Earth and Planetary Science Letters 495 (2018) 157–163

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

www.elsevier.com/locate/epsl

Wave inhibition by sea ice enables trans-Atlantic ice rafting of debris during Heinrich events

Till J.W. Wagner [a,b,*], Rebecca W. Dell [b], Ian Eisenman [b], Ralph F. Keeling [b], Laurie Padman [c], Jeffrey P. Severinghaus [b]

[a] University of North Carolina Wilmington, United States of America
[b] Scripps Institution of Oceanography, University of California San Diego, United States of America
[c] Earth and Space Research, Corvallis, OR, United States of America

ARTICLE INFO

Article history:
Received 6 July 2017
Received in revised form 1 May 2018
Accepted 4 May 2018
Available online xxxx
Editor: M. Frank

Keywords:
icebergs
Heinrich events
ice-rafted debris
melt water
modeling

ABSTRACT

The last glacial period was punctuated by episodes of massive iceberg calving from the Laurentide Ice Sheet, called Heinrich events, which are identified by layers of ice-rafted debris (IRD) in ocean sediment cores from the North Atlantic. The thickness of these IRD layers declines more gradually with distance from the iceberg sources than would be expected based on present-day iceberg drift and decay. Here we model icebergs as passive Lagrangian particles driven by ocean currents, winds, and sea surface temperatures. The icebergs are released in a comprehensive climate model simulation of the last glacial maximum (LGM), as well as a simulation of the modern climate. The two simulated climates result in qualitatively similar distributions of iceberg meltwater and hence debris, with the colder temperatures of the LGM having only a relatively small effect on meltwater spread. In both scenarios, meltwater flux falls off rapidly with zonal distance from the source, in contrast with the more uniform spread of IRD in sediment cores. To address this discrepancy, we propose a physical mechanism that could have prolonged the lifetime of icebergs during Heinrich events. The mechanism involves a surface layer of cold and fresh meltwater formed from, and retained around, large densely packed armadas of icebergs. This leads to wintertime sea ice formation even in relatively low latitudes. The sea ice in turn shields the icebergs from wave erosion, which is the main source of iceberg ablation. We find that sea ice could plausibly have formed around the icebergs during four months each winter. Allowing for four months of sea ice in the model results in a simulated IRD distribution which approximately agrees with the distribution of IRD in sediment cores.

1. Introduction

Layers of sand found in ocean sediment cores throughout much of the North Atlantic indicate several widespread events during the last glacial period. The sand in these layers is too coarse to have been carried by winds or currents, and it is generally believed that this sand was rafted by icebergs during episodes of massive calving from the Laurentide Ice Sheet, called Heinrich events (Heinrich, 1988; Broecker, 1994; Hemming, 2004; Rhodes et al., 2015). These ice-rafted debris (IRD) layers are particularly pronounced in the latitude range 40°N–55°N, which is sometimes referred to as the "IRD belt" (Fig. 1). Large volumes of freshwater rich with debris are expected to have been released from icebergs to produce the observed IRD layers (Dowdeswell et al., 1995; Hemming, 2004; Levine and Bigg, 2008; Roberts et al., 2014), with estimated ice discharges up to 100 times greater than that from present-day Greenland (Hemming, 2004). Such freshwater fluxes have been found in models to cause the Atlantic Meridional Overturning Circulation (AMOC) to weaken, which leads to reduced poleward heat transport and regional cooling (Broecker et al., 1985; Manabe and Stouffer, 1997; Levine and Bigg, 2008; Otto-Bliesner and Brady, 2010). Hence investigating the distribution of meltwater and IRD during Heinrich events could inform projections of future climate change in scenarios involving substantial discharges of icebergs from the Greenland Ice Sheet.

The thickness of IRD layers provides an indication of iceberg meltwater release and drift tracks during the Heinrich events. Recent studies have investigated this in coarse-resolution climate models, and they found a persistent mismatch between modeling results and IRD thickness in sediment cores (Jongma et al., 2013; Roberts et al., 2014). Specifically, the models tend to simulate a rapid decline of meltwater input from west to east across

* Corresponding author at: University of North Carolina Wilmington, USA.
E-mail address: wagnert@uncw.edu (T.J.W. Wagner).

https://doi.org/10.1016/j.epsl.2018.05.006

the North Atlantic which resembles the distribution of modern iceberg sightings (International Ice Patrol, 2009), implying a similar decline in IRD layer thickness (Jongma et al., 2013; Roberts et al., 2014). Ocean sediment cores, by contrast, show a more gradual decrease from west to east (Hemming, 2004). Here we investigate this mismatch with a Lagrangian iceberg drift and decay model forced by output from a comprehensive global climate model (GCM) simulation. In contrast to previous studies, we use a higher resolution ($\sim 1°$) GCM, but the icebergs we simulate are non-interactive, behaving as passive tracers in the climate system.

2. Model setup and simulations

We use a representation of iceberg drift which evolves iceberg velocity, $\vec{v}_i$, subject to wind and ocean current drag, the pressure gradient force, and the Coriolis force (Wagner et al., 2017a). Compared to previous iceberg models (Bigg et al., 1997; Jongma et al., 2009; Martin and Adcroft, 2010; Marsh et al., 2015), this formulation is somewhat idealized, with the main approximations being that (i) the pressure gradient force is derived from the ocean velocity field by assuming geostrophy, (ii) iceberg speed is taken to be much smaller than surface wind speed, (iii) drag from sea ice and wave radiation are neglected, (iv) water drag is computed from the surface current alone (ignoring vertical shear), and (v) the forces on the iceberg are taken to be balanced (neglecting acceleration). This allows for an analytical solution for iceberg velocity in terms of surface air velocity, $\vec{v}_a$, and surface water velocity, $\vec{v}_w$. The solution can be written (Wagner et al., 2017a) as

$$\vec{v}_i = \vec{v}_w + \gamma \left(-\alpha \hat{k} \times \vec{v}_a + \beta \vec{v}_a \right). \tag{1}$$

Here, $\hat{k}$ is the vertical unit vector. The parameter γ is a measure of the relative strength of the air and water drags, and it depends on the densities of ice, water, and air, as well as the air and water drag coefficients. The coefficients α and β depend on wind speed, iceberg size, and the Coriolis parameter. Equation (1) implies that icebergs drift at an angle $\theta = \tan^{-1}(\alpha/\beta)$, relative to the water velocity, with θ depending primarily on wind strength and iceberg size. This solution (1) enables us to efficiently compute large numbers of non-interactive iceberg trajectories from precomputed surface wind and ocean current fields. More details regarding the derivation of equation (1) and the approximations listed above, as well as expressions for α, β, and γ, are given by Wagner et al. (2017a).

We include a representation of iceberg decay that accounts for three main decay processes (Bigg et al., 1997; Martin and Adcroft, 2010; Wagner et al., 2017a): (i) wind-driven wave erosion, M_e, (ii) turbulent basal melt, M_b, and (iii) side wall erosion from buoyant convection, M_v. Iceberg length, L, width, W, and height, H, evolve according to $dL/dt = dW/dt = M_e + M_v$ and $dH/dt = M_b$, with iceberg volume given by $V = LWH$. The individual decay terms are written as follows:

$$\begin{aligned} M_e &= \tfrac{1}{12}\left(1+\cos[\pi A_i^3]\right)(T_w - T_0)\, S(\vec{v}_a, \vec{v}_w), \\ M_v &= b_1 T_w + b_2 T_w^2, \\ M_b &= c\left|\vec{v}_w - \vec{v}_i\right|^{0.8}(T_w - T_i)L^{-0.2}, \end{aligned} \tag{2}$$

where A_i is the fractional sea ice concentration, T_w is the sea surface temperature (SST), $T_0 = -2\,°\mathrm{C}$, S is the sea state, $b_1 = 8.8 \times 10^{-8}\ \mathrm{m\,s^{-1}\,°C^{-1}}$, $b_2 = 1.5 \times 10^{-8}\ \mathrm{m\,s^{-1}\,°C^{-2}}$, $c = 6.7 \times 10^{-6}\ \mathrm{m^{0.4}\,s^{-0.2}\,°C^{-1}}$, and T_i is the temperature of the iceberg which is fixed at $-4\,°\mathrm{C}$. The sea state, S, is computed using a fit to the Beaufort Scale. Finally, we include iceberg capsizing using the stability criterion of Wagner et al. (2017b), which corrects errors in the original criterion of Bigg et al. (1997). For our primary set of simulations we approximate that there is no sea ice around the icebergs ($A_i = 0$) since the icebergs occur mainly in locations that do not have sea ice in the GCM simulations. We subsequently examine the role of a local seasonal sea ice cover around the icebergs, in which case we approximate that the sea ice moves with the icebergs and does not impact the iceberg drift. This approximation would break down in situations where the drift of icebergs is influenced substantially by the presence of thick pack ice covering large regions (e.g., in modern-day conditions north of Greenland or in the Weddell Sea).

The iceberg model described above requires three input fields which we take from GCM simulations: $\vec{v}_w$, $\vec{v}_a$, and SST. We use output from the Community Climate System Model version 4 (CCSM4), a coupled GCM developed by the National Center for Atmospheric Research, which is run at a nominal 1° horizontal resolution (Gent et al., 2011). We force the iceberg model with two previously run CCSM4 simulations: (i) a simulation of the 20th century with historical forcing that spans the period 1850–2005, which is part of the Coupled Model Intercomparison Project phase 5 (CMIP5), henceforth referred to as “20C”, and (ii) a simulation of the last glacial maximum (LGM) with ice sheets, coastlines, greenhouse gases, and solar forcing specified based on paleoclimate estimates (Brady et al., 2013), henceforth referred to as “LGM”.

For each simulation, we consider surface conditions over a 14-yr period. For 20C this period spans the years 1992–2005. For LGM, this period comprises the final 14 yrs of the 1000-yr simulation. The iceberg model is forced with these time-varying climate fields. The analysis of the model output is focused on long-term mean iceberg decay. This is done by computing the iceberg freshwater flux averaged over the 14-yr study period, which removes much of the model’s internal variability. The scale of the additional contributions from longer-term internal climate variability can be assessed by considering the spread over the 6 available CCSM4 historical model realizations. We find this to be relatively small. For example, the spread among model realizations in the 14-yr mean zonal-mean surface wind is $0.2\ \mathrm{m\,s^{-1}}$, which is much smaller than the latitudinal variation of the zonal-mean surface wind (Wagner and Eisenman, 2017, their Fig. S1).

Mean climate conditions for 20C and LGM are shown in Fig. 1. The continental shelf waters off the Labrador Coast feature the strong southward Labrador Current in the 20C simulation. During the LGM, the continental shelf was mostly above sea level and no significant western boundary current is simulated off the LGM Labrador Coast. The eastward flowing North Atlantic Current is notably stronger east of $\sim 40°$W in the LGM simulation. The wind fields show a generally stronger circulation in the LGM case, with particularly strong northwesterlies over the Labrador Coast and elevated wind speeds in the central Atlantic. This is in agreement with previous estimates of enhanced winds during the LGM, a feature that has been attributed to larger atmospheric temperature gradients (McGee et al., 2010). As expected, SSTs are overall colder in the simulated LGM climate. The spatially-averaged cooling in the northern North Atlantic (35°N–65°N) is 4.8 °C. This is somewhat higher than the extra-tropical northern hemisphere average cooling of 3.7 °C computed from the full 1000-yr LGM run (Brady et al., 2013). These values compare to an Atlantic-mean SST cooling of 2.8 °C obtained using a recent LGM state estimate (Kurahashi-Nakamura et al., 2017).

We consider 10 iceberg size classes with initial dimensions ranging from $100 \times 67 \times 67$ m to $1500 \times 1000 \times 300$ m (see Table S1 and Wagner et al., 2017b). A total of 25×10^3 icebergs are released (2500 for each size class) at a constant rate of 1 iceberg of each size class every 2 days throughout the simulations, and each