Heterogeneity in mantle carbon content from CO₂-undersaturated basalts

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The amount of carbon present in Earth’s mantle affects the dynamics of melting, volcanic eruption style and the evolution of Earth’s atmosphere via planetary outgassing. Mantle carbon concentrations are difficult to quantify because most magmas are strongly degassed upon eruption. Here we report undegassed carbon concentrations from a new set of olivine-hosted melt inclusions from the Mid-Atlantic Ridge. We use the correlations of CO₂ with trace elements to define an average carbon abundance for the upper mantle. Our results indicate that the upper mantle carbon content is highly heterogeneous, varying by almost two orders of magnitude globally, with the potential to produce large geographic variations in melt fraction below the volatile-free solidus. Such heterogeneity will manifest as variations in the depths at which melt becomes interconnected and detectable, the CO₂ fluxes at mid-ocean ridges, the depth of the lithosphere-asthenosphere boundary, and mantle conductivity.
The carbon content of undegassed mid-oceanic ridge basalts (MORB) and of the upper mantle has been an ongoing debate for several decades. Carbon is a volatile element that plays a key role in major geodynamical processes such as mantle melting and volcanic degassing. The amount of carbon present in the mantle will affect the onset of deep melting, the geophysical properties of the mantle, as well as long-term climate change when CO$_2$ is released into the atmosphere. Because of its very low solubility in magmas, magmatic degassing depletes carbon in the melt during ascent and eruption, which prevents direct measurements of the original carbon content of most basaltic melts formed in equilibrium with the mantle source. Previous studies have used indirect approaches to correct for degassing, such as isotopic fractionation models, vesicle size distribution or the composition of the gas trapped in vesicles.

Our knowledge of mantle carbon is best constrained by direct measurements of only two undegassed samples: the popping rock and the Siqueiros melt inclusions. Together, these two samples show correlations between CO$_2$ content and non-volatile trace elements (such as Nb) that directly constrain the amount of carbon in the upper mantle, and thus both samples have served as references for CO$_2$ fluxes due to volcanism and its effect on long-term climate changes. Yet, the two most recent studies of these samples show that their CO$_2$/Nb ratios differ by more than a factor of two, suggesting that the mantle CO$_2$/Nb ratio is either variable or constant, and propose upper mantle CO$_2$ abundances and mid-ocean ridge CO$_2$ fluxes that vary by a factor of four. Another study reports unusually high CO$_2$ contents in melts from the Juan de Fuca ridge. However the decoupling between their CO$_2$ and Nb contents indicates that part of the CO$_2$ was lost through degassing before entrapment.

A recent study compiled most CO$_2$ measurements from published MORB glasses and melt inclusions, along with new data from 15 undegassed MORB glasses that are undersaturated in CO$_2$ and do not contain any vesicles. Because of this, the authors argue that they are mostly undegassed. These undegassed MORB glasses have average CO$_2$/Rb, CO$_2$/Ba and CO$_2$/Nb ratios that are close to those of the Siqueiros melt inclusions, also undegassed. Studies of additional undegassed samples are critical to understanding whether these differences are due to sampling bias or real geologic variability.

Here we report the CO$_2$ contents of an independent set of olivine-hosted melt inclusions from the equatorial Mid-Atlantic Ridge (MAR). Using correlations between CO$_2$ and highly incompatible elements, we show that these melt inclusions represent another rare occurrence of undegassed MORB, and we discuss the implications of our results with respect to the carbon content of the mantle and the carbon flux from the global ridge system.

**Results**

**Description of the sample.** We analysed the major, trace and volatile elements of Fo$_{56-90}$ olivine-hosted melt inclusions from the glassy rim of a pillow basalt from the equatorial MAR. Sample EN061-5D-3A was dredged on-axis by R/V *Endeavor* in 1981 (5.185°S, 11.517°W) (refs 14–16). The host glass is a depleted MORB with La/Sm$_N=0.5$, similar to the average depleted MORB composition (D-MORB, La/Sm$_{N}=0.5$ (refs 17,18)). The matrix glasses contain 0.118 ± 0.004 wt% H$_2$O, 206 ± 14 p.p.m. CO$_2$, 102 ± 5 p.p.m. F, 1,014 ± 66 p.p.m. S and 17.8 ± 2 p.p.m. Cl (uncertainty is 2 s.d. on replicate analyses from 15 glass chips), which is typical for a depleted MORB erupted at ~3,300 m below sea level. The matrix glasses are saturated to slightly oversaturated with respect to a CO$_2$-rich vapour phase at the pressure of eruption. Their estimated pressure of vapour saturation is P$_{sat} = 444 ± 30$ bar (ref. 10).

**Description of the melt inclusions.** From this sample we analysed 23 melt inclusions for major, trace and volatile element content; all of the melt inclusions are free of shrinkage bubbles. They are glassy, tholeiitic basalts with MgO contents slightly higher than their host glasses. All but three of the melt inclusions have undergone <3% post entrapment crystallization (PEC), and so although we refer herein to their PEC-corrected chemistry, the correction is inconsequential to our results (see Methods). Incompatible minor and trace element contents range from highly depleted to compositions identical to those of the matrix glass.

**Trace element composition of the melt inclusions.** We used the high-resolution trace element data set to assess which process is responsible for the large range and co-variation of the highly incompatible element contents in the equatorial Atlantic melt inclusions. In a plot of CH/CM versus CM, where CH and CM are the concentrations of a highly incompatible trace element such as Rb or Th and a moderately incompatible element such as Nd, partial melting is expressed as a straight line with a slope = 1, while fractional crystallization would produce an almost horizontal line, and mixing would produce a curve. In Fig. 1a, we see that the equatorial MAR melt inclusions plot on a straight line that is not horizontal, ruling out fractional crystallization as a main relationship explaining the range in

![Figure 1](https://example.com/figure1.png) **Figure 1 | Variation in trace element composition of the equatorial MAR melt inclusions.** (a) Th/Nd ratio as a function of Th content (p.p.m.) for the equatorial melt inclusions (red diamonds) and matrix glasses (pink diamonds). (b) Rb content (p.p.m.) as a function of Rb/Sc ratio for the equatorial MAR melt inclusions and matrix glasses.
composition of these melt inclusions. The small range in the major element contents of the melt inclusions (Supplementary Table 1) support this conclusion as well. Similarly, in a plot of Ce³⁺ versus C¹⁻²⁻²⁻, where Ce³⁺ is the concentration of a compatible element such as Sc, mixing and crystallization would produce hyperbolic curves, while partial melting would generate a straight line²⁵. In Fig. 1b, the straight line defined by the equatorial Atlantic melt inclusions confirms that they are related by relatively few degrees of melting. However, we cannot rule out mixing as a possible cause for the large range in trace element compositions. The range in degree of melting calculated from the major element compositions (11–23%, using Na₂O following ref. 24) is too small to account for the range in incompatible elements observed in the equatorial Atlantic melt inclusions (for example, Rb contents vary from 0.05 to 0.66, a 12-fold increase) using a simple batch melting model of a DMM-type mantle source²⁵. The range in incompatible elements can be easily reproduced by modelling compositions of incremental melts produced during fractional melting, using a melt fraction varying from 0.7 to 1.0% (ref. 26).

The range in incompatible trace element compositions produced during fractional melting, using a melt fraction varying from 0.7 to 1.0% (ref. 26). The range in incompatible trace element compositions could also reflect variable amounts of mixing between very low degree incremental melts early in the melting process (highly enriched in incompatible trace elements, such as the carbon-rich melts formed at depth beneath oceanic ridges¹), and aggregated melts produced later in the melting process by ~14% of melting. The incomplete pooling of melts is similar to the process described for the Famous segment²⁷, where some of the melt inclusions trapped in olivines record compositions systematically more depleted than the matrix glasses. Alternatively, the large variation in trace elements could reflect a variable source composition, that is, a heterogeneous source containing a trace-element depleted reservoir before eruption. If we assume that all melt inclusions had a H₂O/Ce ratio similar to those of the matrix glasses (0.118 ± 0.004 wt%), At the depth of eruption (~3,300 m.b.s.l.), water degassing is only minimal (the equilibrium vapour phase contains <0.3 mol% of H₂O); thus H₂O/Ce should not be affected by degassing. Because most melt inclusions are depleted in trace elements compared to the matrix glasses, they should record lower H₂O contents as well. The similar H₂O contents recorded in the melt inclusions and the matrix glass illustrates a late-stage H₂O enrichment in these depleted melts by diffusive re-equilibration of hydrogen²⁹ between the H₂O-depleted melt inclusions and the H₂O-rich matrix glasses, most likely during storage and differentiation in a shallow magma reservoir before eruption. If we assume that all melt inclusions had a H₂O/Ce ratio similar to those of the matrix glasses (245 ± 12), we can reconstruct the initial H₂O content of the melt inclusions (0.03–0.12 wt% H₂O, average of 0.07 wt% H₂O). The amount of H₂O gain through diffusion varies from none up to 2.8 times compared to the estimated initial H₂O contents. We report in Fig. 2b the Cl content of the melt inclusions and matrix glasses as a function of their Nb content. Unlike most MORB samples, which tend to show variable amounts of seawater-derived Cl contamination, and therefore a variable Cl/Nb, the equatorial Atlantic melt inclusions have a constant Cl/Nb ratio of 14.3 ± 3.8. A similar value is recorded in the matrix glasses. This indicates that the equatorial melt inclusions as well as matrix glasses did

**Volatile element composition of the melt inclusions.** The melt inclusions contain 68–719 p.p.m. CO₂, 52–90 p.p.m. F, 779–1,087 p.p.m. Cl and 0.10–0.14 wt% H₂O. Compared to the matrix glasses, which have suffered from strong CO₂ degassing typical of most MORB glasses, the saturation pressures of the melt inclusions reflect minimum entrapment depths down to 4 km below the sea floor (P_sat = 150–1,500 bar). Because we do not observe shrinkage bubbles in the melt inclusions, the measured contents for the other volatiles, particularly CO₂, represent those of the melts at the time they were trapped in olivine. The large range in volatile content is consistent with the range in other incompatible trace elements found in the melt inclusions. In Fig. 2a, we see that the equatorial MAR melt inclusions span a large range of Ce contents, similarly to other incompatible trace elements. However, H₂O, which partitions similarly to Ce (ref. 28) only spans a very limited range of concentration (0.10–0.14 wt%), with an average H₂O content in the melt inclusions identical to that of the matrix glass (0.118 ± 0.004 wt%). At the depth of eruption (~3,300 m.b.s.l.), water degassing is only minimal (the equilibrium vapour phase contains <0.3 mol% of H₂O); thus H₂O/Ce should not be affected by degassing. Because most melt inclusions are depleted in trace elements compared to the matrix glasses, they should record lower H₂O contents as well. The similar H₂O contents recorded in the melt inclusions and the matrix glass illustrates a late-stage H₂O enrichment in these depleted melts by diffusive re-equilibration of hydrogen²⁹ between the H₂O-depleted melt inclusions and the H₂O-rich matrix glasses, most likely during storage and differentiation in a shallow magma reservoir before eruption. If we assume that all melt inclusions had a H₂O/Ce ratio similar to those of the matrix glasses (245 ± 12), we can reconstruct the initial H₂O content of the melt inclusions (0.03–0.12 wt% H₂O, average of 0.07 wt% H₂O). The amount of H₂O gain through diffusion varies from none up to 2.8 times compared to the estimated initial H₂O contents. We report in Fig. 2b the Cl content of the melt inclusions and matrix glasses as a function of their Nb content. Unlike most MORB samples, which tend to show variable amounts of seawater-derived Cl contamination, and therefore a variable Cl/Nb, the equatorial Atlantic melt inclusions have a constant Cl/Nb ratio of 14.3 ± 3.8. A similar value is recorded in the matrix glasses. This indicates that the equatorial melt inclusions as well as matrix glasses did

**Figure 2 | Behaviour of H₂O and Cl in the equatorial MAR melt inclusions.** (a) Variation in H₂O content (wt%) as a function of Ce content (p.p.m.) for the equatorial MAR melt inclusions and matrix glasses. (b) Variation in Cl content (p.p.m.) as a function of Nb content (p.p.m.) for the equatorial melt inclusions and matrix glasses.
not suffer from significant alteration or assimilation of hydrothermal material.

**Correlation between CO₂ and other trace elements.** The CO₂ content of the melt inclusions strongly correlates with other highly incompatible elements such as Ba, Rb, and Nb, which indicates that these melt inclusions did not lose their initial carbon through degassing (Fig. 3). Note that the CO₂ content of the equatorial melt inclusions also strongly correlates with their Cl content (CO₂/Cl = 39 ± 4, 2 s.d., linear regression of R² = 0.912). For comparison, the CO₂/Cl ratio of Siqueiros melt inclusions is 77 ± 26 (ref. 4), and the CO₂/Cl ratio of popping rock is 46 (ref. 2). However, the CO₂/Cl ratio was not used below because, although the equatorial melt inclusions do not show any signs of seawater contamination, most MORB samples do. Therefore the use of the CO₂/Cl ratio as a source proxy cannot be generalized to the global scale, as most MORB samples are strongly contaminated with seawater-derived Cl (ref. 30).

**Discussion**

Carbon is present as carbonate in the uppermost mantle under the oxidizing conditions relevant for the formation of MORB (refs 4,31). Because very small extents of melting will efficiently remove carbonate minerals from the source, further melting will only dilute carbon in the pooled magma. The CO₂-Nb correlation found in the Siqueiros melt inclusions indicates that C partitions similarly to Nb during melting and crystallization. However, the degree of melting typically of MORB generation, nearly all of the CO₂, Rb, Ba and Nb from the mantle source are contained in the MORB melt, and thus CO₂/Nb is not a perfect canonical ratio at low degrees of melting.

Observations from a global compilation on MORB glasses also suggest that Ba is the best proxy for C to model the CO₂ content of MORB and of the suboceanic mantle. Nonetheless, at the degrees of melting typical of MORB generation, nearly all of the CO₂, Rb, Ba and Nb from the mantle source are contained in the MORB melt, and thus CO₂/Nb is not a perfect canonical ratio at low degrees of melting. Observations from a global compilation on MORB glasses also suggest that Ba is the best proxy for C to model the CO₂ content of MORB and of the suboceanic mantle. Nonetheless, at the degrees of melting typical of MORB generation, nearly all of the CO₂, Rb, Ba and Nb from the mantle source are contained in the MORB melt, and thus CO₂/Nb is not a perfect canonical ratio at low degrees of melting. Observations from a global compilation on MORB glasses also suggest that Ba is the best proxy for C to model the CO₂ content of MORB and of the suboceanic mantle. Nonetheless, at the degrees of melting typical of MORB generation, nearly all of the CO₂, Rb, Ba and Nb from the mantle source are contained in the MORB melt, and thus CO₂/Nb is not a perfect canonical ratio at low degrees of melting.
degree of melting of 10%. The uncertainty on this estimate is conservative, as it takes into account the range in estimates of the Ba, Rb and Nb concentrations in DMM. This CO2 flux corresponds to an average 3He flux released from ridges of $802 \pm 316$ mol yr$^{-1}$, using a constant CO2/3He ratio for MORB of $2.2 \pm 0.7 \times 10^{10}$ (refs 8,9). The calculated CO2 flux is in very good agreement with previous estimates based on popping rock 2nD43 ($2.3 \times 10^{11}$ mol yr$^{-1}$ (ref. 2)), and is twice as high as estimated from Siqueiros melt inclusions ($9.3 \pm 2.8 \times 10^{10}$ mol yr$^{-1}$ (ref. 3)). Our CO2 flux is higher than the global estimates from vescularity ($6.5 \pm 1.8$ to $8.7 \pm 2.8 \times 10^{10}$ mol yr$^{-1}$ (ref. 3)), and is slightly lower than those from a global MORB glass compilation ($2.8 \pm 0.4 \times 10^{11}$ mol yr$^{-1}$ (ref. 13)).

Normalized by the total length of the ridge system (60,864 km (ref. 17)), this corresponds to an average CO2 flux of $2.9 \pm 1.1 \times 10^{11}$ mol yr$^{-1}$ km$^{-1}$. However, the CO2 flux from each segment of the global mid-ocean ridge system may vary as a function of magma flux and mantle source CO2 content. In the case of the ridge segment where sample EN061-5D-3 A was dredged, the local spreading rate of 3.26 mm yr$^{-1}$ (ref. 17) and crustal thickness of 5 km (ref. 14) translate to a local magma flux of 0.0168 km$^{3}$ yr$^{-1}$, a local CO2 flux of $1.4 \pm 0.6 \times 10^{10}$ mol yr$^{-1}$, and a local 3He flux of 0.64 $\pm 0.25$ mol yr$^{-1}$. Normalized by the length of the ridge segment (segment MAR179, 103 km long), the local CO2 flux corresponds to an average flux of $1.4 \pm 0.5 \times 10^{11}$ mol yr$^{-1}$ km$^{-1}$, which is half the average global flux. Thus, the differences observed between local and global CO2 fluxes illustrate geographical variations of at least a factor of two in the CO2 flux from ridges, controlled by variations in mantle carbon concentration and magma flux. This observation agrees with independent estimates based on MORB vescularity.

The equatorial MAR melt inclusions, an example of depleted MORB, have ratios of CO2/Ba, CO2/Rb and CO2/Nb very similar to the popping rock 2nD43, which is a highly enriched MORB (Fig. 3, Table 1). This observation demonstrates that these ratios are not simple functions of the amount of trace element enrichment or depletion in MORB; however, the limited range in these ratios, even taking Siqueiros into account, shows that absolute mantle CO2 abundances will scale with mantle Ba, Rb and Nb abundances. In order to successfully capture the global range of the upper mantle CO2 content, we use the global Rb, Ba and Nb variations from the literature, together with the CO2/Rb, CO2/Ba and CO2/Nb ratios from the equatorial MAR melt inclusions (Table 1). We selected estimates of Rb, Ba and Nb contents in both depleted MORB sources (D-DMM) and enriched MORB sources (E-DMM) (refs 17,32), and obtain a range in DMM CO2 content of $20 - 1,200$ p.p.m., equivalent to $5.5 - 327$ p.p.m. C. This range, reflecting the full spectrum of depleted-to-enriched MORB sources, is wider than previously reported and covers almost two orders of magnitude, including the low14,32 and the high2 mantle CO2 estimates. This range is in good agreement with the global range estimated independently from vescicle size distribution ($27 - 999$ p.p.m. CO2 (ref. 3)).

In a procedure similar to that for CO2, we provide estimates for the H2O, Cl and F content of the DMM. We do not apply this approach to S because the EN061-5D-3Ag are sulfide-saturated37; therefore, their S content is a direct function of the FeO content of the melt and cannot be linked to the S content of the mantle.
depth of metal saturation are uncertain, the depth of melt initiation is also uncertain. If the mantle is more reducing, melting will instead initiate at the carbonated silicate solidus, producing a melt fraction that is an order of magnitude below the threshold for carbonate melts, such as the source for Siqueiros MORB, would produce Fe3+/ΣFe bulk partition coefficients, and primitive mantle Fe2O3 contents89 and the assumption that 37 p.p.m. C has the power to reduce Fe3+/ΣFe by >1% as it oxidizes to carbonate. Because the subsolidus Fe3+/ΣFe ratio of the mantle and the depth of metal saturation are uncertain, the depth of melt initiation is also uncertain.

Our results on the CO2 content of the DMM have important implications for the geophysical detection of melt in the upper mantle and the origin of the asthenosphere. Provided that oxygen fugacity is high enough to stabilize carbonate11, carbonate melting will begin wherever the mantle temperature exceeds the carbonated mantle solidus, producing a melt fraction that is a function of the amount of CO2 in the mantle1. At the MORB mantle CO2 contents that we have constrained, this melt fraction will be vanishingly small. However, carbonate melts can become interconnected at very low melt fractions, as small as 0.05% (ref. 39), and thus there exists a melt fraction threshold below which carbonate melts cannot be extracted from the mantle and would not be detectable by geophysical methods. This threshold melt fraction can be used to define the effective base of the melt regime beneath ridges as well as the effective depth of melting. The seismic low-velocity zone beneath oceanic plates40 and the electrical conductivity structure of the upper mantle41 are both thought to be due to the presence of melt beneath the lithosphere. In particular, carbonate melts are highly conductive, much more so than hydrated mantle or silicate melts42. Using petrologic estimates for the reduction in melting temperature as a function of CO2 content1, a mantle potential temperature of 1,345 °C and an interconnection threshold of 0.05% in melt fraction39, we show that the regional variations in upper mantle CO2 that we have documented here predict large variations in the depth to the effective base of the melting regime beneath ridges, which should correlate with geochemical trace element signatures of depletion and enrichment in MORB (Fig. 5). An upper mantle source with a CO2 abundance of 20 p.p.m. would produce 0.005% melt, which is an order of magnitude below the threshold for interconnection39 and would, in areas of trace element depletion, predict an absence of melt and low electrical conductivity at depths deeper than the nominally anhydrous mantle solidus (85 km (ref. 1)). A less depleted MORB source with 70 p.p.m. CO2, such as the source for Siqueiros MORB, would produce enough carbonated silicate melt to establish an interconnected (and thus conductive) network of melt at depths of ~95 km and above. An average MORB source with 137 p.p.m. CO2 would produce an interconnected network of carbonated silicate melt throughout the entire upper mantle, limited at its base only by redox freezing where carbonate is converted to diamond3,38 (Fig. 5). Given sufficient depth resolution of electrical conductivity measurements, it may be possible to use...
geophysical measurements to determine the depth to the effective base of the carbonated melting regime in areas of trace element depletion, and the depth to the carbonate-diamond transition in areas of trace element enrichment. Given that the carbonate-diamond transition is dependent on oxygen fugacity13,14, variations in the depth of this redox boundary—if resolved—could be indicative of lateral variations in the oxygen fugacity of the mantle.

Methods

Sample preparation. We selected olivine grains that contained fully entrapped melt inclusions, with no cracks or links to the outside glass. We mounted the olivine grains in epoxy, polished them using SiC papers in order to expose the melt inclusions. After polishing, we removed the grains from epoxy using a soldering iron, pressed them into an indium mount, and polished the indium mounts using first diamond paste, then 1/3 μm alumina paste. We washed the mounts using alcohol and water, then stored them for >48 h in a vacuum oven at 70 °C, before applying a gold coat.

Volatile elements. We first analysed the volatile element compositions (H2O, CO2, F, Cl and S) of the matrix glass and the melt inclusions using the NanoSIMS Cameca 50L at the Department of Terrestrial Magnetism, Carnegie Institution, following the procedure described in ref. 43. We used a 12–13 nA Cs+ primary beam to preassure the sample using a 30 × 30 μm2 raster; then, performed the analysis using a 10 × 10 μm2 raster, and collected data on the central 3.8 × 3.8 μm2 of the crater, measuring 5 blocks of 10 analyses each. We assessed blank under these conditions by replicate measurements on Suprasil 3,002 glass and synthetic forsterite, which yield values of 0.2 p.p.m. CO2, 2 p.p.m. H2O and <0.05 p.p.m. F, Cl and S. Typical analytical error (2 s.d. over n analytical cycles) are <2% rel. for all volatiles. We assessed uncertainties by measuring basaltic glass ALV319−41 every 10 analyses, which yield uncertainties of 7% rel. for CO2 and <5% rel. for all other volatiles (2 s.d. over 50 analyses during a 4-day long analytical session). Supplementary Table 1 shows the long-term uncertainties as 2 s.d. external reproducibility of measured repetitions in four basaltic standards.

Major elements. After measuring volatile elements, we slightly polished the samples using 1/3 μm alumina powder in order to remove the gold coat, then we applied a carbon coat. We measured the major element compositions of the host olivines, the matrix glasses and the melt inclusions using a JEOL electron microprobe at the Geophysical Laboratory, Carnegie Institution, using the following conditions: 15 kV accelerating voltage, 30 nA beam, spot mode (for the host olivine, the matrix glasses and the melt inclusions) or beam defocused to 10 μm spot size. We normalized the data to 28Si as the internal standard. We used a set of 11 mafic glass standards (BIR-1g, BCR-2g, BHVO-2g, GSG-1g, GSG-5a, HS90-21g, OR132-g, GOR12-g, KL2-g and ML3F-g) to perform the calibration (linear regressions with $r^2 > 0.995$). We assessed uncertainties and analytical drift using repeated measurement of basaltic glass standard VE-32. Supplementary Table 3 shows the major element compositions of the olivines.

Trace elements. Finally, we analysed the concentrations of 40 trace elements in the melt inclusions and the matrix glasses by Laser-Induced Coupled Mass Spectrometry at the Department of Terrestrial Magnetism, Carnegie Institution, following the procedure described in ref. 44. We used a 35 mW laser (532 nm) to ablate target material. Analyses were carried out using a New Wave Research Nd:YAG laser coupled to an element mass spectrometry (ICP-MS). We used the following conditions: 193 nA beam current, 3 mm ablation spot, and 1 Hz laser frequency. We employed the following standards: BCR-2g and GSG-1g as the major standards, and GSG-5a as the minor standard.

Post-entrapment olivine crystallization. We assessed the major, volatile and trace element compositions of the melt inclusions for post-entrapment olivine crystallization (PEC). We used a Fe3+ / Fe2+ of 0.16 (average value measured in local MORB glasses27 identical to the global average MORB value28), together with the Fe-Mg partition coefficient between olivine and silicate melt47. We corrected the melts by adding olivine back to the melt using increments of 0.1%. All melt inclusions indicated an amount of PEC of <3% of olivine, with the exception of three melt inclusions that indicated PEC of 5, 7 and 9% of olivine. Supplementary Table 5 shows the PEC-corrected compositions of the melt inclusions and matrix glasses. Supplementary Table 6 shows the raw compositions of the melt inclusions and matrix glasses before PEC correction. Note that the main results of this study do not depend on the PEC correction, as both the use of the PEC-corrected compositions and the uncorrected compositions would yield similar conclusions.

Data availability. The authors declare that all data generated during this study are included in this published article (and its Supplementary Information files).

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Acknowledgements

The authors acknowledge the support of the Deep Carbon Observatory, Carnegie Institution of Washington and the Smithsonian Institution. We thank Jianhua Wang for the help with the NanoSIMS measurements, Tim Gooding for assistance with sample preparation and John Armstrong for the assistance with EMP measurements. NSF award OCE# 1258771 provides curatorial support for marine geological samples at the Graduate School of Oceanography, University of Rhode Island.

Author contributions

All authors contributed to the design of the study and the writing of the manuscript. K.A.K. obtained the sample from the URI sample collection. M.L.V. prepared the melt inclusions, performed the analyses and processed the data. E.H.H. and E.C. designed the experiment. K.A.K. obtained the sample from the URI sample collection. M.L.V. prepared the melt inclusions, performed the analyses and processed the data. E.H.H. and E.C. designed the experiment.

Additional information

Supplementary Information accompanies this paper at http://www.nature.com/naturecommunications

Competing financial interests: The authors declare no competing financial interests.

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How to cite this article: Le Voyer, M. et al. Heterogeneity in mantle carbon content from CO2-undersaturated basalts. *Nat. Commun.* 8, 14062 doi: 10.1038/ncomms14062 (2017).

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