048	1.31	1.84	0.031	5.95	1.41	0.13
σ				0.004	0.14	0.34	0.009	0.80	0.16	0.02
1415	1.5	1500	-0.39	0.454	1.18	6.99	0.040	0.88	1.04	0.21
σ				0.028	0.13	0.66	0.005	0.11	0.14	0.05
1416	1.5	1500	0.88	0.006	1.39	2.70	0.067	4.92	1.52	0.23
σ				0.0005	0.18	0.25	0.007	0.56	0.19	0.02
1417	1.5	1500	1.06	0.007	1.20	3.11	0.150	2.71	1.19	0.50
σ				0.001	0.24	0.52	0.042	0.48	0.21	0.30

Partition coefficients in weight ratio. σ = calculated from error propagation. n = number of measurements.