A Mercury–like component of early Earth yields uranium in the core and high mantle $^{142}$Nd

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Recent $^{142}$Nd isotope data indicate that the silicate Earth (its crust plus the mantle) has a samarium to neodymium elemental ratio (Sm/Nd) that is greater than that of 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 ablation$^3$. Since removal of crust by ablation would also remove the heat-producing elements—potassium, uranium and thorium—such removal would make it extremely difficult to balance terrestrial heat production with the observed heat flow$^3$. In the ‘hidden’ reservoir alternative, a complementary low-Sm/Nd layer is usually considered to reside unobserved in the silicate lower mantle. We have previously shown, however, that the core is a likely reservoir for some lithophile elements such as niobium$^4$. We therefore address 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-like body (or, alternatively, an enstatite-chondrite-like body) rich in sulfur to the early Earth would generate a superchondritic Sm/Nd in the mantle and an $^{142}$Nd/$^{144}$Nd anomaly of approximately +14 parts per million relative to chondrite. In addition, the sulfur-rich core would partition uranium strongly and thorium slightly, supplying 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 do 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 (half-life, $t_{1/2}$ = 68 million years) and the presence of a positive $^{142}$Nd anomaly of ~20 parts per million (p.p.m.) (p.p.m.) calculated as

$$10^6 \left( \frac{^{142}Nd_{\text{Earth}}}{^{144}Nd_{\text{Earth}}} - \frac{^{142}Nd_{\text{chondrite}}}{^{144}Nd_{\text{chondrite}}} \right) / \left( \frac{^{142}Nd_{\text{Earth}}}{^{144}Nd_{\text{Earth}}} \right) \times 9 \text{ p.p.m.}$$

in the silicate Earth would require an Sm/Nd ratio higher than chondritic$^{1,2}$. This high Sm/Nd ratio was established early in Earth’s history while $^{146}$Sm was still ‘alive’ (that is, undergoing radioactive decay).

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 Sm$^5$, partial melts have relatively low Sm/Nd and the solid residue has high Sm/Nd. A low-Sm/Nd crust could be completely removed from the mantle system by subduction to an inaccessible region of the deep mantle$^1$ or removed from Earth by impact ablation$^2$. The problem with the former hypothesis is the lack of evidence for a hidden silicate reservoir, while the latter hypothesis suffers from the requirement that much of Earth’s heat production, in the form of radioactive uranium (U), thorium (Th) and potassium (K) be removed together with the low-Sm/Nd crust. Assuming chondritic abundances of U and Th and a K/U ratio of ~12.000 for the silicate Earth$^6$, the heat production in the Earth is only about 0.6 times the current heat loss$^7$. Reducing the heat sources further by ablation loss would make it even more difficult to reconcile heat production with heat loss.

An additional question in the context of heat production is that of the energy source for the Earth’s magnetic field$^8$. Arising from convection in the core, Earth has had a magnetic field for at least 3.5 billion years. The crystallization of the inner core is an important source of energy for the geodynamo$^9$ but most attempts to construct histories of core cooling indicate that the inner core cannot be much older than 1–1.5 billion years$^{10,11}$ unless a source of radioactive heating is present. Numerous studies have focused on $^{40}$K as a potential core heat source, because K, in common with all moderately volatile elements$^8$, is depleted in the silicate Earth relative to the chondritic abundance.

Furthermore, high-pressure experiments$^{12,13}$ indicate that K enters sulfide under oxidizing conditions and sulfur (S) is believed to be a major component of the core’s complement of approximately 10% of elements of low atomic number$^{14}$. 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 sulphide–silicate partitioning$^{15}$ but more recent results indicate very little partitioning of U into S-bearing metals even under extreme conditions$^{16}$.

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 melts$^{17}$. Kiseeva and Wood$^17$ found that the sulfide–silicate partition coefficient for any element $i$, defined as $D_i = [I\text{_{sulf}}]/[I\text{_{sil}}]$, is dependent on the FeO content of the silicate melt, such that for FeS-rich sulfides:

$$\log D_i = A - n \log [\text{FeO}]$$

where $A$ is a constant and $n$ is a constant dependent on the valency of element $i$. Therefore, under strongly reducing conditions, 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 crystallization where the FeO content of the melt is about 8%–10%. This hypothesis is consistent with the data of Murrell and Burnett$^{15}$, who observed strong partitioning of U into sulfide liquid at low oxygen fugacity $f_{O2}$. Given terrestrial accretion models calling for prolonged periods of growth under reduced conditions$^{18,19}$, the demonstration that Mercury is a highly reduced S-rich planet with a liquid core$^{20,21}$ and the association of the rare earth elements (REEs), U and Th with sulfides in enstatite chondrite meteorites$^{22}$, 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 1,400 °C and 1,650 °C using starting materials that 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–Al$_2$O$_3$–SiO$_2$ with variable FeO. Analysis

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was by electron microprobe and laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) (Methods). Table 1 presents a summary of sulphide–silicate partitioning results (see Extended Data Figs 1–3 and Extended Data Tables 1–4 for complete analyses).

Figure 1a shows data from a series of experiments performed at 1,400 °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 1 wt%. The negative slope of logD for f(\text{log}[\text{FeO}_{\text{sil}}]) reverses at high FeO$_{\text{sil}}$, however, because the sulphide dissolves progressively more oxygen as the FeO content of silicate increases and these three lithophile elements (Fig. 1a) follow oxygen into the sulphide. We found similar behaviour in two more series of experiments at higher temperature (Table 1 and Extended Data Fig. 1). Other lithophile elements, notably Ti, Nb and Ta (B.J.W., unpublished data) behave similarly. Importantly, we find D$_{\text{Ti}} >$ D$_{\text{Nd}} >$ D$_{\text{Sm}}$ for partitioning into sulphide in all experiments. At very low FeO contents all D$_{i}$ become >1 (Fig. 1a). Furthermore (Fig. 1b) D$_{\text{Nd}}$ is always appreciably greater than D$_{\text{Sm}}$, with D$_{\text{Nd}}$/D$_{\text{Sm}}$ approaching 1.5 in some cases.

The implications of Fig. 1 are that segregation of sulphide (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 REEs (Table 1 in the primitive mantle of the body (the bulk silicate Earth in this case) would have little effect on its overall REE pattern (Extended Data Fig. 3). Similarly, there would be no observable Eu anomaly despite the fact that Eu is probably in the 2$^{+}$ oxidation state (unlike the other 3$^{+}$ lanthanides) under these conditions (Extended Data Fig. 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 convection$^{10}$ 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-oxidized material later in accretion would lead to the higher current FeO content of the mantle (8.1%)$^{6}$, 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 sulphide to the growing Earth. The Th/U ratio of the silicate Earth would be higher than chondritic (3.8–3.9$^{25}$), which provides an important constraint on how much U can be present in Earth’s core. Based on the Pb-isotopic compositions of Archean galenas$^{24}$ and of 3.5-billion-year-old komatitites$^{25}$ the Th/U ratio of the Archean mantle has been estimated to be ≥4.3. Tatsumoto$^{26}$ argued, on the basis of 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, possibly owing to preferential recycling of the more soluble U$^{27}$.

Figure 2 shows four models of U content of the core and the 142Nd anomaly of the mantle (relative to the bulk Earth), based on our partitioning data. We choose a reduced body of 0.15 mass fraction sulphide, corresponding to the S content of primitive CI chondrites$^{8}$ and use values of D$_{\text{sm}}$ (Sm$_{\text{sulf}}$/Sm$_{\text{sil}}$) that are close to the observed maximum of 0.8–2.2, noting that D$_{\text{Sm}}$ values increase with decreasing temperature and that segregation of sulphide from a crystal-melt mush instead of melt alone would increase them further because of the incompatibility of Sm, Nd and U in crystals. As can be seen (Fig. 2a), adding 20% of such a body to Earth would lead to 4–5 parts per billion (p.p.b.) of U in the core, a Th/U of the silicate Earth of 4.17 and a 142Nd anomaly in the mantle relative to the bulk Earth of ~7 p.p.m. Increasing the reduced body mass to 45% (Fig. 2b) leads to about 8 p.p.b. U in the core, a Th/U of the silicate Earth of 4.5 and a mantle 142Nd anomaly of 13.9 p.p.m. relative to the bulk Earth.

We performed a sensitivity analysis (Extended Data Fig. 2) and find that, if the Th/U of the silicate Earth is ≤4.5, the maximum U content of the core is 8 p.p.b. with a Th content of ~8 p.p.b. These figures increase to ~10 p.p.b. if the Th/U of the silicate Earth is ≤4.7. The 142Nd anomaly is 13.9 p.p.m. in the former case and ~17 p.p.m. in the latter. The estimated U and Th contents of the core would lead to 2–2.4 TW, sufficient to power the geodynamo$^{11}$ even without the potential 0.4–0.8 TW from 40K decay$^{13}$. We can reduce the size of the reduced body by increasing its S content (Fig. 2c, d) and increase D$_{\text{U}}$/D$_{\text{Sm}}$ but the overall effects on the core and mantle 142Nd remain
### Table 1 | Summary of sulfide–silicate partition coefficients

| Run | Pressure, P (GPa) | Temperature, T (°C) | $\log([\text{FeO}]_{\text{sil}} \text{ (wt%)}))$ | $D_{\text{Sm}}$ | $D_{\text{U}}/D_{\text{Sm}}$ | $D_{\text{Nd}}/D_{\text{Sm}}$ | $D_{\text{Th}}/D_{\text{U}}$ | $D_{\text{La}}/D_{\text{Sm}}$ | $D_{\text{Yb}}/D_{\text{Sm}}$ |
|-----|------------------|---------------------|---------------------------------|--------------|----------------|----------------|----------------|----------------|----------------|
| 421 | 1.5              | 1,400               | 0.50                            | 0.005        | 1.42           | 2.47           | 0.036          | 5.85           | 1.38           | 0.16           |
|      | $\sigma$         |                     |                                 | 0.001        | 0.29           | 0.37           | 0.008          | 0.93           | 0.23           | 0.07           |
| 428 | 1.5              | 1,400               | 0.08                            | 0.013        | 1.35           | 1.56           | 0.038          | 5.50           | 1.35           | 0.16           |
|      | $\sigma$         |                     |                                 | 0.001        | 0.19           | 0.17           | 0.005          | 1.06           | 0.20           | 0.02           |
| 427 | 1.5              | 1,400               | $-0.25$                         | 0.062        | 1.30           | 1.81           | 0.028          | 2.36           | 1.23           | 0.13           |
|      | $\sigma$         |                     |                                 | 0.006        | 0.19           | 0.24           | 0.004          | 0.40           | 0.22           | 0.02           |
| 426 | 1.5              | 1,400               | $-0.30$                         | 2.247        | 1.25           | 6.81           | 0.048          | 0.14           | 1.03           | 0.16           |
|      | $\sigma$         |                     |                                 | 0.333        | 0.30           | 1.43           | 0.010          | 0.05           | 0.15           | 0.09           |
| 429 | 1.5              | 1,400               | 1.21                            | 0.005        | 1.04           | 3.68           | 0.200          | 2.04           | 1.10           | 0.67           |
|      | $\sigma$         |                     |                                 | 0.0001       | 0.12           | 0.89           | 0.041          | 0.42           | 0.47           | 0.19           |
| 461 | 1.5              | 1,650               | 0.30                            | 0.023        | 1.22           | 1.92           | 0.058          | 4.09           | 1.18           | 0.21           |
|      | $\sigma$         |                     |                                 | 0.003        | 0.20           | 0.32           | 0.009          | 0.58           | 0.18           | 0.03           |
| 462 | 1.5              | 1,650               | $-0.21$                         | 0.154        | 1.12           | 3.58           | 0.046          | 1.13           | 0.92           | 0.27           |
|      | $\sigma$         |                     |                                 | 0.011        | 0.13           | 0.43           | 0.007          | 0.14           | 0.10           | 0.03           |
| 477 | 1.5              | 1,650               | $-0.29$                         | 0.629        | 1.10           | 9.26           | 0.044          | 0.37           | 0.83           | 0.39           |
|      | $\sigma$         |                     |                                 | 0.038        | 0.11           | 0.73           | 0.043          | 0.38           | 0.80           | 0.39           |
| 464 | 1.5              | 1,650               | $-0.32$                         | 0.751        | 1.13           | 9.41           | 0.035          | 0.21           | 0.55           | 0.28           |
|      | $\sigma$         |                     |                                 | 0.073        | 0.16           | 1.18           | 0.020          | 0.16           | 0.44           | 0.16           |
| 1,414| 1.5              | 1,500               | $-0.21$                         | 0.048        | 1.31           | 1.84           | 0.031          | 5.95           | 1.41           | 0.13           |
|      | $\sigma$         |                     |                                 | 0.004        | 0.14           | 0.34           | 0.009          | 0.80           | 0.16           | 0.02           |
| 1,415| 1.5              | 1,500               | $-0.39$                         | 0.454        | 1.18           | 6.99           | 0.040          | 0.88           | 1.04           | 0.21           |
|      | $\sigma$         |                     |                                 | 0.028        | 0.13           | 0.66           | 0.005          | 0.11           | 0.14           | 0.05           |
| 1,416| 1.5              | 1,500               | 0.88                            | 0.006        | 1.39           | 2.70           | 0.067          | 4.92           | 1.52           | 0.23           |
|      | $\sigma$         |                     |                                 | 0.0005       | 0.18           | 0.25           | 0.007          | 0.56           | 0.19           | 0.02           |
| 1,417| 1.5              | 1,500               | 1.06                            | 0.007        | 1.20           | 3.11           | 0.150          | 2.71           | 1.19           | 0.50           |
|      | $\sigma$         |                     |                                 | 0.001        | 0.24           | 0.52           | 0.042          | 0.48           | 0.21           | 0.30           |

Partition coefficients in weight ratio. $\sigma$ is calculated from error propagation.

Figure 2 | Core content of U (p.p.b.) and mantle $^{142}$Nd anomaly (p.p.m.). a, Calculated effect of adding to the growing Earth a reduced body of 20% of Earth’s mass containing 0.15 mass fraction sulphide. The sulphide is added to the core and the silicate to the mantle. Sulphide/silicate $D_{\text{U}}/D_{\text{Sm}}$ is fixed at 2, $D_{\text{Nd}}/D_{\text{Sm}}$ at 1.4 and $D_{\text{Th}}/D_{\text{U}}$ at 0.1 (Table 1). b, Same as a except the mass of reduced body is 45% of the Earth’s mass. c and d, As for a and b except the reduced body contains 0.22 mass fraction sulphide. The reduced body and remainder of Earth each contain 14 p.p.b. U and 53.5 p.p.b. Th, consistent with chondritic abundances. Sulphide extraction was assumed to take place at the origin of the Solar System.
close to those summarized above if the Th/U of the bulk silicate Earth is constrained to be \( \leq 4.5 \) or \( \leq 4.7 \).

We note that the scenarios shown in Fig. 2 refer to a terrestrial core containing between 3.2 wt% S 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% (ref. 14) to \( \sim 6 \) wt% (ref. 28) from liquid-metal density measurements and 14.7 wt% (ref. 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 measureable (\( \sim +14 \) p.p.m.) \(^{142}\)Nd anomaly in the silicate Earth, in agreement with observations. This would also add sufficient U and Th to the core to generate \( 2 \sim 2.4 \) TW of the energy required to drive the geodynamo.

**Online Content Methods, along with any additional Extended Data display items and Source Data, are available in the online version of the paper; references unique to these sections appear only in the online paper.**

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**Author Information** Reprints and permissions information is available at www.nature.com/reprints. The authors declare no competing financial interests. Readers are welcome to comment on the online version of the paper. Correspondence and requests for materials should be addressed to A.W. (anke.wohlers@earth.ox.ac.uk) or B.J.W. (berniew@earth.ox.ac.uk).
METHODS

Experimental methods. Starting materials for high-pressure experiments consisted of mixtures of ~50 wt% (Fe,Ni)S and ~50% of a synthetic silicate approximating the 1.5 GPa eutectic composition in the anorthite–diopside–forsterite system[30]. 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 1,000 p.p.m., the silicate and sulfide starting materials were mixed in 50:50 proportions and ground under acetone for 20 min, then dried at 110 ºC before the experiment. Starting compositions were loaded into 3 mm outer diameter and 1 mm inner diameter graphite capsules.

Experiments were conducted in a half-inch-diameter piston-cylinder apparatus using external cylinders either of BaCO3-silica glass (at 1,500 ºC and 1,650 ºC) or CaF2 (at 1,400 ºC) and an 8 mm outer diameter graphite furnace with a 1-mm-thick 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 the temperature was maintained within ±1 ºC. Experimental conditions were 1,400 ºC, 1,500 ºC and 1,650 ºC at 1.5 GPa and with experiment durations between 1 h and 4.5 h. These times are sufficient to approach equilibrium in small graphite capsules[17]. Experiments were quenched by turning off the power supply. After quenching, the capsule was extracted from the furnace, mounted in acrylic and polished for further analyses with electron microprobe and LA-ICP-MS. All experimental charges contained sulfide blebs embedded in a silicate glass matrix.

Microanalyses. Samples were analysed on the JEOL 6800 electron microprobe in the Archaeology Department at the University of Oxford. Wavelength dispersive analyses of the major-element compositions of silicate glasses and sulfides were performed at 15 kV with a beam current of 20 nA and a 10 ´ 10 µm defocused beam (Extended Data Tables 1 and 2). At least 20 analyses 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 s on the peak and 15 s background. Minor elements (S, Ni, O) were analysed for 60 s peak and 30 s background. We have previously noted Ni loss from similar experiments[17] and the principal reason for adding Ni was to provide an additional check on LA-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 haematite (Fe, O). Oxygen in the sulfides was determined using the Kr and LDE crystal.

We determined U and Sm contents of three product sulfides as a further check on the LA-ICP-MS analyses. In this case we measured the Mo peak for U and the Lo peak for Sm using standards of UO2 and SmPO4, respectively and a PET crystal. Operating conditions were 15 kV, 40 nA and a 10 µm beam. The count time for U was 120 s on peak and 60 s background. Sm was analysed for 150 s on the peak and 75 s background.

Trace elements in silicates and sulfides were measured by LA-ICP-MS 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 10 Hz and spot size of 25–50 µm were used for silicate glasses and sulfides (Extended Data Tables 3 and 4) with an energy density of ~12 J cm−2. Operating in time-resolved mode, we employed 20 s of background acquisition, followed by ablation for 60 s. Between analyses we employed a 60–90 s ‘wash-out’ time. The following masses were counted: 24Mg, 27Al, 28Si, 32S, 56Fe, 57Fe, 58Ni, 59Ni, 60Ni, 65Cu, 69Cu, 70Zn, 74Zn, 87Sr, 88Sr, 106Ru, 107Pd, 108Pd, 109Ag, 110Ag, 114Cd, 116Cd, 118Cd, 124Sn, 128Sn, 130Te, 134Te, 136Te, 142Nd, 150Nd, 164Dy, 166Dy, 172Yb, 174Yb, 176Yb, 176Yb, 177Yb, 178Yb, 179Yb, 180Yb. Our external standard was NIST610 glass and we typically collected three spectra of this at the beginning and end of each sequence of 10–15 unknowns. The 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 (http://www.glittergemoc.com/) which enabled us to identify occasional sulfide inclusions in the silicate analyses. Since the Fe content of the NIST610 standard is only 460 p.p.m., the background is high and the matrices are very different, so cross-checks on the sulfide analyses were required. Therefore, we measured Ni with the electron microprobe and LA-ICP-MS. In agreement with Kisska and Wood[37], we observed no systematic offset between electron microprobe and LA-ICP-MS analyses for Ni (Extended Data Tables 2 and 4). Additionally, as discussed above, the U and Sm contents of the sulfides were measured by electron microprobe in three experimental charges (numbers 1415, 464, and 477). Between 20 and 43 electron probe analyses were performed on each sample. The highest U and Sm concentrations were measured in experiment 1415 with LA-ICP-MS (U = 2,958 p.p.m., Sm = 719 p.p.m.). Comparative measurements with electron probe yielded values of U = 3,280 ± 490 p.p.m. and Sm = 707 ± 110 p.p.m. (uncertainty is 2 s.e.) and therefore show excellent agreement. Two samples with lower U and Sm concentrations were also analysed. LA-ICP-MS measurements for experiment 464 yielded U = 952 p.p.m. and Sm = 327 p.p.m., while experiment 477 gave U = 927 p.p.m. and Sm = 300 p.p.m. Electron microprobe concentrations of U = 1,164 ± 224 p.p.m. and Sm = 319 ± 87 p.p.m. (experiment 464) and U = 991 ± 69 p.p.m. and Sm = 277 ± 38 p.p.m. (experiment 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.

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Extended Data Table 1 | Major-element composition of silicate glass.

Values are in weight per cent. σ is calculated from error propagation. n is the number of measurements. Σ trace is the sum of trace elements measured using LA-ICP-MS.

| Run  | 421  | 428  | 427  | 426  | 429  |
|------|------|------|------|------|------|
| n    | 25   | 20   | 23   | 19   | 31   |
| SiO₂ | 56.19| 57.69| 59.63| 57.93| 0.01 |
| Al₂O₃| 20.50| 20.40| 20.21| 19.60| 0.09 |
| CaO  | 11.48| 13.02| 10.97| 10.48| 0.16 |
| MgO  | 6.66 | 6.53 | 6.29 | 6.22 | 0.08 |
| FeO  | 3.55 | 3.21 | 3.09 | 3.00 | 0.23 |
| S    | 0.11 | 0.18 | 0.34 | 0.22 | 0.02 |
| Σ trace | 0.81 | 0.90 | 0.78 | 0.84 | 0.30 |
| Total| 98.90| 99.93| 98.99| 99.14| 98.56|

| Run  | 461  | 462  | 477  | 464  |
|------|------|------|------|------|
| n    | 43   | 40   | 31   | 27   |
| SiO₂ | 53.28| 58.38| 58.51| 58.60| 0.31 |
| Al₂O₃| 21.64| 20.32| 19.00| 19.17| 0.23 |
| CaO  | 12.02| 10.78| 9.89 | 9.15 | 0.08 |
| MgO  | 8.12 | 7.01 | 5.99 | 6.20 | 0.06 |
| FeO  | 1.98 | 1.61 | 0.95 | 0.51 | 0.20 |
| S    | 0.42 | 0.39 | 0.93 | 0.40 | 0.08 |
| Σ trace | 0.79 | 0.75 | 0.44 | 0.40 | 0.01 |
| Total| 98.26| 99.24| 100.27| 101.18|

| Run  | 1414 | 1415 | 1416 | 1417 |
|------|------|------|------|------|
| n    | 26   | 22   | 26   | 31   |
| SiO₂ | 48.86| 50.66| 41.12| 39.28| 0.36 |
| Al₂O₃| 16.58| 13.87| 15.05| 15.08| 0.17 |
| CaO  | 15.14| 13.16| 14.01| 14.12| 0.16 |
| MgO  | 15.99| 14.11| 15.51| 15.05| 0.17 |
| FeO  | 0.61 | 0.41 | 0.66 | 0.75 | 0.38 |
| S    | 0.93 | 0.68 | 0.25 | 0.13 | 0.24 |
| Σ trace | 2.01 | 1.55 | 2.28 | 2.16 | 0.10 |
| Total| 99.98| 100.74| 97.53| 97.44|
Extended Data Table 2 | Major-element composition of sulfides.

| Run  | 421 | 428 | 427 | 426 | 429 |
|------|-----|-----|-----|-----|-----|
| P = 1.5 GPa, T = 1400°C |
| a | 21 | a | 27 | a | 23 | a | 25 | a | 23 | a |
| O | 0.92 | 0.8 | 0.26 | 0.95 | 0.21 | 0.04 | 0.2 | 0.15 | 0.16 | 0.21 |
| S | 26.85 | 0.12 | 0.02 | 0.36 | 0.01 | 0.33 | 0.36 | 0.33 | 0.35 | 0.4 |
| Fe | 62.97 | 0.37 | 0.5 | 0.25 | 0.24 | 0.37 | 0.62 | 0.51 | 0.18 | 0.18 |
| Ni | 0.23 | 0.43 | 0.35 | 0.6 | 0.29 | 0.05 | 0.26 | 0.10 | 0.22 | 0.02 |
| 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. |
| Σtrace | 0.005 | 0.0005 | 0.008 | 0.001 | 0.040 | 0.004 | 0.056 | 0.159 | 0.004 | 0.002 |
| Total | 98.98 | 98.87 | 99.74 | 99.25 | 99.14 |

| Run  | 461 | 462 | 477 | 464 |
|------|-----|-----|-----|-----|
| P = 1.5 GPa, T = 1650°C |
| a | 22 | a | 25 | a | 21 | a | 22 | a |
| O | 0.37 | 0.2 | 0.21 | 0.15 | 0.21 | 0.12 | 0.16 | 0.08 |
| S | 36.68 | 0.02 | 0.35 | 0.78 | 0.32 | 0.29 | 1.4 | 33.32 | 0.53 |
| Fe | 61.66 | 0.32 | 0.24 | 0.74 | 0.45 | 0.12 | 0.7 | 62.94 | 0.64 |
| Ni | 0.37 | 0.33 | 0.09 | 0.25 | 0.07 | 0.14 | 0.02 |
| Sm | n.m. | n.m. | n.m. | n.m. | 0.006 | 0.004 | 0.002 |
| U | n.m. | n.m. | n.m. | n.m. | 0.009 | 0.007 | 0.116 | 0.022 |
| Σtrace | 0.019 | 0.002 | 0.195 | 0.009 | 0.223 | 0.016 | 0.206 | 0.073 |
| Total | 99.11 | 99.09 | 98.18 | 97.41 |

| Run  | 1414 | 1415 | 1416 | 1417 |
|------|------|------|------|------|
| P = 1.5 GPa, T = 1500°C |
| a | 26 | a | 40 | a | 35 | a | 35 | a |
| O | 0.24 | 0.18 | 0.31 | 0.22 | 1.08 | 0.77 | 2.18 | 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 | 0.19 | 0.01 | 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. | 0.038 | 0.049 | n.m. | n.m. | n.m. | n.m. |
| Σtrace | 0.030 | 0.011 | 0.030 | 0.014 | 0.001 | 0.002 | 0.005 | 0.00 |
| Total | 96.89 | 96.37 | 95.11 | 95.99 |

Values are in weight per cent. σ calculated from error propagation. n.m., not measured. n is the number of measurements. Σtrace is the sum of trace elements measured using LA-ICP-MS.
Extended Data Table 3 | Trace-element concentration in silicates.

| P = 1.5 GPa, T = 1400°C |  |  |  |  |  |  |  |
|---|---|---|---|---|---|---|---|
| Run | 421 | 428 | 427 | 426 | 429 |  |  |
| n | 5 | 5 | 5 | 5 | 5 | 5 | a |
| Zr90 | 1226 | 39 | 1153 | 43 | 1171 | 49 | 2086 | 55 | 56 | 72 |
| La139 | 841 | 21 | 574 | 8 | 667 | 31 | 247 | 25 | 571 | 29 |
| Ce40 | 4.0 | 0.1 | 4 | 0.1 | 3.7 | 0.4 | 2 | 0 | 3.4 | 0.4 |
| Nd40 | 494 | 15 | 450 | 6 | 474 | 27 | 167 | 25 | 397 | 29 |
| Sm152 | 796 | 25 | 728 | 9 | 775 | 39 | 304 | 30 | 645 | 50 |
| Eu153 | 709 | 13 | 670 | 11 | 636 | 37 | 452 | 24 | 578 | 47 |
| Yb174 | 1002 | 33 | 903 | 6 | 1010 | 42 | 590 | 25 | 769 | 57 |
| Tb232 | 1019 | 21 | 967 | 14 | 984 | 51 | 465 | 37 | 813 | 57 |
| U238 | 961 | 21 | 929 | 22 | 893 | 45 | 117 | 4 | 782 | 52 |
| N60 | n.m. | n.m. | n.m. | n.m. | 34 | 12 | 108 | 3 | 71 | 1 |

| P = 1.5 GPa, T = 1650°C |  |  |  |  |  |  |  |
|---|---|---|---|---|---|---|---|
| Run | 461 | 462 | 477 | 464 |  |  |  |
| n | 5 | 5 | 5 | 5 | 5 | 5 | a |
| Zr90 | 1114 | 26 | 1191 | 19 | 413 | 4 | 311 | 8 |
| La139 | 632 | 24 | 568 | 2 | 387 | 8 | 357 | 11 |
| Ce40 | 4.7 | 0.4 | 3.8 | 0.2 | 3 | 0.1 | 2.5 | 0.1 |
| Nd42 | 489 | 19 | 441 | 6 | 276 | 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 | 29 |
| Tb232 | 980 | 24 | 1030 | 17 | 712 | 12 | 640 | 16 |
| U238 | 922 | 35 | 680 | 12 | 159 | 1 | 135 | 3 |
| N60 | n.m. | n.m. | 20 | 11 | 9 | 8 | 4 | 0 |

| P = 1.5 GPa, T = 1500°C |  |  |  |  |  |  |  |
|---|---|---|---|---|---|---|---|
| Run | 1414 | 1415 | 1416 | 1417 |  |  |  |
| n | 5 | 5 | 5 | 5 | 5 | 5 | a |
| Zr90 | 3648 | 52 | 2237 | 53 | 3455 | 35 | 3133 | 38 |
| La139 | 1558 | 25 | 1190 | 29 | 1752 | 27 | 1746 | 20 |
| Ce40 | 3.9 | 0.2 | 2.0 | 0.2 | 3.1 | 0.1 | 3.0 | 0.2 |
| Nd42 | 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 |
| Tb232 | 2553 | 51 | 2454 | 59 | 2887 | 32 | 2636 | 51 |
| U238 | 2439 | 49 | 932 | 27 | 2688 | 38 | 2606 | 45 |
| N60 | 17 | 10 | 10 | 5 | 13 | 1 | 21 | 3 |

Values are in parts per million. n is the number of measurements. n.m., not measured.
Extended Data Table 4 | Trace-element concentration in sulfides.

| P = 1.5 GPa, T = 1400°C | n | 421 | 438 | 437 | 426 | 429 |
|---------------------------|---|-----|-----|-----|-----|-----|
| Zr90                      | 5 | 5   | 5   | 5   | 5   | 5   |
| La139                     | 4.1 | 10 | 1  | 47 | 7  | 573 | 12 | 2.8 | 1.7 |
| 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                     | 3.3 | 2  | 47 | 8  | 94 | 11 | 140 | 1  | 5.0 | 2.5 |
| Yb174                     | 0.7 | 0.3 | 1.8 | 0.1 | 8  | 1  | 218 | 12 | 2.7 | 2.9 |
| Tb232                     | 0.4 | 0.1 | 0.7 | 0.1 | 3.1 | 0.3 | 339 | 6  | 5.6 | 2.8 |
| U238                      | 11 | 1  | 19 | 1  | 101 | 6  | 1792 | 94 | 14 | 3 |
| N60                       | 2837 | 64 | 3596 | 138 | 3320 | 88 | 3764 | 159 | 3153 | 51 |

| P = 1.5 GPa, T = 1650°C | n | 461 | 462 | 477 | 464 |
|---------------------------|---|-----|-----|-----|-----|
| Zr90                      | 3.5 | 0.4 | 77 | 3  | 909 | 66 | 640 | 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  |
| Tb232                     | 2.5 | 0.3 | 26 | 3  | 184 | 15 | 156 | 121 |
| U238                      | 41 | 4  | 380 | 36 | 927 | 43 | 952 | 74  |
| N60                       | 3643 | 99 | 3130 | 133 | 2573 | 181 | 1538 | 103 |

| P = 1.5 GPa, T = 1500°C | n | 1414 | 1415 | 1416 | 1417 |
|---------------------------|---|-----|-----|-----|-----|
| Zr90                      | 12 | 3   | 1629 | 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 | 0.08 | 0.02 |
| Nd142                     | 77 | 5   | 491 | 42  | 12  | 1   | 11  | 1   |
| Sm152                     | 100 | 7  | 719  | 42  | 14  | 1   | 15  | 2   |
| Eu153                     | 471 | 51  | 560  | 63  | 59  | 5   | 35  | 3   |
| Yb174                     | 16 | 2   | 225 | 54  | 3.9 | 0.2 | 8   | 5   |
| Tb232                     | 7  | 2   | 315 | 31  | 3.1 | 0.3 | 8   | 2   |
| U238                      | 216 | 36 | 2058 | 196 | 44  | 2   | 53  | 4   |
| N60                       | 10728 | 381 | 8480 | 396 | 9234 | 149 | 11866 | 72  |

Values are in parts per million. n is the number of measurements.
Extended Data Figure 1 | Partition coefficients for U, Nd and Sm with changing log[FeO] content in silicate melt (wt%). a, Results for $D$ values of experiments performed at 1.5 GPa and 1,500 °C. b, $D$ value results at 1.5 GPa and 1,650 °C.
Extended Data Figure 2 | Core content of U (p.p.b.) and mantle $^{142}$Nd anomaly (p.p.m.) at $D_{\text{Nd}}/D_{\text{Sm}}$=3. The effect on the Nd and U content using the same parameters ($D_{\text{Nd}}/D_{\text{Sm}}$ at 1.4 and $D_{\text{Th}}/D_{\text{U}}$ at 0.1) as in Fig. 2 but with a higher $D_{\text{U}}/D_{\text{Sm}}$ ratio. a and b show the calculated effect of adding to Earth a reduced body of 20% of the Earth’s mass or 45% of the Earth’s mass, containing 0.15 mass fraction sulfide. c and d illustrate the same scenario except that the reduced body contains 0.22 mass fraction sulfide.
Extended Data Figure 3 | REE fractionation at 3.2% and 8.1% S in the core. The calculated REE pattern in the bulk silicate Earth (BSE) for the two extreme cases of Fig. 1a (3.2% S) and Extended Data Fig. 2d (8.1% S). Black diamonds represent REE concentrations relative to chondritic abundances and normalized to Yb = 1, at 3.2% S in the core (20% reduced mass impactor containing 0.15 mass fraction sulphide). White diamonds illustrate the REE fractionation at elevated S content (8.1% S in the core, 35% reduced mass impactor containing 0.22 mass fraction sulphide). We assumed $D_{Sm} = [(Sm \text{ in sulphide})/(Sm \text{ in silicate})] = 1$ 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 the BSE. The trend is broadly consistent with that seen in the depleted mid-ocean-ridge basalt (MORB)–mantle composition (blue diamonds) but much smaller. The effect on the REE pattern of the BSE would, as can be seen, be undetectable. Blue diamonds illustrate the measured ratio of depleted MORB mantle (from Salters and Stracke$^{31}$) to the BSE (Palme and O’Neill$^8$, assuming chondritic abundances of refractory lithophile elements in the latter. Error bars are from propagated error calculation and correspond to 1 s.d.