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Role of the Bering Strait in the thermohaline circulation and abrupt climate change

Aixue Hu,[1] Gerald A. Meehl,[1] and Weiqing Han[2]

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[1] Here we investigate the role of the Bering Strait (BS) in the thermohaline circulation (THC) response to added freshwater forcing (hosing) in the subpolar North Atlantic, through analyzing simulations of a fully coupled climate model with an open and closed BS. Results show that the THC declines similarly with an open and closed BS during hosing. However, the recovery of the THC is delayed by about a century in the closed BS simulation than in the open BS one after the hosing is off. The closed BS prevents the added freshwater being transported from the Atlantic into the Pacific via the Arctic as in the open BS case. Further, the freshwater supply is elevated significantly after the hosing by exporting the freshwater stored in the Arctic during hosing, as sea ice, back to the North Atlantic. This stabilizes the surface stratification there and suppresses the recovery of the deep convection. **Citation:** Hu, A., G. A. Meehl, and W. Han (2007), Role of the Bering Strait in the thermohaline circulation and abrupt climate change, *Geophys. Res. Lett.*, *34*, L05704, doi:10.1029/2006GL028906.

1. Introduction

[2] One of the challenging questions that climate scientists face is why the climate in the current interglacial period is more stable than in the last glacial period. During the last glacial period, abrupt climate change occurred at frequencies from about a millennium to multi-millennium, such as the Dansgaard/Oeschger (D/O) events (oscillations) [*Dansgaard et al.*, 1993; *Ditlevsen et al.*, 2005], the Heinrich events [*Heinrich*, 1988; *Hemming*, 2004] and the Younger Dryas [*Hemming et al.*, 2000] cold event. The Greenland ice core data indicate a temperature change up to 10°C in a few decades in these events [*Grootes et al.*, 1993; *Alley et al.*, 1993, 2003]. In the Holocene, the only abrupt climate change comparable to these events is the cold event that happened about 8200 years ago (8.2 ky event), however, the amplitude of the temperature change may be only about 50% of that in the D/O events or Heinrich events, and the duration of this 8.2 ky event is much shorter than the D/O events or Younger Dryas event [*Alley et al.*, 1997].

[3] The thermohaline circulation transports warmer, saltier upper ocean water northward to the North Atlantic marginal seas where this water is cooled and sinks to depth, then flows southward as deep water. Although existing studies have linked all of those abrupt climate change events to abrupt changes of the thermohaline circulation (THC) in the North Atlantic [*Clark et al.*, 2002; *Rahmstorf*, 2002] due to its capacity to carry huge amounts of heat to the subpolar North Atlantic [*Ganachaud and Wunsch*, 2000], it is still not clear whether the differences of those abrupt climate change events between the last glacial period and the Holocene are caused by the different forcing strength of the glacier discharge or the climate boundary conditions, such as sea level. Since the start of the Holocene stable climate is likely to coincide with the opening of the Bering Strait [*Dyke et al.*, 1996], here we investigate the role of the Bering Strait condition in determining the response timescale of the climate system, especially the THC, to a freshwater burst in the North Atlantic associated with the discharge of land glacier melt water. The possible impact of the Bering Strait on the glacial and inter-glacial climate will also be discussed.

[4] The Bering Strait is the only pathway that connects the Pacific and the Arctic, and subsequently the North Atlantic. This strait has a depth of approximately 50 meters and a width of about 150 km at present. About 0.8 Sverdrup (Sverdrup, Sv $\equiv 10^6$ m^3s^{-1}) fresher North Pacific water is transported through this strait into the Arctic, and subsequently into the North Atlantic [*Aagaard and Carmack*, 1989]. Variations of this transport, thus the amount of freshwater into the North Atlantic, can affect the strength of the THC by altering the intensity of the deep convection in the North Atlantic marginal seas, which could potentially impact the global climate [*Wadley and Bigg*, 2002; *Reason and Power*, 1994; *De Boer and Nof*, 2004; *Shaffer and Bendtsen*, 1994].

[5] Although the basic conclusions in this study are consistent with the theoretical works of *De Boer and Nof* [2004], more detailed physical processes and associated feedbacks contributing to a delayed THC recovery in a closed Bering Strait simulation are revealed in this study, such as an elevated freshwater input in the North Atlantic after the added freshwater forcing is terminated. Thus, this study provides new insights into why the Holocene climate is more stable compared to the last glacial period as the Bering Strait was open in the former period and closed for the latter.

2. Model and Experiments

[6] For our study, we use a fully coupled state-of-the-art global climate model — the National Center for Atmospheric Research (NCAR) Community Climate System Model version 2 (CCSM2) [*Kiehl and Gent*, 2004]. CCSM2 includes the Community Atmospheric Model (CAM2) at T42 resolution (roughly 2.8°) and 26 hybrid levels in the vertical, a version of the Parallel Ocean Program (POP)

[1]Climate and Global Dynamics Division, National Center for Atmospheric Research, Boulder, Colorado, USA.

[2]Department of Atmospheric and Oceanic Sciences, University of Colorado, Boulder, Colorado, USA.

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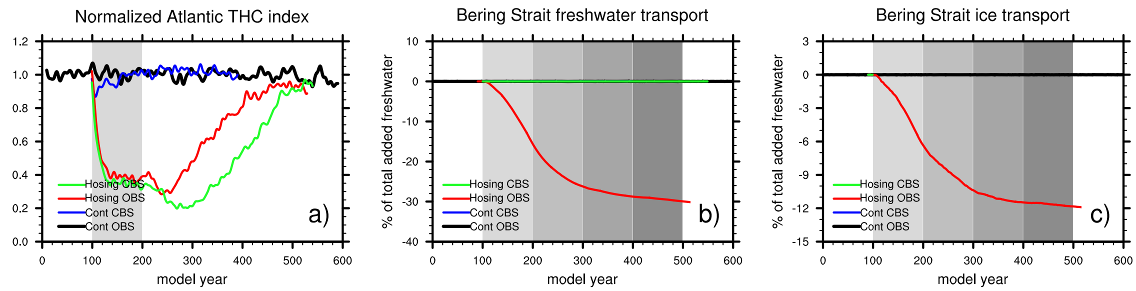

Figure 1. (a) Time evolution of a 13-year low pass filtered THC index normalized by its control mean, and (b) the cumulative anomalous transport of the freshwater, and (c) sea ice at the Bering Strait relative to the control run mean. The cumulative anomalous transports shown in Figures 1b or 1c are a percentage of the total added freshwater in the subpolar North Atlantic in the 100-year hosing period ($Q_{fwaddtot}$). The unit for these transports are % $Q_{fwaddtot}$ years. Each of the gray shades indicates a period of 100 model years, and the lightest gray shade represents the period of hosing.

developed at Los Alamos National Lab with 1° horizontal resolution and enhanced meridional resolution (1/2°) in the equatorial tropics, and with 40 vertical levels, the Community Sea Ice Model (CSIM4) with Elastic-viscous-plastic dynamics, a subgrid-scale thickness distribution, and energy conserving thermodynamics, and the Community Land Model (CLM).

[7] Two sets of experiments conducted with this model are analyzed. Each of the two sets includes a control simulation with fixed 1990s forcing conditions (with CO_2 concentration of 355 parts per million, ppm), and a freshwater perturbation (hosing) simulation. The only difference between the two sets of simulations is that one has an open Bering Strait (OBS), and the other, a closed Bering Strait (CBS). The strength of the freshwater perturbation is 1 Sv distributed uniformly in the North Atlantic between 50 and 70°N for a 100-year period (hosing period). Then this freshwater perturbation is turned off to allow the climate system to recover (recovery period). The open Bering Strait control run is a segment of 600-years of the CCSM2 millennial present day control run [*Kiehl and Gent*, 2004]. The closed Bering Strait control run and the two hosing runs are branched at year 100 of this segment. The closed Bering Strait control run is integrated for 300 years, and the two hosing runs are integrated for 450 years.

3. Results

[8] Time evolution of a 13-year lowpass filtered THC index defined as the maximum strength of the annual mean Atlantic meridional streamfuction below 500-meter depth normalized by the control mean is shown in Figure 1a. The mean THC is 15.6 and 18.3 Sv in the OBS and CBS control runs, respectively. The stronger THC in the CBS run results from the closed Bering Strait, which prevents the transport of the fresher Pacific water into the Arctic and the Atlantic, inducing an intensified deep convection in the North Atlantic, consistent with previous studies with stand-alone ocean models [*Wadley and Bigg*, 2002; *Reason and Power*, 1994]. When the freshwater perturbation is on, the THC responds similarly with an open or closed Bering Strait: a sharp decline in the first decade and a slower rate of decline thereafter. In the recovery period, the THC continues to weaken for some time and then begins to strengthen. The THC recovery starts 60 years after the hosing period in the OBS run, but 100 years in the CBS run.

[9] During the hosing, the deep convection in the North Atlantic marginal seas is almost completely shut down. In the OBS run, deep convection re-emerges in about 50 years after the hosing period in the Nordic Sea, 80 years in the Irminger Sea located at south of the Denmark Strait, and 160 years in the Labrador Sea. In the CBS run, deep convection restarts about 85 years after the hosing in the Nordic Sea, 170 years in the Irminger Sea and 210 years in the Labrador Sea. The THC indices shown in Figure 1a represent combined effects of these deep convection sites on the THC. It appears that once the THC starts to recover, the rates of the THC strengthening in these two hosing runs are similar except for the first few years. The timescale of the THC recovery in the CBS run, however, is delayed by about a century relative to the OBS run.

[10] Because the only difference between the OBS and CBS runs is the Bering Strait, the delayed recovery of the THC must be related to what happened at the Bering Strait. In the OBS control run, the northward mass, liquid freshwater and ice transports at the Bering Strait are 0.81 Sv, 0.062 Sv and 0.003 Sv, respectively, agreeing well with observed estimations (e.g., 0.8 Sv, 0.053 to 0.77 Sv, and 0.003 Sv [*Aagaard and Carmack*, 1989; *Woodgate and Aagaard*, 2005]). In the OBS hosing run, the mass transport at the Bering Strait, as well as the freshwater and sea ice transport, reverses its direction a decade after the start of the hosing (Figures 1b and 1c). Instead of the freshwater and sea ice being imported from the Pacific into the Arctic, they are exported from the Arctic into the Pacific from year 110 to year 280. Note due to the import of the freshwater added in the subpolar North Atlantic, the Arctic surface water becomes fresher than the surface water in the North Pacific in the hosing experiments. In this 170-year period with a reversed Bering Strait throughflow, about 23% of the total freshwater added into the subpolar North Atlantic ($Q_{fwaddtot}$, hereafter) exits directly through the OBS into the Pacific as liquid freshwater (14.5%) and ice (8.6%). If the differences of the total freshwater transport (liquid freshwater and ice) between the OBS hosing run and the control run are taken, we find that the North Pacific obtains an equivalent of about 22% of the $Q_{fwaddtot}$ via the status changes of the Bering Strait throughflow during that 100-year hosing period. This

Table 1. Bering Strait Transports and the Freshwater Budget Between 40 and 80°N[a]

			P − E + R		Ice-melt		Q_{OT40N}		Q_{OT80N}		Budget	
Period	FW_{BS} OBS	Ice_{BS} OBS	OBS	CBS	OBS	CBS	OBS	CBS	OBS	CBS	OBS	CBS
0–100	15.4	6.3	0.8	−1.1	−6.1	−9.5	59.5	74.9	36.2	21.3	90.4	85.6
0–200	26.2	10.4	0	−3.1	−9.4	−18.4	50.6	84.6	50.6	26.1	91.8	89.2
0–300	28.7	11.5	−1.4	−6.1	−10.4	−22.2	54.2	88.5	53.9	27.7	96.3	87.9
0–400	29.9	11.8	−2.2	−8.2	−9.6	−22.4	56.7	96.6	53.9	26.7	98.8	92.7
0–450	30.6	12.1	−2.6	−8.7	−9.1	−22.5	56.8	98.4	54.3	26.2	99.4	93.4

[a]Cumulative anomalous freshwater (FW_{BS}) and ice (Ice_{BS}) transport at the Bering Strait; the cumulative freshwater transport by the ocean at 40°N (Q_{OT40N}) and 80°N (Q_{OC80N}), and the cumulative freshwater input at the ocean surface due to precipitation, evaporation and river runoff (P − E + R), and ice-melt, and the cumulative freshwater divergence between 40° and 80°N of the Atlantic (budget, a summation of P − E + R, Ice-melt, Q_{OT40N} and Q_{OT80}N). All numbers related to freshwater shown in this table are the percentage of the total added freshwater in the subpolar North Atlantic ($Q_{fwaddtot}$) with a unit of % $Q_{fwaddtot}$ years. Period indicates the start and end years when the cumulative anomalous freshwater transport is calculated. Positive (negative) numbers represent freshwater being exported out (added into) the region between 40° and 80°N.

number increases to 37% 100 years after the termination of the additional freshwater flux, and to about 43% for the whole duration of the OBS hosing experiment (Table 1). However, this important process is absent in the CBS run since the Bering Strait is closed.

[11] The insight into the importance of this Bering Strait transport reversal is provided by the theoretical studies of *De Boer and Nof* [2004]. Their studies show with a collapsed THC, instead of returning to the Southern Oceans via the lower limb of the THC under present conditions, the 4 Sv of upper ocean water forced into the Atlantic by Southern Ocean winds will exit to the Pacific via the OBS. Thus, any strong freshwater flux into the North Atlantic which could induce a THC shutdown would be quickly flushed out of the North Atlantic into the North Pacific via the OBS by that 4 Sv upper Southern Ocean water. Consequently, the THC would be re-established fairly quickly. In the OBS hosing run, since the THC is not completely shutdown (possibly associated to the cooling in the North Atlantic induced by the weakened THC), only a bit less than a quarter of that 4 Sv upper ocean water forced into the Atlantic at 30°S by winds exits into the Pacific in the last decade of the hosing, the rest of that water still flows back to the southern oceans via the lower limb of the THC [*Hu and Meehl*, 2005]. As a result, less freshwater is pushed into the Pacific via the OBS in our simulation, preventing the THC in the OBS run to be recovered as quickly as in De Boer and Nof's studies.

[12] Another way to look into why the Bering Strait throughflow would reverse its direction is given by *Overland and Roach* [1987]. They found that the mean flow at the Bering Strait is primarily geostrophically controlled by the dynamic sea level differences between the Pacific and the Arctic with about 40 cm higher sea level in the former than in the latter. *Shaffer and Bendtsen* [1994] further indicated that the flow strength at the Bering Strait is proportional to this sea level difference. In the OBS control run, this sea level difference is about 30 cm and the resulting northward throughflow is 0.81 Sv. As indicated by *Levermann et al.* [2005] and many others, a shutdown of the THC can raise the sea level in the North Atlantic and lower it in the Pacific. This sea level change induced by the weakened THC causes a reversal of the pressure gradient between the North Pacific and the Arctic, and a reversed flow at the Bering Strait in the OBS hosing run. The dynamic sea level difference across the Bering Strait varies from 30 cm from west (Alaska side) to east (Siberian side) as in the control run to 0 cm a decade after the hosing, then to −31 cm in the last decade of the hosing. After the termination of the hosing, this sea level difference goes back to 0 at year 280, and returns to the control value around year 400, so as the flow at the Bering Strait.

[13] Next, we are going to further study what processes, other than those that have been mentioned in the previous two paragraphs, may have contributed to the delayed recovery of the THC in the CBS run through analyzing the possible direct and indirect effect of this Bering Strait throughflow reversal on the deep convection in the North Atlantic. The direct local effect is studied by a freshwater budget analysis between 40° and 80°N in the Atlantic, and the remote effect is evaluated through the changes of the sea surface salinity contrast between North Atlantic and North Pacific [*Seidov and Haupt*, 2003; *Hu et al.*, 2004] and the changes of the meridional steric height gradient in the Atlantic basin [*Hughes and Weaver*, 1994; *Thorpe et al.*, 2001; *Hu et al.*, 2004]. The freshwater budget is calculated as a summation of the oceanic freshwater transport between 40° and 80°N in the Atlantic, the surface freshwater input into the ocean due to precipitation, evaporation and river runoff (P − E + R), and the ice melt water associated to sea ice exported from the Arctic into this region.

[14] When the freshwater is added in the hosing zone, a thin freshwater layer, less than 100-meters thick, is quickly formed which produces a shallow, but very strong halocline layer in the hosing zone and its vicinity. This shallow halocline layer suppresses the deep convection in the North Atlantic marginal seas, and leads to a near collapsed THC. However, most of the freshwater contained in this thin layer is quickly transported southward to the South Atlantic or northward into the Arctic from the hosing zone by the meridional Ekman transport proportional to the surface zonal wind stress in both runs during the hosing. By the end of the hosing, roughly the same amount of the freshwater – 96% of $Q_{fwaddtot}$ is diverged out of the subpolar North Atlantic region by oceanic transports through the northern (80°N) and southern (40°N) boundaries in the OBS and CBS runs. The pathway of this oceanic transport, however, is quite different in these two hosing runs. The southward transport at 40°N is about 60% and 75% of $Q_{fwaddtot}$ in the OBS and CBS runs, respectively, and the rest (36% and 21%) goes northward at 80°N into the Arctic. Of the freshwater transported northward, about two thirds of

that is subsequently exported into the Pacific via the Bering Strait and the other one sixth (6% of $Q_{fwaddtot}$) is exported back into the North Atlantic as sea ice in the OBS run. In the CBS run, close to one half (9.5% of $Q_{fwaddtot}$) of that freshwater transported into the Arctic is exported back to the North Atlantic as sea ice. Taking into account the contribution from P − E + R, about 9% and 15% of the $Q_{fwaddtot}$ is left in this region by the end of the hosing in the OBS and CBS runs, respectively (Table 1 and Figure S1 in the auxiliary material[1]). In other words, the surface water in the subpolar North Atlantic is saltier in the OBS run than in the CBS run at the end of the hosing, promoting a better condition for the recovery of the deep convection there in the OBS run.

[15] In the recovery period, the oceanic freshwater divergence decreases dramatically as the strong halocline layer formed during the hosing weakens. However, the rate of this halocline layer weakening is faster in the OBS run than in the CBS run due to less surface freshwater input from P − E + R and melt ice after the hosing. As shown in Table 1, in comparison to the control runs, an additional surface freshwater flux equivalent to about 11% of $Q_{fwaddtot}$ is added into the Atlantic between 40 and 80°N due to increased melt-ice imported from the Arctic and the P − E + R in the first century of the recovery period in the CBS run, but with only a third of that in the OBS run. Of the increased surface freshwater input after the hosing, more than 2/3 is from the increased melt-ice water (Table 1) and the rest is contributed by a greater reduction of the evaporation associated with the colder climate induced by the weakened THC.

[16] The increased sea ice export from the Arctic into the subpolar North Atlantic is associated with the freshwater transported by oceanic current from the North Atlantic hosing zone into the Arctic during and after the hosing. When this freshwater reaches the Arctic, a thin freshwater layer forms in the upper Arctic. This thin freshwater layer stabilizes the stratification in the upper Arctic ocean and reduces the heat flux from the interior ocean into the mixed layer (note: in the Arctic, due to the ice cover, water in the mixed layer is at the freezing point, but the water in the interior ocean is above the freezing point). Combined with the colder climate induced by the weakened THC, the ice formation in the Arctic is intensified during and a period after the hosing. Since most of the freshwater imported into the Arctic is subsequently exported into the Pacific via the Bering Strait in the OBS run, only a small portion of that freshwater is exported from the Arctic into the Nordic Sea as sea ice after the hosing. In contrast, the majority of the freshwater imported into the Arctic in the CBS run is exported into the North Atlantic as sea ice. The total freshwater exported from the Arctic after the hosing in the CBS run is close to 3 times that in the OBS run (Table 1). As a result, a strong fresher surface condition which suppresses the recovery of the deep convection in the subpolar North Atlantic is maintained longer in the CBS run than in the OBS run. Therefore, the export of the freshwater via the Bering Strait into the Pacific in the OBS run significantly reduces the amount of the freshwater being exported into the North Atlantic from the Arctic after the hosing, resulting in a more favourable condition for a quicker recovery of the THC than in the CBS run.

[17] Remotely, the transport of the freshwater flux from the North Atlantic to North Pacific via the Bering Strait in the OBS run reduces the sea surface salinity of the North Pacific. Studies show that the sea surface salinity contrast between these two basins is an indicator of the variations of the THC strength under both equilibrium [*Seidov and Haupt*, 2003] and transient states [*Hu et al.*, 2004]. As this contrast increases, the THC strengthens, and vice versa. In our case, as freshwater is dumped into the sub-polar North Atlantic, this surface salinity contrast decreases (Figure S2a). After the hosing, freshwater export at the Bering Strait leads to a quicker recovery of this salinity contrast in the OBS hosing run than in the CBS run, contributing to a quicker THC recovery.

[18] Another remote measurement of the THC strength is the meridional steric height gradient between 60°N and 30°S in the Atlantic where the steric height anomaly is defined as the zonal and vertical integral of the density anomaly divided by the reference density (1000 kg/m^3) [*Thorpe et al.*, 2001]. Studies have shown that this steric height gradient is correlated with the strength of the THC with a higher gradient associated with a stronger THC [*Hughes and Weaver*, 1994; *Thorpe et al.*, 2001; *Hu et al.*, 2004]. As freshwater is added in the north, the steric height anomaly decreases, leading to a decrease of its meridional gradient (Figure S2b). In the OBS case, since the removal of the freshwater is quicker, the recovery of steric height anomaly at 60°N is also faster, leading to a quicker recovery of the meridional steric height gradient, thus associated to an earlier recovery of the deep convection in the sub-polar North Atlantic and the THC compared to the CBS case.

4. Conclusion and Discussion

[19] As summarized in Figure 2, our coupled model study shows that the status of the Bering Strait can significantly affect the response of the THC to freshwater forcing added into the sub-polar North Atlantic region, agreeing with the earlier theoretical works [*De Boer and Nof*, 2004]. With an open Bering Strait, the added freshwater in the North Atlantic can be exported into the North Pacific via the Bering Strait. Locally, this freshwater export reduces the amount of freshwater being exported back to the subpolar North Atlantic after the hosing, leading to an earlier recovery of the deep convection there. Remotely, it leads to a quicker restoration of the sea surface salinity contrast between North Atlantic and North Pacific, as well as the meridional steric height gradient, to their values in the control runs by lowering the salinity in the North Pacific and raising the steric height anomaly in the North Atlantic after the hosing. Thus, a condition in favour of a quicker recovery of the THC after the hosing is produced in the OBS run. With a closed Bering Strait, the freshwater stored in the Arctic during the hosing is exported back to the North Atlantic as sea ice and liquid water after the hosing. This freshwater export from the Arctic, along with an increased P − E + R in the subpolar North Atlantic, produces a much higher surface freshwater input after the hosing in the CBS run than in the OBS run. As a result, a more stable surface

[1]Auxiliary materials are available in the HTML. doi:10.1029/2006GL028906.

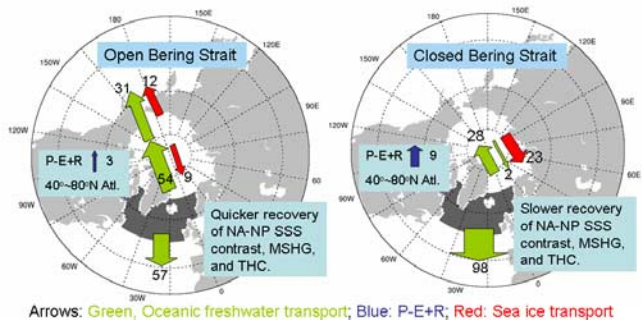

Figure 2. Schematic diagram. The arrows indicate the direction of the freshwater transport, and the numbers next to the arrows are the cumulative anomalous freshwater transport or input in the 450-year hosing period normalized by the total freshwater added into the sub-polar North Atlantic ($Q_{fwaddtot}$) with a unit of % $Q_{fwaddtot}$ years. The P − E + R denote the surface freshwater input into the ocean by precipitation, evaporation, and river runoff, and the dark gray shaded region denotes where the additional freshwater is added. The green arrows denote the oceanic freshwater transport, the red arrows the sea ice transport, and the blue arrows the surface freshwater input by P − E + R (upward arrow indicates increased freshwater input). In the CBS run, the closed Bering Strait prevents the transports of the freshwater added in the subpolar North Atlantic into the North Pacific via the Bering Strait as in the OBS run, but exports this water back to the North Atlantic as sea ice after the hosing. The resulting elevated freshwater input in the CBS run after the hosing delays the recovery of the deep convection in the North Atlantic, the sea surface salinity contrast between the North Atlantic and the North Pacific, and the meridional steric height gradient, thus a delayed recovery of the THC.

stratification is maintained longer in the CBS hosing run, leading to a delayed recovery of the deep convection, the sea surface salinity contrast, and the meridional steric height gradient, thus a delayed recovery of the THC. Further, the speculated effect of the closed Bering Strait on THC discussed by *Hu and Meehl* [2005] is confirmed here.

[20] Good analogues for the OBS and CBS hosing experiments in the real world could be the short-lived 8.2 k event during the Holocene and the long-lived Heinrich events in the last glacial period, respectively. In the 8.2 k event, a large portion of freshwater released from Lake Agassiz into the North Atlantic might have been transported into the Pacific via the Bering Strait, leading to a quicker recovery of the THC. This may also help to set up a relationship between the 8.2 k event and the stepwise freshening trend of the subarctic North Pacific in the early to mid-Holocene suggested from marine core data [*Sarnthein et al.*, 2004]. For the Heinrich events, evidence seems to suggest that freshwater derived from the massive discharge of icebergs into the North Atlantic reaches the Nordic Sea [*Hemming et al.*, 2000] and might likely further flow into the Arctic. Since the Bering Strait was closed at that time, this freshwater would have to be exported back to the North Atlantic from the Arctic after the end of the iceberg discharge, producing an elevated freshwater input, suppressing the recovery of the deep convection in the North Atlantic, and delaying the recovery of the THC and the climate.

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References

Aagaard, K., and E. C. Carmack (1989), The role of sea ice and other fresh water in the Arctic circulation, *J. Geophys. Res.*, *94*, 14,485–14,498.
Alley, R. B., et al. (1993), Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event, *Nature*, *362*, 527–529.
Alley, R. B., et al. (1997), Holocene climatic instability: A prominent, widespread event 8200 yr ago, *Geology*, *25*, 483–486.
Alley, R. B., et al. (2003), Abrupt climate change, *Science*, *299*, 2005–2010.
Clark, P. U., N. G. Pisias, T. F. Stocker, and A. J. Weaver (2002), The role of the thermohaline circulation in abrupt climate change, *Nature*, *415*, 863–869.
Dansgaard, W., et al. (1993), Evidence for general instability for past climate from a 250-kyr ice-core record, *Nature*, *364*, 218–220.
De Boer, A. M., and D. Nof (2004), The exhaust valve of the North Atlantic, *J. Clim.*, *17*, 417–422.
Ditlevsen, P. D., M. S. Kristensen, and K. K. Andersen (2005), The recurrence time of Dansgaard-Oeschger events and limits on the possible periodic component, *J. Clim.*, *18*, 2594–2603.
Dyke, A. S., J. E. Dale, and R. N. McNeely (1996), Marine mollusks as indicators of environmental change in glaciated North America and Greenland during the last 18,000 years, *Geogr. Phys. Quat.*, *50*(2), 125–184.
Ganachaud, A., and C. Wunsch (2000), Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data, *Nature*, *408*, 453–457.
Grootes, P. M., M. Stulver, J. W. C. White, S. Johnsen, and J. Jouzel (1993), Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores, *Nature*, *366*, 552–554.
Heinrich, H. (1988), Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years, *Quat. Res.*, *29*, 143–152.
Hemming, S. R. (2004), Heinrich events: Massive late Pleistocene detritus layers of the North Atlantic and their global climate imprint, *Rev. Geophys.*, *42*, RG1005, doi:10.1029/2003RG000128.
Hemming, S. R., G. C. Bond, W. S. Broecker, W. D. Sharp, and M. Klas-Mendelson (2000), Evidence from Ar40/Ar39 ages of individual hornblende grains for varying Laurentide sources of iceberg discharges 22,000 to 10,500 yr BP, *Quat. Res.*, *54*, 372–383.
Hu, A., and G. A. Meehl (2005), Bering Strait throughflow and the thermohaline circulation, *Geophys. Res. Lett.*, *32*, L24610, doi:10.1029/2005GL024424.
Hu, A., G. A. Meehl, and W. Han (2004), Detecting thermohaline circulation changes from ocean properties in a coupled model, *Geophys. Res. Lett.*, *31*, L13204, doi:10.1029/2004GL020218.
Hughes, T. M. C., and A. J. Weaver (1994), Multiple equilibria of an asymmetric two-basin ocean model, *J. Phys. Oceanogr.*, *24*, 619–637.
Kiehl, J. T., and P. R. Gent (2004), The community climate system model, version two, *J. Clim.*, *17*, 3666–3682.
Levermann, A., et al. (2005), Dynamic sea level changes following changes in the thermohaline circulation, *Clim. Dyn.*, *24*, 347–354.
Overland, J. E., and A. T. Roach (1987), Northward flow in the Bering and Chukchi Seas, *J. Geophys. Res.*, *92*, 7097–7105.
Rahmstorf, S. (2002), Ocean circulation and climate during the last 120,000 years, *Nature*, *419*, 207–214.
Reason, C. J. C., and S. B. Power (1994), The influence of the Bering Strait on the circulation in a coarse resolution global ocean model, *Clim. Dyn.*, *9*, 363–369.
Sarnthein, M., et al. (2004), Mid-Holocene origin of the sea-surface salinity low in the subarctic North Pacific, *Quat. Sci. Rev.*, *23*, 2089–2099.
Seidov, D., and B. J. Haupt (2003), Freshwater teleconnections and ocean thermohaline circulation, *Geophys. Res. Lett.*, *30*(6), 1329, doi:10.1029/2002GL016564.

Shaffer, G., and J. Bendtsen (1994), Role of the Bering Strait in controlling North Atlantic Ocean circulation and climate, *Nature*, *367*, 354–357.

Thorpe, R. B., et al. (2001), Mechanisms determining the Atlantic thermohaline circulation response to greenhouse gas forcing in a non-flux-adjusted coupled climate model, *J. Clim.*, *14*, 3102–3116.

Wadley, M. R., and G. R. Bigg (2002), Impact of flow through the Canadian Archipelago and Bering Strait on the North Atlantic and Arctic circulation: An ocean modelling study, *Q. J. R. Meteorol. Soc.*, *128*, 2187–2203.

Woodgate, R. A., and K. Aagaard (2005), Revising the Bering Strait freshwater flux into the Arctic Ocean, *Geophys. Res. Lett.*, *32*, L02602, doi:10.1029/2004GL021747.

W. Han, Department of Atmospheric and Oceanic Sciences, University of Colorado, Boulder, Campus Box 311, CO 80309, USA.

A. Hu and G. A. Meehl, Climate and Global Dynamics Division, National Center for Atmospheric Research, Boulder, CO 80305, USA. (ahu@ucar.edu)