The carbon content of Earth and its core

Rebecca A. Fischer1,2,3, Elizabeth Cottrellb, Erik Haurid2, Kanani K. M. Lee3,3, and Marion Le Voyerb

1Department of Earth and Planetary Sciences, Harvard University, Cambridge, MA 02138; 2Department of Mineral Sciences, National Museum of Natural History, Smithsonian Institution, Washington, DC 20560; 3Department of Earth and Planetary Sciences, University of California, Santa Cruz, CA 95064; 4Department of Terrestrial Magnetism, Carnegie Institution of Washington, Washington, DC 20015; 5and 6Department of Geology and Geophysics, Yale University, New Haven, CT 06511

Edited by David Walker, Columbia University, Palisades, NY, and approved March 2, 2020 (received for review November 12, 2019)

Earth’s core is the largest reservoir of carbon (C) in the planet, but its C abundance has been poorly constrained because measurements of carbon’s preference for core versus mantle materials at the pressures and temperatures of core formation are lacking. Using metal–silicate partitioning experiments in a laser-heated diamond anvil cell, we show that carbon becomes significantly less siderophile as pressures and temperatures increase to those expected in a deep magma ocean during formation of Earth’s core. Based on a multistage model of core formation, the core likely contains a maximum of 0.09(4) to 0.20(10) wt% C, making carbon a negligible contributor to the core’s composition and density. However, this accounts for ~80 to 90% of Earth’s overall carbon inventory, which totals 370(150) to 740(370) ppm. The bulk Earth’s carbon/sulfur ratio is best explained by the delivery of most of Earth’s volatiles from carbonaceous chondrite-like precursors.

Carbon plays a key role in living organisms, climate regulation, plate tectonics, and possibly the geodynamo (e.g., ref. 1). Deep carbon influences mantle dynamics, fluxing melting, and mobilizing structurally bound water (e.g., refs. 2 and 3). If abundant in Earth’s iron-rich metallic core (e.g., ref. 4), carbon would play an important role in core dynamics, thermal structure, and physical properties (e.g., ref. 5). Earth’s bulk carbon content informs our understanding of the planet’s building blocks, its bulk composition, and the processes of volatile delivery and loss (6). However, the bulk carbon content of Earth is poorly constrained, because a large and uncertain amount of carbon is “hidden” in Earth’s core.

Earth’s core is ~10% less dense than pure iron–nickel under the same pressures and temperatures, due primarily to the presence of “light elements” such as Si, O, S, C, and/or H (e.g., refs. 4 and 7). It is important to understand both the identities and abundances of these elements due to the roles they play in depressing the melting point of pure iron and driving compositional convection in the outer core that contributes to magnetic field generation. Knowledge of the core’s composition would greatly enhance our understanding of the thermal structure in the core, whose temperature is equal to the liquidus temperature of the core alloy at the inner core boundary, and would also constrain Earth’s bulk composition, and hence the identities of Earth’s precursory building blocks. Carbon is a good candidate light element due to its siderophile nature (e.g., ref. 8), abundance in primitive chondritic meteorites (e.g., ref. 9), and ability to match inner core seismic properties when alloyed with iron (10, 11).

Due to its volatility, carbon is expected to be depleted in Earth relative to primitive meteorites. However, estimating the bulk Earth carbon content from volatile depletion trends is challenging because the condensation temperature of C (a common measure of volatility) is difficult to define (5). While there is considerable uncertainty in the carbon abundance of Earth’s mantle, the uncertainty in the bulk Earth C content is dominated by the uncertainty in the core carbon content, as the core likely contains most of Earth’s carbon (e.g., ref. 7) and its composition is the subject of much debate. Previous estimates of Earth’s core carbon content include ~0.2 wt% based on volatility trends (7); 0.25(15) wt% based on mass balance, assuming a bulk Earth carbon content of 1,000 ppm (12); 0.6 wt% based on Mo and W metal–silicate partitioning (5); 0.5 to 2 wt% based on carbon isotopic fractionation during core formation (5); <1.1 wt% based on an estimated S content of the core and a CI chondritic C/S ratio (5); and ~2 to 4 wt% based on volatility and solubility arguments (8, 13). It has been suggested, based on the core’s shear wave velocity structure, that the core may contain enough carbon to precipitate a carbide inner core (e.g., ref. 10).

A different approach to quantifying core carbon, and thus bulk planetary carbon, is to determine how much C should have entered the core during its formation, by assessing carbon’s preference for metallic versus silicate melts under core formation conditions. This preference is quantified by a partition coefficient $D_C = C_{\text{Metal}}/C_{\text{Silicate}}$, where $C_{\text{Metal}}$ and $C_{\text{Silicate}}$ are the concentrations (in wt%) of carbon in the equilibrating metallic and silicate melts, respectively. A number of previous studies have measured the metal–silicate partitioning behavior of carbon at moderate pressures and temperatures (8, 14–22), finding a dependence on variables including pressure (P), temperature (T), oxygen fugacity ($\text{O}_2$), metallic melt composition, and nonbridging oxygen atoms per tetrahedrally coordinated cation (NBO/T, a measure of the degree of silicate melt polymerization). However, these data only extend to a maximum pressure of 15 GPa and temperature of 2573 K (19), significantly lower than the average conditions of core formation in the Earth of 45 to 55 GPa and the mantle liquidus temperature at that pressure (e.g., refs. 23 and 24). Extrapolations of these data to higher P–T suggest $D_C$ for terrestrial core formation ranging from ~10−7 (14) to ~6 × 10−7 (15). Such siderophile behavior for carbon during core formation would require an additional mechanism for increasing the carbon content of the mantle

Carbon is critical to life on Earth, climate regulation, and many geological processes. Despite its importance, the amount of carbon in the planet has been poorly understood due to uncertainty in the composition of Earth’s core, likely the largest reservoir of carbon. Here, we demonstrate that carbon becomes increasingly compatible in silicate compared to metal at the high pressures and temperatures at which Earth’s core formed. Therefore, carbon can only be present in the core in low abundances, although the core still likely holds the vast majority of the planet’s carbon.

**Significance**

Carbon is critical to life on Earth, climate regulation, and many geological processes. Despite its importance, the amount of carbon in the planet has been poorly understood due to uncertainty in the composition of Earth’s core, likely the largest reservoir of carbon. Here, we demonstrate that carbon becomes increasingly compatible in silicate compared to metal at the high pressures and temperatures at which Earth’s core formed. Therefore, carbon can only be present in the core in low abundances, although the core still likely holds the vast majority of the planet’s carbon.


The authors declare no competing interest.

This article is a PNAS Direct Submission. Published under the PNAS license.

1To whom correspondence may be addressed. Email: rebeccafischer@g.harvard.edu.

2Deceased September 5, 2018.

3Present address: Physics Division, Lawrence Livermore National Laboratory, Livermore, CA 94550.

This article contains supporting information online at https://www.pnas.org/lookup/suppl/doi:10.1073/pnas.1919930117/-/DCSupplemental.

after core formation to match the observed value (e.g., refs. 8 and 18), such as delivery of carbon in the late veneer (Discussion).

Results
To investigate the metal–silicate partitioning behavior of carbon at conditions more directly applicable to Earth’s core formation, we performed experiments on mixtures of olivine, iron, nickel, and graphite (undersaturated with respect to carbon), using a laser-heated diamond anvil cell to achieve pressures of 37 to 59 GPa and temperatures above the silicate liquidus, 4200 to 5200 K (SI Appendix, Table S1). Multispectral imaging radiometry was used to map temperatures in two dimensions during the experiments (25). Following the experiments, we recovered the samples to ambient conditions and hand polished them. We measured carbon in the silicate glass using nano-secondary ionization mass spectrometry (nanoSIMS), and we measured all major and minor elements in the silicate and metal, including carbon, using a field emission electron probe microanalyzer (EPMA) (Fig. 1 and SI Appendix, Tables S1 and S2). Silicate carbon analyses using the two methods were used to bracket the partitioning behavior of carbon (Materials and Methods and SI Appendix).

We find that carbon is approximately two orders of magnitude less siderophile at the deep magma ocean conditions of this study compared to results of previous studies at more modest pressures and temperatures (8, 14–22) ($D_C = 1$ to 100 vs. $\sim 100$ to 10,000; Fig. 2), consistent with a previous ab initio calculation of $D_C = 9(3)$ at 40 GPa and 3200 K (26). We performed a stepwise, unweighted linear regression to the data of this study and 100 measurements of C metal–silicate partitioning in the literature (8, 14–20) (Materials and Methods, Dataset S1, and SI Appendix, Fig. S1), with C partitioning exhibiting a statistically significant ($P < 0.05$) dependence on $T$, $P$, mole fractions of sulfur and oxygen in the metal ($X_S$ and $X_O$), NBO/T, and $fO_2$ in the combined dataset. Using either the nanoSIMS (Eq. 1) or microprobe (Eq. 2) analyses of carbon in the silicate from this study, the resulting fits are, respectively:

$$\log_{10}(D_C) = 1.49(70) + \frac{3.000 (1,110)}{T} - \frac{235 (60) * P}{T}$$

$$+ 9.6(14) * \log_{10}(1 - X_S) - 19.5(89) * \log_{10}(1 - X_O)$$

$$- 0.118(42) * \text{NBO/T} - 0.238(92) * \Delta \text{IW},$$

[1]

$$\log_{10}(D_C) = 1.81(74) + \frac{2.470 (1,170)}{T} - \frac{227 (63) * P}{T}$$

$$+ 9.7(14) * \log_{10}(1 - X_S) - 30.6(99) * \log_{10}(1 - X_O)$$

$$- 0.123(46) * \text{NBO/T} - 0.211(94) * \Delta \text{IW},$$

[2]

with $T$ in kelvin, $P$ in gigapascals, and $fO_2$ in log$_{10}$ units relative to the iron–wüstite (IW) buffer (numbers in parentheses are uncertainties on the last digit[s]; covariance matrices of these fits are shown in SI Appendix, Tables S3 and S4). Measured versus predicted partition coefficients from Eq. 1 are shown in SI Appendix, Fig. S2. The magnitudes of the terms in the two equations are in agreement within their mutual uncertainties. Eqs. 1 and 2 indicate that carbon becomes less siderophile with increasing pressure, the opposite dependence than previously determined in low pressure studies from 1 to 5 GPa (8, 15, 16), but in agreement with the three previous highest-pressure studies of carbon partitioning, which spanned pressures up to 12 to 15 GPa (19, 21, 22). C also becomes less siderophile with increasing NBO/T (in agreement with many previous studies, e.g., refs. 8, 15, 17–19, and 21), increasing sulfur in the metal (in agreement with refs. 16 and 22), decreasing oxygen in the metal (not previously investigated, since oxygen only dissolves into metallic melts at high $P$–$T$; e.g., ref. 23), and increasing oxygen fugacity (in agreement with many previous studies, e.g., refs. 14, 15 and 19–21; although this behavior seems to reverse at sufficiently low $fO_2$; refs. 17 and 18). At low pressures, carbon becomes less siderophile with increasing temperature (Fig. 2), in agreement with most previous studies (8, 16, 18, 19, 22). At pressures above ~15 GPa, Eqs. 1 and 2 predict that carbon becomes more siderophile with increasing temperature (Fig. 2), although we caution that it is difficult to deconvolve pressure and temperature effects as these correlate in experimental datasets. Eqs. 1 and 2 should be regarded as empirical fits that adequately describe the behavior of carbon over the conditions of the experimental data (Fig. 2).
et al. (14) in silicate glass from similar conditions. The carbon bands resemble those reported by Armstrong of carbon speciation do not affect our major findings or conclusions. However, the details of the metal and several interpretations are possible. However, the data of carbon speciation in the volatile species were present in detectable quantities. Glass speciation is consistent with a low water abundance in these experiments. 

Discussion

The metal–silicate partition coefficients determined in this study are smaller than those previously reported at lower P–T (8, 14–20), although there is some overlap, especially with the previous studies that extended to the highest pressures (e.g., ref. 19) (SI Appendix). A change in partition coefficient of similar magnitude is seen in silicon over the same P–T range, for example (e.g., refs. 8, 16, 18, 19, and 22). This partitioning behavior for carbon is consistent with a positive 1/T dependence found by most previous studies (e.g., refs. 8, 16, 18, 19, and 22). It is also consistent with a negative pressure dependence found by the previous studies that spanned the largest pressure ranges (19, 21, 22), a factor of >2 greater P range than other studies (8, 14–18, 20). Additionally, the experiments presented here were undersaturated with respect to carbon, while the experiments in most lower P–T studies (8, 14–20, 22) were performed at carbon-saturated conditions. A recent study (21) suggested that lower partition coefficients result from carbon-undersaturated conditions, which are more applicable to conditions in the Earth. For single-stage core formation at 54 GPa (23), the peridotite liquidus temperature, 4100 K (27), NBO/T = 2.6, IW–2.2, XCO = 0.06 (e.g., ref. 28), and XH = 0.03 (e.g., ref. 7), Eqs. 1 and 2 yield a partition coefficient for carbon of 0.5 to 1.8. At these conditions, extrapolations of the parameterizations of previous studies give partition coefficients ranging from 0.5 (19), in good agreement with our results, to 2 × 10^7 (16). For a mantle carbon content of 120 ppb (29), a partition coefficient of 0.5 to 1.8 implies a core carbon content of 66(55) to 220(190) ppb. More oxygen in the core would result in higher core carbon contents; for example, using XCO = 0.12 (e.g., ref. 30) results in a core with 0.024(15) to 0.16(11) wt% C. Fig. 3 illustrates the trade-offs between core formation depth (pressure and temperature), core oxygen content, and core carbon content in this single-stage core formation model. Abundances of most major, minor, and trace elements in the Earth’s mantle and core are most consistent with a series of metal–silicate equilibration events at increasing P–T, although there are

![Fig. 2. Metal–silicate partition coefficients of carbon from this study (solid rectangles) and previous studies (open symbols; Dataset SI). Data are color-coded by pressure and are corrected to a common NBO/T = 2.6, XCO = XH = 0, and fO2 = IW–2.2 using Eq. 2 (or Eq. 1 for the nanoSIMS data from this study). Triangles: 14. Circles: 15. Squares: 8. Hexagons: 16. Stars: 17. Pentagons: 18. Inverted triangles: 19. Diamonds: 20. × symbol: ab initio calculation (26, uncorrected). Data from this study are shown as ranges encompassing microprobe and nanoSIMS analyses of carbon in the silicate glass (one datum only has a lower bound, shown as an arrow). Solid lines: fit to complete dataset, using nanoSIMS analyses for silicate C from the present study (Eq. 1). Dashed lines: fit to complete dataset, using microprobe analyses for silicate C from the present study (Eq. 2). Curves are shown for NBO/T = 2.6, XCO = XH = 0, fO2 = IW–2.2, and pressures of 0, 15, 30, 45, and 60 GPa. (Inset) Uncorrected metal–silicate partition coefficients.](https://www.pnas.org/content/117/16/8745)

![Fig. 3. Core carbon content as a function of the pressure of core formation (with temperature along the peridotite liquidus of 27) and oxygen content of the core. Calculation is for single-stage core formation, assuming a mantle carbon content of 120 ppm (29), a peridotitic NBO/T of 2.6, oxygen fugacity of IW–2.2, and a core sulfur content of XH = 0.03 (e.g., ref. 7). Dotted lines: approximate core oxygen contents from previous studies (28, 30, 34). (A) Based on fit using nanoSIMS analyses for silicate C from the present study (Eq. 1). (B) Based on fit using microprobe analyses for silicate C from the present study (Eq. 2).](https://www.pnas.org/content/117/16/8745)
shown to be necessary to reproduce the mantle in oxygen fugacity as accretion proceeds has been previously to approximately reproduce the FeO, NiO, and SiO2 contents of equilibration (e.g., ref. 39). These model conditions were chosen content of 120(60) ppm (29), the core contains 0.09(4) to 0.20(10) 1.2 wt% O, 0.9 wt% S, and 5.2 wt% Ni. For a mantle carbon the mantle (29) (Fig. 4). The resulting core contains 7.8 wt% Si, the accretion of more oxidized material from the outer disk to- ward the end of planetary accretion (e.g., ref. 37) and the redox reaction SiO2 → 2FeO + Si, which proceeds to the right as pressures and temperatures increase (e.g., refs. 23, 34, and 38). The degree of metal equilibration was k = 0.4, consistent with the Earth’s Hf–W isotopic composition for the case of whole-mantle equilibration (e.g., ref. 39). These model conditions were chosen to approximately reproduce the FeO, NiO, and SiO2 contents of the mantle (29) (Fig. 4). The resulting core contains 7.8 wt% Si, 1.2 wt% O, 0.9 wt% S, and 5.2 wt% Ni. For a mantle carbon content of 120(60) ppm (29), the core contains 0.09(4) to 0.20(10) wt% C (range corresponding to the use of Eqs. 1 and 2, respectively).

These core carbon abundances indicate that carbon is unlikely to be a major contributor to Earth’s core density deficit with respect to pure iron. To explain the seismologically observed density, Earth’s core must contain significant amounts of other light elements such as silicon and oxygen (e.g., refs. 24, 28, and 34), both of which become more siderophile at the high temperatures of Earth’s core formation (e.g., refs. 23 and 24), and/or sulfur and hydrogen. Despite carbon comprising a small fraction of the core, the core remains the largest reservoir of carbon in the Earth, containing 78(12) to 89(7)% of the planet’s carbon budget. Therefore, any analysis of the Earth’s carbon content must account for this dominant reservoir, which has implications for understanding of the planet’s bulk composition and volatile inventory and delivery. This core composition corresponds to a bulk Earth carbon content of 2.2(9) to 4.2(22) × 1021 kg, or 370(150) to 740(370) ppm.

This analysis assumes that carbon was delivered throughout Earth’s accretion history, such that the depth of metal–silicate equilibration that reproduces the mantle’s major and trace element composition also applies to carbon. N-body simulations of terres- trial planet accretion show that Earth-like planets preferentially accrete material from the outer disk, which is likely more volatile-rich, later in their growth histories (e.g., ref. 37). If more of Earth’s carbon were accreted later in its history, when the Earth was larger, the average P–T conditions of metal–silicate equilibration relevant for carbon may have been significantly higher. With increasing depth along the peridotite liquidus (27), carbon generally becomes even less siderophile (Eqs. 1 and 2). Therefore, the actual core carbon abundance, and thus the bulk Earth carbon abundance, may be even lower than that determined here. For example, if carbon equilibration occurred at 50% higher pressures (representing

Fig. 4. Results of a multistage model of core formation. (A) Evolution of the carbon metal–silicate partition coefficient Dc as Earth grows, using either Eq. 1 (fit based on nanoSIMS measurements; gray symbols and curve) or Eq. 2 (fit based on microprobe measurements; black symbols and curve). (B) Mantle carbon content. Colors are as in part A. (C) Core silicon (red symbols and curve) and oxygen (dark blue symbols and curve) contents. (D) Mantle FeO (light blue symbols and curve, left axis) and NiO (orange symbols and curve, right axis) contents.
The carbon content derived from C/Nb and C/Ba ratios of undegassed Earth (36), we find a bulk Earth C/S ratio of 0.12(5) to 0.25(11). This ratio is lower than the bulk silicate Earth C/S ratio of 0.49(14) (41), assuming that the MgO insulation layer remained transparent, so it exhibited no wavelength-dependent absorption and emission, and that the unreacted silicate starting material remained olivine (Fo90) and retained low absorption coefficients (<100 cm⁻¹) (48). Taking into account these assumptions and the difference in temperatures between the two sides of the experiment, estimated temperature uncertainties are ±10%. Experimental pressures, temperatures, and accompanying uncertainties are summarized in SI Appendix, Table S1.

Sample Recovery and Analysis. Following decomposition, we mounted the entire assembly in epoxy under vacuum. We polished either parallel or perpendicular to the compression axis to expose the laser-heated spot for analysis. We cut away the epoxy medium and embedded the sample in a new epoxy medium, then coated all samples and standards with Ir for analysis of light elements first by nanoSIMS, followed by EPMA. After obtaining surprisingly high carbon contents in the quenched silicate melt (glass) by nanoSIMS, we analyzed all elements by EPMA, including direct analysis of carbon and oxygen. Silicate carbon contents measured by nanoSIMS were higher than the carbon contents measured by EPMA, and therefore the nanoSIMS measurements constrain the minimum partition coefficients of carbon. In light of this surprising result, we take the conservative approach of bracketing the carbon contents of the experiments with results from nanoSIMS and EPMA (SI Appendix).

We carried out nanoSIMS analyses on the quenched silicate using the Cameca NanoSIMS 50L at the Carnegie Institution of Washington (e.g., ref. 49). We measured isotopes 12C, 13C, 16O, 28Si, 56Fe, and 60Ni simultaneously using a pulsed IR laser scanning ion implanter (41) and both mass 28Si and 26Si measured. We added a Cs+ primary beam. We applied a beam current of 4 to 10 pA, with a raster size ranging from 10 × 10 μm² to 25 × 25 μm² after a 160-s presputter over an area 5 μm larger (to remove surface contamination before the analyses). We collected 15 frames per image with a 16-s count time, at 256 × 256 pixels. We used 12C/13C to quantify carbon abundances based on a calibration (linear regression with R² > 0.98) built from five silicate glasses with known carbon contents in the range 45 to 1,500 ppm C. SIMS carbon calibrations are linear up to at least >2.4 wt% C when normalizing against 60Ni (50); we expect similar linear behavior when normalizing against 18O (e.g., refs. 50–52) (SI Appendix). We processed the data using the L’Image software package for IDL (L. R. Nittler, Carnegie Institution) and corrected all data for spatial drift and deadtime. We averaged over 15 to 30 nonoverlapping regions of interest (ROIs), with ROIs chosen at the edges of the silicate glass and any decompression fractures. Results are insensitive to the number/size of ROIs used. Quoted uncertainties include contributions from the spatial variability of the silicate glasses (the SD of multiple ROIs) and the uncertainty on the calibration fit and extrapolation. Analytical uncertainties (typically 1 to 3% relative) are negligible in comparison to these other sources. Carbon abundances measured by nanoSIMS can be found in SI Appendix, Table S1.

We quantified all major and minor elements in the quenched metal and silicate at the Smithsonian National Museum of Natural History using a JEOL JXA-8530F field emission electron probe microanalyzer. We Ir-coated all standards and unknowns simultaneously. We employed a liquid-nitrogen cold-finger to minimize the carbon background. The accelerating voltage was 10 keV, beam current was 10 nA, and beam diameter ranged from 3 μm when the available area was sufficiently large, down to a point analysis when it was not. The beam size (including activation volume) was large enough to average over quench features, and measured carbon was insensitive to beam size. We used the following crystals for quantification: LIFL for Fe kα and Ni kα; TAPl for Si kα and Mg kα; LDE1L for O kα; and LDESL for C kα. We counted for 10 s on peak and for 5 s at high and low background off-peak positions. We employed the following primary standards (Ir-coated) for analysis of the melt phase: stoichiometric Fe₂C for carbon, Ni₁₆Fe₂₈ for nickel and iron, Mg metal for magnesium, Harvard hematite 92649 for oxygen, and Si metal for silicon. We employed the following primary standards (Ir-coated) for analysis of the silicate phase: stoichiometric...
FeC for carbon, Ni0.3Fe0.7 for nickel, and San Carlos olivine for magnesium, iron, and silicon (oxygen assigned stoichiometrically). We monitored for heated spots (0.3 wt% C in silicates and 0.3 to 0.6 wt% C in metals, based on analysis of multiple carbon-free standards and ferropericlase elements except carbon, which was exponential. The matrix correction for the runs as secondary standards. Off-peak corrections were linear for all totals in the metals of some samples, possibly due to quench textures (53).

We performed confocal Raman spectroscopy on the quenched silicate glass of sample R193 (SI Appendix, Fig. S3). This sample exhibited areas with few heated spots and possible carbon contamination from the diamond anvils (e.g., refs 23 and 54).

We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.

ACKNOWLEDGMENTS. This work was funded by a National Science Foundation Postdoctoral Fellowship (EAR-1452626) to R.A.F. This work was performed in part under the auspices of the US Department of Energy by Lawrence Livermore National Laboratory under Contract DE-AC52-07NA27344. We are grateful to Jackie Li and David Walker for providing microprobe standards; Jennifer Girard and Terry-Ann Suer for assistance with laser-heating experiments; Tim Gooding, Tim Rose, Colin Jackson, Emma Bullock, Rob Wardell, and Jianhua Zhang with sample preparation and analysis; My Bieba for assistance with nanoSIMS data processing; and Francis Nimmo and Jie Deng for helpful advice. We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.

FeC for carbon, Ni0.3Fe0.7 for nickel, and San Carlos olivine for magnesium, iron, and silicon (oxygen assigned stoichiometrically). We monitored for heated spots (0.3 wt% C in silicates and 0.3 to 0.6 wt% C in metals, based on analysis of multiple carbon-free standards and ferropericlase elements except carbon, which was exponential. The matrix correction for the runs as secondary standards. Off-peak corrections were linear for all totals in the metals of some samples, possibly due to quench textures (53).

We performed confocal Raman spectroscopy on the quenched silicate glass of sample R193 (SI Appendix, Fig. S3). This sample exhibited areas with few heated spots and possible carbon contamination from the diamond anvils (e.g., refs 23 and 54).

We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.

ACKNOWLEDGMENTS. This work was funded by a National Science Foundation Postdoctoral Fellowship (EAR-1452626) to R.A.F. This work was performed in part under the auspices of the US Department of Energy by Lawrence Livermore National Laboratory under Contract DE-AC52-07NA27344. We are grateful to Jackie Li and David Walker for providing microprobe standards; Jennifer Girard and Terry-Ann Suer for assistance with laser-heating experiments; Tim Gooding, Tim Rose, Colin Jackson, Emma Bullock, Rob Wardell, and Jianhua Zhang with sample preparation and analysis; My Bieba for assistance with nanoSIMS data processing; and Francis Nimmo and Jie Deng for helpful advice. We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.

FeC for carbon, Ni0.3Fe0.7 for nickel, and San Carlos olivine for magnesium, iron, and silicon (oxygen assigned stoichiometrically). We monitored for heated spots (0.3 wt% C in silicates and 0.3 to 0.6 wt% C in metals, based on analysis of multiple carbon-free standards and ferropericlase elements except carbon, which was exponential. The matrix correction for the runs as secondary standards. Off-peak corrections were linear for all totals in the metals of some samples, possibly due to quench textures (53).

We performed confocal Raman spectroscopy on the quenched silicate glass of sample R193 (SI Appendix, Fig. S3). This sample exhibited areas with few heated spots and possible carbon contamination from the diamond anvils (e.g., refs 23 and 54).

We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.

ACKNOWLEDGMENTS. This work was funded by a National Science Foundation Postdoctoral Fellowship (EAR-1452626) to R.A.F. This work was performed in part under the auspices of the US Department of Energy by Lawrence Livermore National Laboratory under Contract DE-AC52-07NA27344. We are grateful to Jackie Li and David Walker for providing microprobe standards; Jennifer Girard and Terry-Ann Suer for assistance with laser-heating experiments; Tim Gooding, Tim Rose, Colin Jackson, Emma Bullock, Rob Wardell, and Jianhua Zhang with sample preparation and analysis; My Bieba for assistance with nanoSIMS data processing; and Francis Nimmo and Jie Deng for helpful advice. We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.

FeC for carbon, Ni0.3Fe0.7 for nickel, and San Carlos olivine for magnesium, iron, and silicon (oxygen assigned stoichiometrically). We monitored for heated spots (0.3 wt% C in silicates and 0.3 to 0.6 wt% C in metals, based on analysis of multiple carbon-free standards and ferropericlase elements except carbon, which was exponential. The matrix correction for the runs as secondary standards. Off-peak corrections were linear for all totals in the metals of some samples, possibly due to quench textures (53).

We performed confocal Raman spectroscopy on the quenched silicate glass of sample R193 (SI Appendix, Fig. S3). This sample exhibited areas with few heated spots and possible carbon contamination from the diamond anvils (e.g., refs 23 and 54).

We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.

ACKNOWLEDGMENTS. This work was funded by a National Science Foundation Postdoctoral Fellowship (EAR-1452626) to R.A.F. This work was performed in part under the auspices of the US Department of Energy by Lawrence Livermore National Laboratory under Contract DE-AC52-07NA27344. We are grateful to Jackie Li and David Walker for providing microprobe standards; Jennifer Girard and Terry-Ann Suer for assistance with laser-heating experiments; Tim Gooding, Tim Rose, Colin Jackson, Emma Bullock, Rob Wardell, and Jianhua Zhang with sample preparation and analysis; My Bieba for assistance with nanoSIMS data processing; and Francis Nimmo and Jie Deng for helpful advice. We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.

FeC for carbon, Ni0.3Fe0.7 for nickel, and San Carlos olivine for magnesium, iron, and silicon (oxygen assigned stoichiometrically). We monitored for heated spots (0.3 wt% C in silicates and 0.3 to 0.6 wt% C in metals, based on analysis of multiple carbon-free standards and ferropericlase elements except carbon, which was exponential. The matrix correction for the runs as secondary standards. Off-peak corrections were linear for all totals in the metals of some samples, possibly due to quench textures (53).

We performed confocal Raman spectroscopy on the quenched silicate glass of sample R193 (SI Appendix, Fig. S3). This sample exhibited areas with few heated spots and possible carbon contamination from the diamond anvils (e.g., refs 23 and 54).

We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.

ACKNOWLEDGMENTS. This work was funded by a National Science Foundation Postdoctoral Fellowship (EAR-1452626) to R.A.F. This work was performed in part under the auspices of the US Department of Energy by Lawrence Livermore National Laboratory under Contract DE-AC52-07NA27344. We are grateful to Jackie Li and David Walker for providing microprobe standards; Jennifer Girard and Terry-Ann Suer for assistance with laser-heating experiments; Tim Gooding, Tim Rose, Colin Jackson, Emma Bullock, Rob Wardell, and Jianhua Zhang with sample preparation and analysis; My Bieba for assistance with nanoSIMS data processing; and Francis Nimmo and Jie Deng for helpful advice. We thank the editor for handling our submission and the two anonymous reviewers for helpful comments.