

## The Fingerprint of Geologic Carbon on Glacial/Interglacial Marine Carbonate Chemistry and Atmospheric CO<sub>2</sub>

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The rise in atmospheric pCO<sub>2</sub> during the last glacial termination was accompanied by a 190‰ decrease in surface ocean Δ<sup>14</sup>C between 17 to 10 kyB.P., which cannot be explained without calling upon an input of <sup>14</sup>C-depleted carbon from either a formally isolated deep ocean reservoir or input of geologic carbon. There are now enough marine Δ<sup>14</sup>C records spanning the deglaciation to make clear that Pacific deep water Δ<sup>14</sup>C did not track the atmosphere whereas the Δ<sup>14</sup>C of intermediate waters did, except in the eastern equatorial Pacific (EEP) where there were large negative excursions in Δ<sup>14</sup>C. To explain the contrasting deep and upper ocean deglacial Δ<sup>14</sup>C histories a new hypothesis calls upon a release of <sup>14</sup>C-depleted CO<sub>2</sub>-rich hydrothermal fluids at intermediate water depths as the ocean warmed during the deglaciation. Here we use an Earth System model of intermediate complexity (cGENIE) with new geochemical records from the tropical Pacific to test this hypothesis. We show that a total release of 1400Gt of <sup>14</sup>C-dead DIC into intermediate waters of the EEP (700Gt) and Arabian Sea (700Gt) causes Δ<sup>14</sup>C changes throughout the ocean that agree with observations. The injection also causes a ~40 μmol/kg drop in [CO<sub>3</sub><sup>2-</sup>] of tropical surface waters. Trace metal proxies of [CO<sub>3</sub><sup>2-</sup>] and carbonate preservation data are presented for the EEP that document a transient decrease in [CO<sub>3</sub><sup>2-</sup>] in association with the deglacial Δ<sup>14</sup>C excursions. This carbonate preservation event is associated with elevated V/Ca, Zn/Ca and Cu/Ca in planktonic foraminifera from the EEP.

Taken together, the model and geochemical data provide strong support for the hypothesis that there was a release of geologic carbon during the deglaciation that contributed as much as 60ppm to the rise in atmospheric pCO<sub>2</sub>. If the geochemical signatures documented across glacial Termination 1 are found to occur at earlier glacial terminations as well, these findings will have profound implications for our understanding of glacial/interglacial CO<sub>2</sub> variability. Hydrothermal systems in the oceans may act as a CO<sub>2</sub> capacitor, regulating storage and release of carbon, and in doing so, affect the radiative balance that determines Earth's climate on orbital time scales.

## Inter-mineral Mg isotope fractionation in mantle xenoliths

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The bulk Mg isotope composition of the silicate Earth is homogeneous and identical within analytical error to that of chondritic meteorites [1-7]. Systematic fractionations among minerals in mantle peridotites [1,2,4,8-10] are resolvable, however, and are broadly consistent with theoretical predictions for high-T equilibrium Mg isotope fractionation [10]. Theory predicts that tetrahedral crystallographic sites have higher <sup>26</sup>Mg/<sup>24</sup>Mg than octahedral sites; up to 0.5 – 0.8‰ for spinel – silicate pairs at 1000K [10], which is in agreement with recent data from two spinel peridotites [9].

Here we investigate the inter-mineral Mg isotope fractionation in 5 spinel peridotite and 5 garnet pyroxenite xenoliths from Salt Lake Crater, Hawaii. Calculated whole rock δ<sup>26</sup>Mg compositions of both rock types are within error of those of bulk silicate Earth, suggesting that inter-mineral differences are due to equilibrium isotope fractionation. In the spinel peridotites, the δ<sup>26</sup>Mg of olivine (ol), orthopyroxene (opx) and clinopyroxene (cpx) are indistinguishable. In the three measured olivine-spinel pairs, the spinels (sp) have consistently higher δ<sup>26</sup>Mg values than the olivines, by 0.21 - 0.28‰. The sp-ol difference in δ<sup>26</sup>Mg suggests an equilibration temperature of ca. 1700°C, considerably higher than the calculated opx-cpx mineral equilibration temperature of ca. 1100°C. The δ<sup>26</sup>Mg of garnet in the garnet pyroxenites is consistently lower by 0.38 - 0.45‰ than in the coexisting cpx. Hence, the δ<sup>26</sup>Mg values increase from garnet with a Mg coordination number of 8, to the octahedrally coordinated silicates (ol ≤ opx ≤ cpx) and the tetrahedrally coordinated spinels, consistent with coordination number exerting a first-order control on inter-mineral high-T equilibrium Mg isotope fractionation [10].

However, a strong positive correlation of δ<sup>26</sup>Mg<sub>spinel</sub> with the spinel Cr (Al) content or spinel Cr# shows that composition also influences the δ<sup>26</sup>Mg values of individual minerals. The latter observation may explain the discrepancy between calculated equilibration temperatures based on mineral equilibria (opx-cpx) and Mg isotope fractionation (δ<sup>26</sup>Mg<sub>ol-sp</sub>), and may hamper the use of spinel-silicate Mg isotope fractionation as a reliable geothermometer in magmatic rocks.

- [1] Yang et al. (2009), *Earth Planet. Sci. Lett.* **288**, 475-482. [2] Handler et al. (2009) *Earth Planet. Sci. Lett.* **282**, 306-313. [3] Bourdon et al. (2010) *Geochim. Cosmochim. Acta* **74**, 5069-5083. [4] Chakrabarti & Jacobsen (2010) *Earth Planet. Sci. Lett.* **293**, 349-358. [5] Schiller et al. (2010) *Earth Planet. Sci. Lett.* **297**, 165-173. [6] Teng et al. (2010) *Geochim. Cosmochim. Acta* **74**, 4150-4166. [7] Pogge van Strandmann et al. (2011) *Geochim. Cosmochim. Acta* **75**, 5247-5268. [8] Wiechert & Halliday (2007) *Earth Planet. Sci. Lett.* **256**, 360-371. [9] Young et al. (2009) *Earth Planet. Sci. Lett.* **288**, 524-533. [10] Schauble (2011) *Geochim. Cosmochim. Acta* **75**, 844-869.