A deglacial ^{14}C budget

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Using the simplest possible model, an ocean split into upper (32%) and deep (63%) plus biosphere/atmosphere (5%), we construct a best estimate LGM ^{14}C budget based on ocean and atmospheric Δ^{14}C measurements. This budget is consistent with the estimates for the production-based global ^{14}C inventory during the LGM. Since the inferred ^{14}C ventilation ages for the mid-depth and deep ocean are significantly older than pre-industrial, some combination of sluggish LGM ocean overturning and greater surface reservoir ages is implicated. However, this budget leaves little room for a sizable, severely ^{14}C deplete Mystery Reservoir in the deep LGM ocean.

Given this preliminary LGM budget, it is very difficult, yet possible, to explain some of the deglacial observations by simple redistribution of ^{14}C from the upper ocean into the deep ocean. Specifically, the modest deglacial Δ^{14}C decline of southern-sourced mid-depth waters [1,2] can be accounted for by rising Δ^{14}C in the deep Southern Ocean [3]. However, for the Heinrich stadial 1, most records show constant or modestly decreasing planktic-benthic Δ^{14}C difference, suggesting a decline in the absolute Δ^{14}C of the deep ocean [4-6]. If these observations collectively reflect global deep ocean Δ^{14}C, then (a) an imbalance between deglacial production and decay of ^{14}C, (b) the addition of ^{14}C-dead geologic carbon, and/or (c) a substantial deepening of the thermocline are required to close the deglacial budget. These alternative mechanisms would also significantly contribute to the deglacial 190‰ decline in atmospheric Δ^{14}C.

[1] De Pol-Holz *et al.* (2010) *Nature Geosc.* doi:10.1038/NGEO745. [2] Rose *et al.* (2010) *Nature* doi:10.1038/nature09288. [3] Skinner *et al.* (2010) *Science* doi:10.1126/science.1183627. [4] Robinson *et al.* (2005) *Science* doi:10.1126/science.1114832. [5] Galbraith *et al.* (2007) *Nature* doi:10.1038/nature06227. [6] Broecker *et al.* (2008) *EPSL* doi:10.1016/j.epsl.2008.07.035.

The lithium isotopic signature of carbonatites

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Carbonatites are mantle-derived intraplate magmas that provide a means of monitoring the chemical secular evolution of the Earth's mantle [e.g., 1]. Speculations that mantle-derived rocks may record secular lithium (Li) isotopic variations as Li is progressively recycled into the mantle [2] have been fueled by the diverse Li isotopic composition of subducted sediments and basaltic ocean crust. The Li isotopic composition of carbonatites reflects their mantle source because there is no, or very little, Li isotope fractionation during their differentiation [3] and partial melting. A correlation is expected between recycled materials and Li isotopic anomalies, if a) non-mantle-like Li is introduced to the mantle through subduction [2], or b) Li isotopic heterogeneities are generated kinetically in the mantle by diffusion from subducted, Li-rich materials [4]. Li isotopic compositions of Archean to Recent carbonatites from several continents (δ^7Li = +4.1±1.3; n=23) overlap the range typical for modern mantle-derived rocks (MORB and OIB) and show no variation with time [5]. If the mantle sources of carbonatites are related to subduction and recycling of oceanic lithosphere, as suggested by some studies [e.g., 6], we see no evidence for this in terms of their Li isotopes, suggesting that neither of the two conditions above hold and that one or both of the following is true: a) the bulk composition of subducted material does not deviate greatly from the average mantle value (δ^7Li = +4±2), in agreement with a bulk δ^7Li of +3.1 for a typical subducting oceanic slab [7], b) crustal Li is effectively homogenized upon subduction into the mantle [5]. Alternatively, carbonatites may derive from primitive or depleted mantle sources, which have not been influenced by crustal recycling.

[1] Bell & Blenkinsop (1987) *GCA* **51**, 291-298. [2] Elliott *et al.* (2004) *EPSL* **220**, 231-245. [3] Halama *et al.* (2007) *EPSL* **254**, 77-89. [4] Vlastélic *et al.* (2009) *EPSL* **286**, 456-466. [5] Halama *et al.* (2008) *EPSL* **265**, 726-742. [6] Nelson *et al.* (1988) *GCA* **52**, 1-17. [7] Marschall *et al.* (2007) *EPSL* **262**, 563-580.