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Global-mean marine $\delta^{13}C$ and its uncertainty in a glacial state estimate

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ABSTRACT

A paleo-data compilation with 492 $\delta^{13}C$ and $\delta^{18}O$ observations provides the opportunity to better sample the Last Glacial Maximum (LGM) and infer its global properties, such as the mean $\delta^{13}C$ of dissolved inorganic carbon. Here, the paleo-compilation is used to reconstruct a steady-state water-mass distribution for the LGM, that in turn is used to map the data onto a 3D global grid. A global-mean marine $\delta^{13}C$ value and a self-consistent uncertainty estimate are derived using the framework of state estimation (i.e., combining a numerical model and observations). The LGM global-mean $\delta^{13}C$ is estimated to be 0.14‰ ± 0.20‰ at the two standard error level, giving a glacial-to-modern change of 0.32‰ ± 0.20‰. The magnitude of the error bar is attributed to the uncertain glacial ocean circulation and the lack of observational constraints in the Pacific, Indian, and Southern Oceans. To halve the error bar, roughly four times more observations are needed, although strategic sampling may reduce this number. If dynamical constraints can be used to better characterize the LGM circulation, the error bar can also be reduced to 0.05 to 0.1‰, emphasizing that knowledge of the circulation is vital to accurately map $\delta^{13}C$ in three dimensions.

1. Introduction

Carbon-13 to carbon-12 ratios (i.e., $\delta^{13}C$) can chemically fingerprint different carbon reservoirs, and thus glacial-interglacial changes in $\delta^{13}C$ of oceanic dissolved inorganic carbon (i.e., $\delta^{13}C_{DIC}$) reflect the carbon partitioning between terrestrial, atmospheric, and marine reservoirs. Dramatic environmental changes during the Last Glacial Maximum (LGM, 23,000 to 19,000 years before present) altered the terrestrial biosphere, and some of the low isotopic signature of terrestrial carbon ($\delta^{13}C \approx -25$‰) was transferred to the glacial ocean, consistent with observations of benthic foraminiferal $\delta^{13}C$ lower than the modern-day (e.g., Shackleton, 1977; Curry et al., 1988; Duplessy et al., 1988). The glacial atmosphere held approximately 170 gigatons (Gt) less carbon (e.g., Monnin et al., 2001), leaving the ocean as the most readily available source of compensation for the other two reservoirs. Pollen records and vegetation models that more directly reflect terrestrial carbon change yield higher estimates of glacial-to-modern carbon transfer (e.g., 750–1900 Gt C, Crowley, 1995; Adams and Faure, 1998; Kaplan et al., 2002) than the marine-based estimates (e.g., 330–650 Gt C, Shackleton, 1977; Curry et al., 1988; Duplessy et al., 1988; Köhler et al., 2010), although an inert terrestrial carbon pool may reconcile the difference (Ciais et al., 2012). A recent compilation of benthic *Cibicidoides* spp. $\delta^{13}C$ has nearly twice the data points of previous compilations and coverage of the Atlantic, Pacific, and Indian Oceans (Peterson et al., 2014), and thus motivates the re-investigation of the marine-based whole-ocean $\delta^{13}C$ estimates.

Determining the mean value of a spatially-distributed tracer field reduces to a linear operation in most cases (i.e., an inner vector product):

$$\bar{c} = \mathbf{w}^T\mathbf{y} + \bar{c}_0 = \sum_{i=1}^{N} w_i y_i + \bar{c}_0 \tag{1}$$

where $\bar{c}$ is the global mean value of a tracer c, $\mathbf{w}$ is a vector of weights with w_i for the ith element, ^{T} is the vector transpose, $\mathbf{y}$ is a vector containing N observations of y_i, and $\bar{c}_0$ is a constant included for full generality. If all observations are assumed to contain equal information about the global mean and no other information is available (i.e., $\bar{c}_0 = 0$), the optimal weights would all be $1/N$, and Equation (1) reverts to the basic sample mean. This assumption is invalidated if there are differing noise levels in the measurements.

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The sparse, irregularly-spaced nature of glacial observations also invalidates this assumption, of course. Originally, paleoceanographers best dealt with this issue by choosing cores from what was thought to be the most representative oceanic regions (e.g., Shackleton, 1977). As more data became available, basin-wide or regional means were computed as a preliminary step before global averaging (e.g. Curry et al., 1988; Boyle, 1992; Matsumoto and Lynch-Stieglitz, 1999; Peterson et al., 2014). This multi-step process naturally leads to non-uniform weights on the observations in Equation (1).

When the global-mean oceanic $\delta^{13}C_{DIC}$ is computed as a succession of sub-averages, the result may be sensitive to the size and location of the chosen sub-domains, and only by producing $\delta^{13}C_{DIC}$ maps at higher spatial resolution will this sensitivity be reduced. The distance between LGM observations, however, is often greater than the decorrelation lengthscale of oceanic property fields, and thus the typical method of "objectively" mapping the observations onto a regular grid (e.g., optimal interpolation or objective mapping, Bretherton et al., 1976) reverts to a first-guess estimate in many locations. In other words, large regions of the LGM ocean would be unconstrained by the data, especially at intermediate depths where little core coverage is available. Furthermore, the objectively-mapped estimate will leave local extrema in the estimated tracer field around the data points. Such features are undesirable because they are not physically sustainable in equilibrium when diffusion has sufficient time to act (e.g., for atmospheric momentum, Hide, 1969). It is not clear, however, how equilibrated the glacial ocean was and whether eddy processes can be accurately modeled as a diffusive process. Computation of an accurate global mean is challenging even for modern-day cases, such as sealevel rise (e.g., Wunsch et al., 2007). A new method is needed to create a map with sparse LGM observations that addresses these complications.

Here we suggest that a method originally developed for estimating the oceanic water-mass distribution from sparse observations (Gebbie, 2014) is also well-suited to make three-dimensional global maps. Specifically, we combine a tracer transport model (Section 2.1) with observations (Section 2.2) to produce an LGM state estimate. Rather than using the assumed statistics of circulation lengthscales, like optimal interpolation, we illustrate that the circulation itself can be used to make a gridded field (Section 2.3). The numerical model serves a dual purpose: 1) a means to readily interpret the sources, sinks, and pathways of tracer, and 2) a kinematic interpolator and extrapolator that allows large-scale information to be extracted from the observations. Here we extend the state estimation framework by deriving a self-consistent formula for the global-mean uncertainty (Section 2.4).

This work has two major results: 1) an estimate of the LGM global-mean $\delta^{13}C_{DIC}$, and 2) its uncertainty within a explicit set of assumptions. To connect these results to deglacial climate dynamics and the carbon cycle, we reconstruct a global map of LGM $\delta^{13}C_{DIC}$ and detect a largescale, coherent pattern of LGM-to-modern changes (Section 3). The glacial-mean $\delta^{13}C_{DIC}$ uncertainty is partially attributed to the sparsity and measurement error in the observations, but also due to the difficulty in accurately modeling the LGM circulation (Section 4). Our results are discussed in the context of previous observational techniques (Section 5), especially how the observational weights in the averaging Equation (1) are modified by the assumed circulation regime. We conclude by emphasizing the importance of circulation knowledge in the goal of further reducing the global-mean $\delta^{13}C_{DIC}$ uncertainty (Section 6).

2. Global LGM state estimate

The global LGM state estimate is produced by combining a kinematic tracer transport model with a global array of benthic foraminiferal observations of $\delta^{13}C$ and $\delta^{18}O$. Global, three-dimensional gridded distributions are produced for multiple tracers: $\delta^{13}C_{DIC}$, seawater $\delta^{18}O$ (i.e., $\delta^{18}O_w$), potential temperature, practical salinity, and phosphate. The model, observations, and state estimation method are detailed next.

2.1. Model

The model is a statistically steady-state conservation equation that is assumed to hold for, C, a general tracer: $\nabla\cdot(\vec{F}C) = Q$, where $\vec{F}$ is the mass flux and Q is a local source or sink. In the statistical steady state, any temporal variability that has a net diffusive or advective effect is represented by the model used here.

In practice, the model equations are discretized on a global, three-dimensional grid. Here the grid is defined with $4^\circ \times 4^\circ$ horizontal resolution and 33 vertical levels with enhanced resolution near the surface. Glacial ocean computations are undertaken on the same grid as a modern-day reference case, but gridcells shallower than 120 m modern-day water depth are discarded due to the sealevel drop. After discretization, the equations are normalized by the sum of all mass fluxes into the gridcell, $\phi_i = \sum_{j=1}^{N} f_{ij}$, where f_{ij} is the flux from gridcell j to i, and there are N neighboring gridcells. Then the tracer transport equation at gridcell i becomes more similar to a water-mass mixing model (following Gebbie and Huybers, 2012):

$$\sum_{j=1}^{N} m_{ij} c_j - c_i = q_i \quad (2)$$

where m_{ij} is the ratio of the inward flux from j to the total flux ($m_{ij} = f_{ij}/\phi_i$), c_i is the tracer concentration in cell i, and q_i is the equilibrium tracer source with units of the tracer concentration itself ($q_i = Q/\phi_i$). For conservative tracers, the source and sink vanishes ($Q = 0$). These algebraic manipulations lead to a well-conditioned set of equations that can be solved quickly, but with the tradeoff that information is lost regarding the absolute rate of circulation.

The isotope variables, $\delta^{13}C_{DIC}$ and $\delta^{18}O_w$, require some further consideration. In particular, the sink of $\delta^{13}C_{DIC}$ due to remineralization is assumed to be equal to -0.95‰/(μmol/kg) times the source of remineralized phosphate, which is adjusted relative to the modern ratio of -1.1‰/(μmol/kg) due to changes in whole-ocean $\delta^{13}C_{DIC}$ and upper-ocean biological fractionation (e.g., following Broecker and Maier-Reimer, 1992). Here we model the ratio (delta value) rather than the individual isotopes which incurs an error (e.g., Walker, 1991), but it is small because the $^{18}O/^{16}O$ ratio in Vienna Standard Mean Ocean Water (VSMOW) standard is about 1/500, and the $^{13}C/^{12}C$ ratio in the Vienna Pee Dee Belemnite (VPDB) standard is about 1/90. Furthermore, this error is damped in the vicinity of observations by the formal data constraints.

For reasons that should become clear below, the state vector, $\mathbf{x}$, is defined to contain both tracer and circulation information, i.e., $\mathbf{x}^T = [\mathbf{c};\mathbf{m}]^T$, where $\mathbf{c}$ is a vector that represents all of the global three-dimensional tracer distributions and $\mathbf{m}$ describes the circulation by concatenating all of the mass-flux ratios, m_{ij} (e.g., Gebbie and Huybers, 2010). This state vector definition is not unique, but it provides sufficient information to permit a steady-state tracer distribution to be computed, and thus is an acceptable definition of the state. All of the tracer transport equations are combined and symbolically represented as: $\mathscr{L}[\mathbf{x}] = \mathbf{q} + \mathbf{v}$, where $\mathscr{L}$ is a nonlinear operator due to the multiplication of the tracer concentration and flow field that encapsulates advective and diffusive processes, and $\mathbf{v}$ is the source deviation from the modern-day first-guess field, $\mathbf{q}$. The model equation includes surface concentration (i.e., Dirichlet) boundary conditions for completeness.