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Modeling the surface mass-balance response of the Laurentide Ice Sheet to Bølling warming and its contribution to Meltwater Pulse 1A

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ABSTRACT

Meltwater Pulse (MWP) 1A occurred ~14.5–14 ka and is the largest abrupt rise in sea level (10–20 m of sea-level rise) of the last deglaciation. The timing of MWP-1A is coincident with or shortly follows the abrupt warming of the North Atlantic region into the Bølling warm period, which could have triggered a large Laurentide Ice Sheet (LIS) contribution to MWP-1A. Given that outside of the Arctic, LIS iceberg discharge probably did not increase during the Bølling, much of the LIS MWP-1A contribution likely occurred through surface ablation. Here we test the response of LIS surface mass-balance to Bølling warming by forcing a LIS energy–mass balance model with climate from an atmosphere–ocean general circulation model. Our modeling approach neglects changes in LIS mass from dynamics and iceberg calving, allowing us to isolate the surface mass balance response. Model results suggest that LIS surface ablation can explain much of the sea-level rise just prior to MWP-1A. LIS surface mass-balance becomes more negative in response to the Bølling warming, contributing an additional 2.9 ± 1.0 m of sea-level rise in 500 yr in addition to the background contribution of 4.0 ± 0.8 m. The modeled LIS MWP-1A contribution is less than previous assumptions but agrees with geochemical runoff and LIS area-volume estimates. The fraction of MWP-1A attributable to other ice sheets, particularly Antarctica, depends on the total sea-level rise that occurred during this MWP.

1. Introduction

During the last deglaciation ~20–6 ka, sea level rose 125–130 m at an average rate of ~1 cm yr^{-1} (Bard et al., 1990; Fairbanks, 1989). The rate of sea-level rise was not constant with several intervals of more rapid rise, the most notable of which is Meltwater Pulse (MWP) 1A (Fairbanks, 1989; Hanebuth et al., 2000; Peltier and Fairbanks, 2006). During MWP-1A, sea level rose 10–20 m in <500 yrs, indicating rapid mass loss from one or more ice sheets to the global oceans (Fig. 1C). The timing of MWP-1A is roughly coincident with the abrupt warming of the North Atlantic region into the Bølling ~14.6 ka (Fig. 1A). The precise relationship is debatable, however, with the onset of MWP-1A ranging from ~14.6 to 14.3 ka and termination between ~14.6 and 13.8 ka (Fig. 1C) (Bard et al., 1990; Edwards et al., 1993; Hanebuth et al., 2000; Peltier and Fairbanks, 2006; Stanford et al., 2006; Weaver et al., 2003). Thus MWP-1A could be coincident with or lag the onset of the Bølling warm period, with implications for the sources and climatic impacts of this MWP (Stanford et al., 2006; Tarasov and Peltier, 2005; Weaver et al., 2003).

The Laurentide Ice Sheet (LIS) was originally assumed to be the dominant source of MWP-1A (e.g., Fairbanks, 1989; Peltier, 1994) with subsequent work suggesting additional contributions from the Scandinavian, Barents–Kara and Cordilleran Ice Sheets (Peltier, 2004; Tarasov and Peltier, 2005). A large LIS contribution could be explained as a response to the abrupt Bølling warming with the majority of the meltwater delivered to the ocean through increased surface ablation of its southern margin (Peltier, 2004; Tarasov and Peltier, 2005). In as much as increased deposition of iceberg rafted debris (IRD) sourced from the LIS reflects increased iceberg calving, IRD records suggest that eastern LIS iceberg discharge did not significantly increase during the Bølling (Fig. 1B) (e.g., Andrews and Tedesco, 1992; Bond et al., 1999; Hemming, 2004; Keigwin and Jones, 1995). IRD deposition in the Arctic Ocean, however, increased and may record the onset of northern LIS retreat and greater iceberg calving during the Bølling (Darby et al., 2002; Dyke, 2004; England et al., 2009). Here we investigate the LIS surface mass-balance response-alone to Bølling warming by forcing an energy–mass balance model (EMBM) with climate fields from an atmosphere–ocean general circulation model

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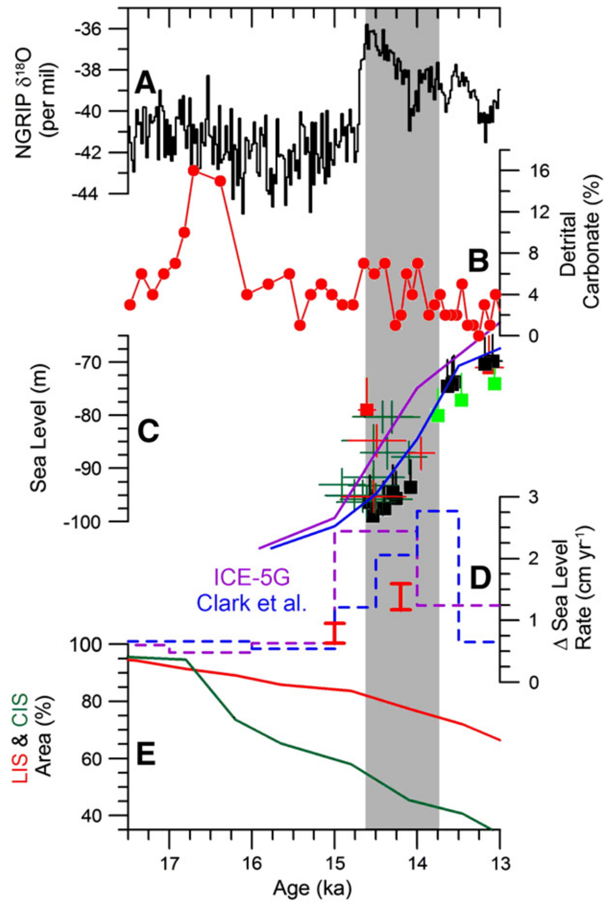

Fig. 1. Climate and sea level. (A) Greenland $\delta^{18}O$ (Svensson et al., 2008). (B) North Atlantic % detrital carbonate from core V23-81 (Bond et al., 1999). (C) Eustatic and relative sea level. Purple line is ICE-5G eustatic (Peltier, 2004); blue line is eustatic sea level from Clark et al. (2009). Squares are coral sea-level data (red from New Guinea, light green from Tahiti, black from Barbados) (Bard et al., 1996; Edwards et al., 1993; Peltier and Fairbanks, 2006). Crosses are mangrove sea-level data from Sunda Shelf (red are in situ, dark green are not in situ) (Hanebuth et al., 2000). Depth and age range indicated for relative sea-level data. (D) Rate of eustatic sea-level rise (Clark et al., 2009; Peltier, 2004) and EMBM-modeled LIS contributions (red vertical bars). (E) Percent of the LIS (red) and CIS (dark green) remaining relative to Last Glacial Maximum extent (Dyke, 2004). Vertical gray bar denotes the range in MWP-1A timing with its duration being <500 yr (Bard et al., 1990; Edwards et al., 1993; Hanebuth et al., 2000; Stanford et al., 2006; Weaver et al., 2003).

(AOGCM) that simulated the abrupt onset of the Bølling following the Oldest Dryas cold period ~18.5–14.6 ka (Liu et al., 2009).

2. Modeling LIS surface mass-balance

We use the EMBM of Anslow et al. (2008) to simulate the surface mass-balance of the LIS near the end of the Oldest Dryas (~15 ka) and during the Bølling (~14.2 ka). The EMBM accounts for spatial and temporal changes in a melting snow or ice surface including surface roughness and geometry with respect to incoming shortwave radiation. Meltwater refreezing is simulated following Huybrechts and deWolde (1999) (Carlson et al., 2009). Snow and ice albedos are assumed 0.8 and 0.5, respectively, based on observations of the Greenland Ice Sheet (e.g., Greuell, 2000), which is the only potential modern analog for the LIS. We use a range of surface roughness lengths for snow (0.001–0.0001 m) and ice (0.01–0.05 m) also based on Greenland Ice Sheet measurements (e.g., Duynkerke and van den Broeke, 1994; Grainger and Lister, 1966; Greuell and Konzelmann, 1994; Smeets and van den Broeke, 2008). This model only accounts for changes in surface ablation, and neglects dynamic feedbacks such as ice surface lowering and mass loss from calving to the ocean.

The EMBM is forced with air temperature, wind speed and direction, humidity, surface shortwave radiation, downward longwave radiation, and precipitation taken from the National Center for Atmospheric Research Community Climate System Model 3 (NCAR CCSM3; Collins et al., 2006) transient simulation described in Liu et al. (2009). CCSM3 is a fully coupled AOGCM with dynamic-vegetation and sea-ice modules. The atmosphere has 26 levels and ~3.75° horizontal resolution. The ocean has 25 levels, longitudinal resolution of 3.6° and varying latitudinal resolution that increases to ~0.9° at the equator, with higher resolution also in the North Atlantic. CCSM3 was forced with transient changes in the orbit of the Earth, greenhouse gasses, reconstructed ice sheets (ICE-5G; Peltier, 2004) and meltwater flux to the oceans (Peltier, 2004) from 22 to 14.5 ka. The meltwater flux deviated from that of Peltier (2004) after 14.5 ka, and MWP-1A was not applied in these simulations. If such a flux had been added to the North Atlantic, the model would have failed to produce the Bølling warming (Liu et al., 2009).

The AOGCM successfully simulated cooling during the Oldest Dryas in response to Northern Hemisphere ice-sheet retreat and the magnitude of abrupt warming into the Bølling from a subsequent reduction of meltwater discharge to the North Atlantic (Liu et al., 2009). In particular, this AOGCM reproduced the Bølling warming of 9 ± 3 °C observed over Greenland (Severinghaus and Brook, 1999) and in the North Atlantic of ~6 °C (Bard et al., 2000), providing us confidence in using it as a reasonable climate forcing. We note, however, that AOGCMs can have biases in their absolute simulated climate. Of importance for our study is the sensitivity of CCSM3 to the reduction in meltwater discharge during the Oldest Dryas-Bølling transition, because CO_2 did not significantly change 14.5–14.0 ka. CCSM3 simulates a 25–40% reduction in Atlantic meridional overturning strength in response to 0.1 Sverdrups (Sv; 10^6 m^3 s^{-1}) of freshwater forcing to the North Atlantic, the range depending on the location and duration of the freshwater forcing, with a subsequent recovery upon removal of the freshwater forcing (Otto-Bliesner and Brady, 2010). This compares well with the average reduction of ~30% (range 5–60%) for a 0.1 Sv freshwater forcing for 100 yr followed by recovery as determined from a suite of climate models (Stouffer et al., 2006), suggesting CCSM3 has a reasonable sensitivity to freshwater discharge. Nevertheless, we have greater confidence in the simulated change in the LIS mass balance from Bølling warming, upon which we focus discussion.

We use 50-year averages from the transient DGL-A simulation at 15.05–15.00 ka and 14.25–14.20 ka. Of the two CCSM3 simulations performed with different meltwater forcing schemes (DGL-A and DGL-B), DGL-A showed the greatest agreement with proxy climate reconstructions (Liu et al., 2009) and thus was selected for our study. This simulation also had the greatest abrupt Bølling warming and thus the greatest difference in climate between ~15.0 and 14.2 ka. As mentioned above, MWP-1A was not applied to the global oceans in this simulation, which if applied to the North Atlantic would have significantly reduced Atlantic meridional overturning circulation in the model (Liu et al., 2009). By excluding this freshwater forcing, we allow for the maximum overturning circulation recovery at the time of the Bølling warming and thus maximum warming. Given this obvious extreme forcing on the EMBM, our results should be viewed as maximum estimates of the LIS surface mass-balance response to Bølling warming as simulated by this AOGCM.

We downscale vertically and horizontally the CCSM3 simulation following standard methods to a 50×50 km LIS topography taken from ICE-5G for 15 ka (pre-Bølling) and 14.5 ka (Bølling) (Peltier, 2004) using LIS-appropriate atmospheric lapse rates (-5 °C km^{-1} for temperature and 0.1 km^{-1} for precipitation) for elevation-sensitive variables (Abe-Ouchi et al., 2007; Carlson et al., 2009; Marshall et al., 2002; Pollard et al., 2000). We note that changes in the ICE-5G LIS topography are minimal between 15 and 14.5 ka (Peltier, 2004) and