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Atmospheric transmission of North Atlantic Heinrich events

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Abstract. We model the response of the climate system during Heinrich event 2 (H2) by employing an atmospheric general circulation model, using boundary conditions based on the concept of a "canonical" Heinrich event. The canonical event is initialized with a full-height Laurentide ice sheet (LIS) and CLIMAP sea surface temperatures (SSTs), followed by lowering of the LIS, then warming of North Atlantic SSTs. Our modeled temperature and wind fields exhibit spatially variable responses over the Northern Hemisphere at each stage of the H2 event. In some regions the climatic responses are additive, whereas in other regions they cancel or are of opposite sign, suggesting that Heinrich event climatic variations may have left complex signatures in geologic records. We find variations in the tropical water balance and the mass balance of ice sheets, and implications for variations in terrestrial methane production from the contraction of northern permafrost regions and the expansion of tropical wetlands.

1. Introduction

Heinrich events are instances of increased fluxes of ice-rafted debris to the North Atlantic that are thought to record massive discharges of icebergs from a partial collapse of one or more circum-North Atlantic ice sheets during the last glaciation [*Broecker et al.*, 1992; *Bond and Lotti*, 1995]. The ice discharge events, which occurred every 3-10 kyr, are part of a package of glaciologic, oceanographic, and climatic changes in the North Atlantic and adjacent regions [*Bond et al.*, 1993; *Grimm et al.*, 1993; *Broecker*, 1994]. Marine and terrestrial data indicate that the climate of regions distant from the North Atlantic also varied during Heinrich events [*Clark and Bartlein*, 1995; *Lowell et al.*, 1995; *Porter and An*, 1995; *Phillips et al.*, 1996; *Lund and Mix*, 1998], which requires either the transmission of paleoclimatic events from the North Atlantic elsewhere [*Broecker*, 1994] or a common forcing mechanism [*Bond and Lotti*, 1995], or both.

As currently understood, the general sequence of changes associated with a Heinrich event began with a period of cooling in the North Atlantic region (Figure 1). Concurrently, the Laurentide ice sheet (LIS) grew to a size at which it became unstable [*MacAyeal*, 1993], and partially collapsed, thereby increasing the flux of icebergs to the North Atlantic [*Broecker et al.*, 1992]. The fresh water supplied by the melted icebergs reduced the rate at which North Atlantic Deep Water (NADW) formed [*Keigwin and Lehman*, 1994; *Vidal et al.*, 1997], which further cooled the North Atlantic region [*Bond et al.*, 1993; *Cortijo et al.*, 1997] by reducing northward advection of relatively warm upper ocean water. The end of a Heinrich event (the cessation of the iceberg flux) was marked by increased NADW formation and abrupt warming in the circum-North Atlantic region (Figure 1) [*Bond et al.*, 1993]. Changes in atmospheric composition, in particular the concentration of methane [*Brook et al.*, 1996] and perhaps CO_2 [*Stauffer et al.*, 1998], also accompanied Heinrich events.

There is still debate on the particular timing and sequence of ice-rafting events and sea surface temperature (SST) changes in the North Atlantic, and on the specific roles of the LIS, other continental ice sheets and the ocean thermohaline circulation in generating Heinrich events [*Alley*, 1998; *Bond and Lotti*, 1995; *McCabe and Clark*, 1998]. Moreover, the structure of individual events may depart somewhat from the canonical Heinrich event outlined above. For this reason, we isolate the effects of changes in LIS ice volume and North Atlantic SSTs (and their combined effects) on climate, recognizing that both effects combine in different ways in various events.

Global registration of regional climate changes that appear to accompany Heinrich events (including H0, or the Younger Dryas climate reversal) requires a mechanism for transmitting the climate changes of the circum-North Atlantic region, or alternatively, for generating a synchronous response in distant regions to a common forcing mechanism ("joint dependency" [*Clark and Bartlein*, 1995]). Among the mechanisms that have been proposed for transmission are changes in atmospheric circulation [*Broecker*, 1994; *Clark and Bartlein*, 1995; *Porter and An*, 1995; *Mikolajewicz et al.*, 1997], in atmospheric composition (including water vapor) [*Broecker*, 1994, 1997], or in the oceanic thermohaline circulation [*Broecker*, 1994; *Behl and Kennett*, 1996; *Lund and Mix*, 1998; *Alley and Clark*, 1999].

In this paper, we use an atmospheric general circulation model (AGCM) to investigate mechanisms by which events generated in the North Atlantic region may be transmitted elsewhere through the atmosphere. Although not explicitly including an ocean response, our results suggest mechanisms (e.g., changes in the global water balance and winds) that may have influenced ocean circulation (e.g., thermohaline circulation, upwelling). We specifically examine the response of the climate system to changes that are initiated in the North At-

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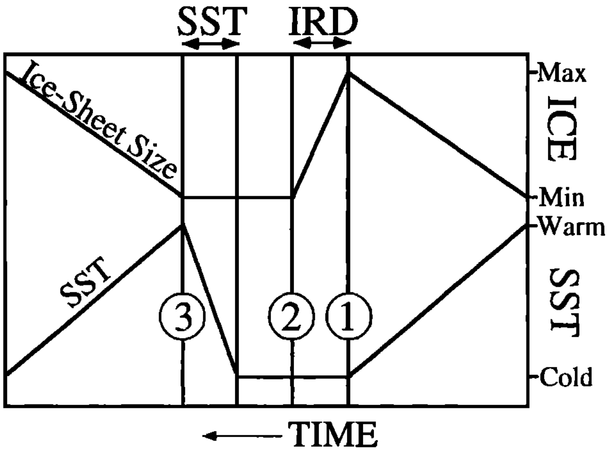

Figure 1. Conceptual model of a canonical Heinrich event as represented in our modeling experiments. Prior to a Heinrich event, ice sheet growth is accompanied by decreasing sea surface temperatures (SSTs). During the Heinrich event, SSTs remain at their coldest while the ice sheet surges (as indicated by ice rafted debris (IRD)), and rapidly increase sometime after the end of a surge [*Bond et al.*, 1993]. Boundary conditions for our three atmospheric general circulation model experiments are identified as follows: 1 MaxCold (maximum ice sheet height/cold SSTs); 2 MinCold (minimum ice sheet height/cold SSTs), and 3 MinWarm (minimum ice sheet height/warm SSTs).

lantic region through a surge (and thus lowering) of the LIS and subsequent warming of North Atlantic sea surface temperatures (Figure 1), events that are all constrained by data [*Bond et al.*, 1993; *Bond and Lotti*, 1995; *Cortijo et al.*, 1997]. Future studies, using more sophisticated coupled ocean-atmosphere-ice sheet models than now exist, will be required to evaluate some common forcing mechanism (joint dependency) that may have generated a Heinrich event in addition to other climatic responses.

2. Model and Experiments

We used the GENESIS AGCM [*Thompson and Pollard*, 1995] to examine the global sensitivity of the climate system to a partial collapse in the height of the LIS and to an abrupt increase of North Atlantic SSTs. Previous experiments with atmospheric and oceanic models have evaluated the sensitivity of regional and global climate to changes in North Atlantic SSTs or NADW formation [*Fawcett et al.*, 1997; *Manabe and Stouffer*, 1988, 1997; *Mikolajewicz et al.*, 1997; *Rind et al.*, 1986]. With the exception of three experiments that focused on the Younger Dryas [*Fawcett et al.*, 1997; *Renssen*, 1997; *Rind et al.*, 1986], however, these simulations were conducted with essentially modern boundary conditions (e.g., modern global SSTs, ice sheet configurations, insolation, and atmospheric composition), and therefore may not adequately represent the mechanisms that may be contingent on the presence of large ice sheets, generally lower SSTs, and reduced concentrations of greenhouse gasses. Similarly, experiments involving changes in the height of the ice sheets do not usually consider accompanying changes in SSTs [e.g., *Rind*, 1987]. We focus on Heinrich event 2 (H2; ca. 21 ^{14}C ka or 24.5 calendar ka) [*Bond and Lotti*, 1995], a time for which boundary conditions are relatively well known, although our results may apply to other Heinrich events.

The experiments were conducted with the T31 version of GENESIS (version 2.01) which resolves the atmosphere on an equivalent 3.75° rectangular grid, and the surface on a 2.0° grid. In all our experiments, we prescribed (modern) Dorman-Sellers vegetation types [*Dorman and Sellers*, 1989]. Otherwise, boundary conditions were specified to be consistent with the conditions that prevailed at the time of H2. For a "paleoclimatic control" simulation (MaxCold; see Figure 1) we prescribed a full-height LIS [*Licciardi et al.*, 1998] in addition to other last glacial maximum (LGM) ice sheets [*Peltier*, 1994], SSTs based on Climate: Long-Range Investigation, Mapping and Prediction (CLIMAP) [*CLIMAP Project Members*, 1981], appropriate concentration of atmospheric CO_2 (200 ppm), and insolation.

In the first "experiment" (MinCold), a post-Heinrich event lowering of the LIS (Figure 1) was approximated by replacing the full-height LIS with one ~1500 m lower over the central Hudson Bay region [*Licciardi et al.*, 1998]. In the second experiment (MinWarm), post-surge warming of the North Atlantic (Figure 1) was approximated by replacing the standard CLIMAP North Atlantic SSTs northward of 30°N latitude with SSTs prescribed to be three quarters of the way from full glacial to modern values [*Cortijo et al.*, 1997].

Each simulation was initialized from the same nominal 20-year LGM simulation and run out for 15 years; the results presented here are averaged over the last 12 years, allowing three years for the model to equilibrate to new boundary conditions. Differences ("anomalies") between simulations portray the separate effects of ice sheet collapse (MinCold - MaxCold), the combined effects of ice sheet collapse and SST increase (MinWarm - MaxCold), and the effects of SST increase alone (MinWarm - MinCold). The order of presentation of the first two anomalies below (Plate 1, left and middle columns) approximates the likely sequence of individual changes in forcing and response during a Heinrich event (Figure 1). The third set of anomalies (Plate 1, right column) may be interpreted broadly to approximate the climatic response to Dansgaard-Oeschger events.

3. Model Responses

Relative to the MaxCold simulation, the MinCold experiment produces a variety of temperature changes around North America (Plate 1, left column) that can be understood in light of the direct effects of reduced ice sheet elevation, and indirect effects generated by changes in atmospheric circulation that allow colder airmasses to dominate the region south of the ice sheet [*Kutzbach et al.*, 1993]. The net result of these changes is cooling and drying of much of North America. The lower ice sheet induces accompanying changes in atmospheric circulation, both at the surface and aloft; these extend downstream to the northeastern Pacific Ocean and include amplification of the upper-level trough and intensification of the Icelandic low over the North Atlantic, alternation of regions of lower and higher pressure over Eurasia, and reduction in the intensity of the Aleutian low in the North Pacific.

Temperature anomalies between the MaxCold simulation and the MinWarm experiment (Plate 1, middle column) are generally greater than those of the MinCold experiment, owing to the large effect of the warm North Atlantic (Plate 1, right column). Changes in circulation aloft (establishment of

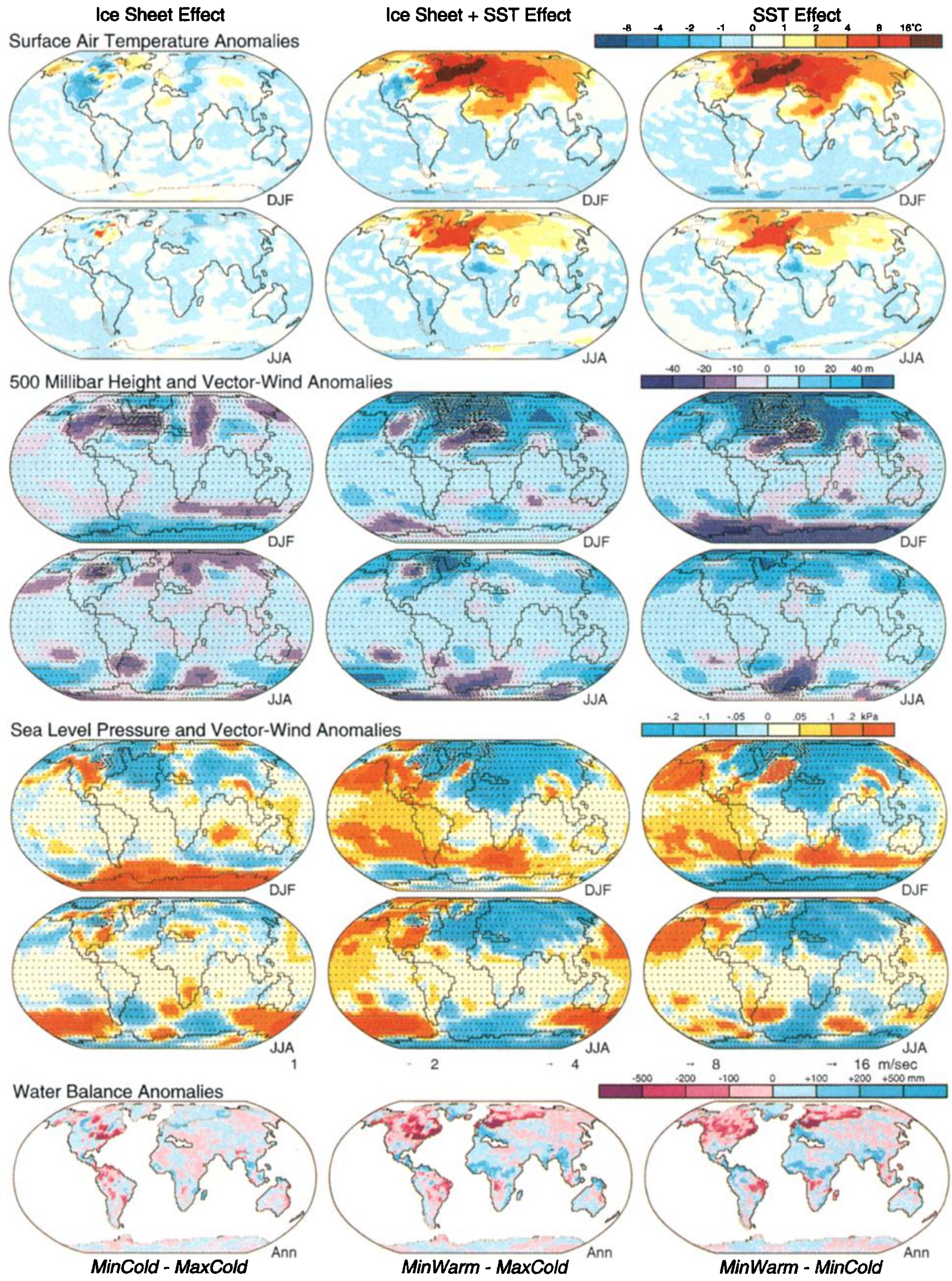

Plate 1. Seasonal (December through February (DJF) and June through August (JJA) anomalies of near-surface air temperatures, 500-mbar heights and winds, sea level pressure and winds, and terrestrial water balance ([P-E] - runoff – drainage, where P and E are precipitation and evaporation) averaged over the last 11 years of simulation. (left) Effect of lowered LIS (MinCold - MaxCold). (middle) Combined effects of lowered LIS and increased North Atlantic SSTs (MinWarm minus MaxCold). (right) Effect of increased North Atlantic SSTs alone (MinWarm minus MinCold).

an upper-level ridge over the North Atlantic) and at the surface (low pressure over Eurasia) in the MinWarm experiment advect warm air from the North Atlantic into Eurasia. The reduction of sea ice and increase in SSTs in the North Atlantic increase winter surface air temperatures up to 16°C in the North Atlantic region, and from 2° to 8°C over areas of Europe, Siberia, Asia, northern Africa, and Alaska. In the northern summer the temperature effects of the warmer North Atlantic are less because the relative magnitude of the SST anomaly is smaller. Nonetheless there are areas over the LIS, Greenland and the Eurasian ice sheets where temperatures are 4°-8°C warmer than in both the MaxCold and MinCold experiments. The SST increase thus reverses the cooling of parts of North America in the MinCold experiment.

In the MinWarm simulation, warmer SSTs in the North Atlantic cause lower sea level pressure over Eurasia in both seasons, which acts to weaken monsoonal circulation during the boreal winter and strengthen it during the boreal summer. The net result leads to positive water-balance anomalies (Plate 1, middle and right columns) throughout Central and South America, Africa, and Asia. *Fawcett et al.* [1997] reported similar patterns of change in precipitation have been reported for simulations of the Younger Dryas climate reversal conducted with GENESIS. Through this monsoonal connection, changes in LIS and North Atlantic SSTs can be transmitted into the northern tropics, and possibly across the equator. Changes in the tropical water balance also stem from modified Hadley circulation induced by changing the pole-to-equator temperature gradient. Consistent with the results of *Rind* [1998], increasing the pole-to-equator temperature gradient, as in our MaxCold simulation, tends to make the water balance more positive over the tropics and more negative over the subtropics (not shown). In our MinWarm experiment, in which the North Atlantic temperature gradient is reduced, the water balance is more positive over the northern subtropics and more negative over the tropics. While our results are consistent with those of *Rind* [1998], *Crowley and Baum* [1997] note that shifting of the band of tropical wetness may be a model-dependent feature of GENESIS. Parallel simulations with another AGCM are needed to confirm this response.

The simulated temperature and water balance anomalies suggest additional mechanisms that may be involved in the global registration of Heinrich events and in the response of paleoclimatic indicators in particular regions. For example, water balance anomalies indicate that changes in the mass balance of Northern Hemisphere ice sheets and alpine glaciers, and thus in the position of their margins, would have accompanied a Heinrich event [*Clark and Bartlein*, 1995; *Giraudi and Frezzotti*, 1997; *Lowell et al.*, 1995; *McCabe and Clark*, 1998]. The generally negative water balance anomalies over the mid and high northern latitudes (including the ice sheets) are consistent with a reduced fresh water flux to the oceans, which would contribute to restarting or reinvigorating thermohaline circulation [*Rahmstorf*, 1995; *Manabe and Stouffer*, 1997].

4. Implications for Paleoclimate Records

Abrupt increases in atmospheric methane concentrations during the interstadials that follow Heinrich events and the cold-phases of Dansgaard-Oeschger events [*Brook et al.*, 1996] are consistent with simulated warmer conditions over mid- and high-latitude Eurasia. Compared with the MaxCold simulation, in the MinWarm experiment there is a simulated decrease of ~2 × 10^7 km^2 in the areal coverage of "permafrost" (relative volume of soil ice at 0.75 to 4.25 m depth; not shown) primarily over Eurasia. Assuming a methane concentration in permafrost of 4.3 g m^{-3} [*Rasmussen et al.*, 1993], the loss of permafrost could have yielded a total methane release on the order of 10 Tg from gas hydrates stored in the extensive regions of Eurasia [*Baulin and Danilova*, 1984]. An increase in the flux of methane also may have been associated with the ensuing boreal wetlands [*Christensen et al.*, 1995].

The potential for increased methane is also suggested by the positive water balance anomalies over the African and Asian monsoon regions in the MinWarm experiment. The monsoon regions, which support a large fraction of the modern tropical wetlands [*Bartlett and Harriss*, 1993], were reduced in extent during the last glacial period [*Stokes et al.*, 1997], and may have expanded during some phases of Heinrich events [*Porter and An*, 1995; *Schultz et al.*, 1998]. Interstadial/post-Heinrich event increases in the terrestrial water balance, along with concomitant increases in precipitation minus evaporation, also point to a supplemental role (with methane) for increased tropical water vapor [*Broecker*, 1994, 1997; *Lowell et al.*, 1995] (and its inherent greenhouse effect) in transmitting northern hemisphere climate changes southward.

Terrestrial temperature and water-balance anomalies vary considerably, both spatially and among simulations (Plate 1 and Figure 2), which may explain abrupt fluctuations in lacustrine [*Benson et al.*, 1996, 1998; *Oviatt*, 1997], glacial [*Clark and Bartlein*, 1995; *Giraudi and Frezzotti*, 1997;

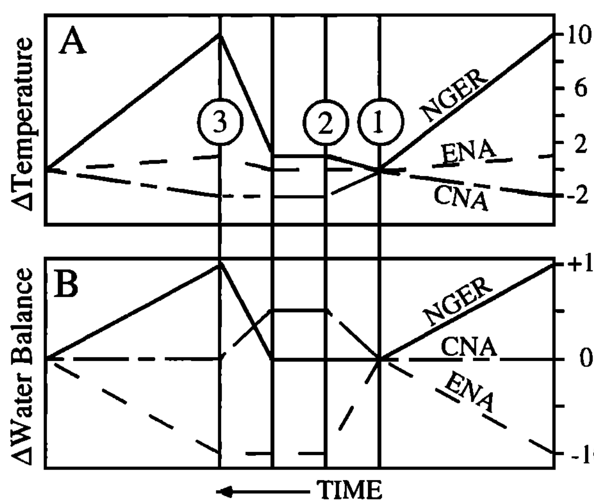

Figure 2. Our model results (Figure 2) demonstrate substantial spatial and temporal variability in response to changes in the ice sheet and SSTs. That variability is illustrated here as schematic time series of anomalies in temperature (A) and water balance (B) relative to our control LGM experiment MaxCold (1 on graph, whereas 2 = MinCold simulation and 3 = MinWarm simulation; see Figure 1) for three locales: NGER = northern Germany, centered on Hamburg, ENA = eastern North America, centered on Princeton, NJ, and CNA = central North America, centered on Boulder, CO.

Lowell et al., 1995; *Phillips et al.*, 1996] and ecological [*Grimm et al.*, 1993; *Lowell et al.*, 1995; *Watts et al.*, 1996] paleoenvironmental indicators for which temperature and/or effective-moisture variables have been implicated as controls. In some regions (e.g., the southeastern quadrant of the LIS or western Europe) the anomalies are similar in sign among simulations, but in others (e.g., central North America or the Mediterranean), the signs of the anomalies alternate (Figure 2). Paleoclimatic records from these latter regions might therefore be expected to be more variable on the timescales of Heinrich events.

Although some regions show clear responses to the changes in controls, the absence of a response in a region cannot be construed as indicating that the locale (and its paleoclimatic record) is insensitive to climatic variations related to Heinrich events. In our experiments the ice and SST boundary condition changes were discrete and sequential; this was likely not the case in the real world. Different boundary conditions (e.g., different extent of continental ice sheets) and small differences in the relative timing of the changes in controls could modify the specific response of a region during other Heinrich events. The superimposition of the effects of several potential controls, and multiple pathways through which the paleoclimatic variations in the circum-North Atlantic region may be registered globally (with different spatial patterns and potentially with different timing), call for collaborative design of climate model experiments and the comparison of their results to a broad spatial array of paleoclimatic records (both terrestrial and marine) with secure chronologies and interpretations in an hypothesis-testing framework.

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