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LGM hosing approach to Heinrich Event 1: results and perspectives from data–model integration using water isotopes

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A B S T R A C T

Freshwater hosing has long been used in climate models to represent the perturbation associated with the release of icebergs during Heinrich Events. However, freshwater hosing ignores some of the icebergs-induced feedbacks on climate. It therefore bears the question whether the perturbed climate obtained through classical freshwater hosing experiments is realistic enough to be used as a proxy for a Heinrich Event.

Using the fully coupled, water isotope enabled, climate model *i*LOVECLIM, we conducted a series of freshwater hosing experiments, accounting for the depleted signature of ice-sheet sourced water, under LGM conditions. Using paleoceanographic data, we integrated simulated and measured $\delta^{18}O_{calcite}$ at the surface and within the ocean column to evaluate whether (i) LGM freshwater hosing is a good proxy for Heinrich Event 1 and whether (ii) different sources for the freshwater in the North Atlantic result in different spatial anomalies that could be used to fingerprint the source of melt water.

Results obtained with two sets of experiments indicate that first order simulated $\delta^{18}O_{calcite}$ anomalies are broadly consistent with the measured anomalies and that a complete shutdown of the thermohaline circulation is not consistent with the reconstructed patterns from foraminiferal calcite. A more detailed statistical analysis of the planktic and benthic $\delta^{18}O_{calcite}$ anomalies shows that the best experiment is obtained for a severely disturbed thermohaline circulation but not a complete shutdown. With our model sensitivity, this corresponds to a freshwater flux between 0.18 Sv and 0.3 Sv. Our results also indicate a better agreement when the freshwater flux is applied in the Labrador Sea than when directly applied to the Ruddiman Belt.

1. Introduction

Current human-induced climate warming is likely to trigger at least a partial deglaciation of the Greenland and western Antarctic in the near future (e.g. Huybrechts et al., 2004). Within this context, several model studies (e.g. Manabe and Stouffer, 1995; Ganopolski and Rahmstorf, 2001; Hewitt et al., 2006) have shown that the additional meltwater input to the ocean might trigger a slowdown of the Atlantic Meridional Overturning Circulation (AMOC) thereby providing a climate feedback. However, the AMOC response to freshwater input is known to be both model dependent (e.g. Stouffer et al., 2006; Kageyama et al., 2010; Otto-Bliesner and Brady, 2010; Kageyama et al., 2013) and climate dependent (e.g. Ganopolski and Rahmstorf, 2001; Kageyama et al., 2010; Van Meerbeeck et al., 2011), reducing our capability of predicting the risk of such an event.

Turning to paleoclimate evidence, reconstructing variations in the oceanic circulation is notoriously difficult since there is no proxy that records those changes only. Proxies that are responding to oceanic circulation changes are also sensitive to other factors such as biological productivity (benthic foraminifer $\delta^{13}C$) or particle fluxes (Pa/Th) to cite a few (Lynch-Stieglitz et al., 1995; Geibert and Usbeck, 2004). When measured over the last deglaciation (the last 21,000 years) both Pa/Th (McManus et al., 2004; Gherardi et al., 2005, 2009) and benthic $\delta^{13}C$ indicate that there is a strong reduction of deep oceanic circulation and/or ventilation during Heinrich Event 1, a major iceberg surge at the beginning of the deglaciation (~17–15 ka BP). Pa/Th and benthic $\delta^{13}C$ values during Heinrich Event 1 are compatible with a near-complete shutdown of the thermohaline circulation in the North Atlantic below ~3000 m depth, while they indicate that there was still a vigorous export of North Atlantic waters to the South Atlantic above ~2000 m depth

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(Waelbroeck et al., 2006, 2011; Gherardi et al., 2009; Thornalley et al., 2010).

It is a usual approach to mimic the climatic evolution over Heinrich Events by applying freshwater flux on a constant glacial climate background. The input location used for the freshwater bears generally little resemblance to the region in which any Heinrich Event induced freshwater forcing may end up. Since in general, the purpose is to assess the climatic impact of an AMOC shutdown, the amount of freshwater flux imposed is large enough to trigger such a shutdown in the given model and does not always bear a relationship to the estimated volume flux of icebergs of the Heinrich Event. So far, few studies have tried to assess the amount of freshwater released during a Heinrich Event.

To our knowledge, only one previous study tried to invert, in a 2-D oceanic model, the $\delta^{18}O_{calcite}$ signal from planktic foraminifera to evaluate both the amount of freshwater that entered the North Atlantic Ocean and the necessity of an AMOC shutdown (Roche et al., 2004). In this study of Heinrich Event 4 (~40 ka BP), the freshwater flux in best accordance with proxy data was found to be of magnitude 0.3 ± 0.1 Sv (1 Sv = 10^6 m^3s^{-1}) and of duration 250 ± 150 years. In addition, this study indicated that an AMOC shutdown was necessary for that model to simulate surface $\delta^{18}O_{calcite}$ anomalies that agree with planktic foraminifera proxy data records.

While freshwater hosing is the most common method employed to simulate circulation perturbations that are associated with Heinrich Events, it neglects the effect of latent heat consumption from the ocean to melt the icebergs and the fact that icebergs are a mobile source of freshwater, affected by winds and oceanic currents. Investigation of such effects in coupled climate models where icebergs are included showed that: 1) a release of icebergs in the Labrador Sea leads to iceberg trajectories in accordance with reconstructed Ice Rafted Debris spatial distribution (Ruddiman, 1977; Hemming, 2004; Levine and Bigg, 2008; Jongma et al., 2013) 2) the latent heat effect is important in that it facilitates sea-ice expansion (Jongma et al., 2013) 3) the associated freshwater release occurs primarily in the western Atlantic (Jongma et al., 2013; Roberts et al., 2014).

In the present manuscript, we take Heinrich Event 1 as a case study for freshwater hosing under glacial climate conditions using for the first time an oxygen water isotope enabled coupled climate model including a 3-D oceanic model. Our aim is to provide preliminary answers to the following questions: can freshwater hosing under glacial climate conditions be used as a good representation of deglacial Heinrich Event 1? Can we evaluate, using different freshwater inputs, the magnitude of AMOC reduction that is most consistent with proxy data records?

2. Model description

We use the *i*LOVECLIM coupled climate model version 1.0, including the simulation of water isotopes in the atmosphere, ocean and continental surface (Roche, 2013). It is a code fork and an evolution of the LOVECLIM model in its version 1.2 (Goosse et al., 2010), from which it retains the Atmosphere, Ocean and Vegetation (AOV) components. In the following we provide a brief description of the AOV components taken from Roche et al. (2013).

The atmospheric component ECBilt was developed at the Dutch Royal Meteorological Institute (KNMI) (Opsteegh et al., 1998). Its dynamical core is based on the quasi-geostrophic approximation with additional ageostrophic terms added to improve the representation of the Hadley cell dynamics. It is run on a spectral grid with a T21 truncation (5.6° in latitude/longitude in the physical space). ECBilt has three vertical layers at 800, 500 and 200 hPa. Only the first layer contains humidity as a prognostic variable. The time step of integration of ECBilt is 4 h.

The oceanic component (CLIO) is a 3-D oceanic general circulation model (Goosse and Fichefet, 1999) based on the Navier–-Stokes equations. It is discretized on an Arakawa B-grid at approximately 3° × 3° resolution. The vertical discretization follows a “z-coordinate” on 20 levels. It has a free surface that allows the use of real freshwater fluxes, a parameterisation of down-sloping currents (Campin and Goosse, 1999) and a realistic bathymetry. CLIO includes a dynamical-thermodynamical sea-ice component that is an updated version of Fichefet and Morales Maqueda (1997, 1999).

The dynamic vegetation model (VECODE) was specifically designed for long-term computation and coupling to coarse resolution models (Brovkin et al., 1997). VECODE consists of three submodels: (1) a model of vegetation structure (bioclimatic classification) calculates plant functional type (PFT) fractions in equilibrium with climate; (2) a biogeochemical model computes net primary productivity (NPP), allocation of NPP, and carbon pool dynamics (leaves, trunks, soil carbon pools) (3) a vegetation dynamics model. The latter computes two Plant Functionnal Types (PFT: trees & grass) and a dummy type (bare soil).The vegetation model is resolved on the atmospheric grid (hence at T21 resolution) and allows fractional allocation of PFTs in the same grid cell to account for the small spatial scale needed by vegetation.

Water isotopes have been implemented in the atmosphere following exactly what is done for the moisture (Roche, 2013) to ensure consistency between freshwater fluxes and isotopic fluxes. In the ocean, water isotopes are treated as passive tracers. Under pre-industrial conditions, $\delta^{18}O$ in precipitation and in oceanic waters of *i*LOVECLIM was found to reproduce the main features observed in present-day measurements (Roche and Caley, 2013). To further ascertain the capability of the model to reproduce what is measured in proxy data, it was compared to a set of proxy data from the Late Holocene (Caley and Roche, 2013). A very good agreement was found between oceanic $\delta^{18}O_{calcite}$ proxy data and simulated $\delta^{18}O_{calcite}$, of prime importance for the present study.

3. Rationale for using oxygen isotopes in calcite

The carbonate oxygen isotopic concentration from various organisms such as foraminifers is mainly controlled by temperature and by the isotopic composition of seawater ($\delta^{18}O_{sw}$) during shell formation (Urey, 1947; Shackleton, 1974).

The temperature dependence of the equilibrium fractionation of inorganic calcite precipitation around 16.9 °C is given in Shackleton (1974) as:

$$T = 16.9 - 4.38\left(\delta^{18}O_c(\mathrm{PDB}) - \delta^{18}O_{sw}(\mathrm{SMOW})\right) + 0.1\left(\delta^{18}O_c(\mathrm{PDB}) - \delta^{18}O_{sw}(\mathrm{SMOW})\right)^2 \quad (1)$$

where $\delta^{18}O_c$(PDB) refers to the measured isotopic value in the given calcite sample relative to the PDB (Pee Dee Belemnite) standard for carbonate (Craig, 1957) and $\delta^{18}O_{sw}$(SMOW) refers to the isotopic value of seawater in which the carbonate was formed, relative to the SMOW (Standard Mean Ocean Water) standard for water (Baertschi, 1976).

We use Shackleton (1974) equation to compute the modelled $\delta^{18}O_c$ from modelled temperature and modelled $\delta^{18}O_{sw}$ as follow:

$$\delta^{18}O_c(\mathrm{PDB}) = 21.9 - 0.27 + \delta^{18}O_{sw}(\mathrm{SMOW}) - \sqrt{310.61 + 10{*}T} \quad (2)$$