



Rates of thermohaline recovery from freshwater pulses in modern, Last Glacial Maximum, and greenhouse warming climates

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Received 29 December 2006; revised 21 February 2007; accepted 8 March 2007; published 11 April 2007.

[1] Recovery rates of the thermohaline circulation after a freshwater pulse in the North Atlantic vary considerably depending on the background climate, as demonstrated in the Community Climate System Model. The recovery is slowest in a Last Glacial Maximum (LGM) climate, fastest in a modern climate, and intermediate between the two in a greenhouse warming (4XCO₂) climate. Previously proposed mechanisms to explain thermohaline circulation stability involving altered horizontal freshwater transport in the North Atlantic are consistent with relative recovery rates in the modern and 4XCO₂ climates, but fail to explain the slow LGM recovery. Instead, sea ice expansion inhibits deep-water formation after freshening in the LGM climate by reducing heat loss to the atmosphere and providing additional surface freshwater. In addition, anomalous vertical freshwater transport across ~1 km depth after freshening is most effective at weakening the stratification in the modern case but is negligible in the LGM case. **Citation:** Bitz, C. M., J. C. H. Chiang, W. Cheng, and J. J. Barsugli (2007), Rates of thermohaline recovery from freshwater pulses in modern, Last Glacial Maximum, and greenhouse warming climates, *Geophys. Res. Lett.*, 34, L07708, doi:10.1029/2006GL029237.

1. Introduction

[2] Shutdown of the thermohaline circulation (THC) in a modern climate from enhanced freshwater input of roughly ~10 Sv-yr (1 Sv = 10⁶ m³ s⁻¹) to the North Atlantic is usually only temporary in climate models: Once the freshwater input returns to normal, the circulation recovers in a few decades to a century at most [Stouffer *et al.*, 2006]. These models agree roughly with the duration of climate change that followed meltwater pulses approximately 8200 year ago, which resulted in the largest global cooling in the Holocene [Ellison *et al.*, 2006]. However, meltwater pulses of a similar magnitude during the most recent glacial are hypothesized to have caused the THC to remain collapsed or greatly reduced for much longer durations [see, e.g., Clark *et al.*, 2002]. It remains to be seen how anthropogenic climate change and the anticipated increase in precipitation and runoff in the northern North Atlantic will influence the THC response to freshwater pulses, say from Greenland.

[3] Here we investigate the influence of the climate state on THC recovery following a freshwater addition to the North Atlantic using the Community Climate System Model Version 3 (CCSM3). We add the same freshwater anomaly to modern, greenhouse warming, and glacial climates to investigate the mechanisms of THC recovery in a spectrum of background climates.

2. Model

[4] We impose freshwater perturbations to three nearly equilibrated background climate cases: a 1,000 yr integration with “modern” (i.e., 1990s) conditions [Collins *et al.*, 2006], a 440 yr Last Glacial Maximum LGM (with ice-sheet topography, ocean bathymetry, orbital configuration, and greenhouse gases prescribed for ~21 ka) integration [Otto-Bliesner *et al.*, 2006], and a 4XCO₂ stabilization integration [Bryan *et al.*, 2006] where CO₂ is initially increased at the rate of 1% per year from present-day to four times present-day levels and then held fixed for 300 yr. These particular background climates, or “controls”, were chosen to sample a wide range of initial THC strengths and structures and surface heat and freshwater fluxes.

[5] We branched freshwater pulse experiments from each control by instantaneously freshening the upper 970 m of the North Atlantic and Arctic Oceans from 55–90°N, 90°W–20°E by an average of 2 psu (higher at the top and tapering with depth). This is equivalent to adding 16 Sv-yr of freshwater. Our method is similar to that of Vellinga *et al.* [2002] and contrasts with that used in the intercomparison study described by Stouffer *et al.* [2006], where freshwater is added over 100 yr to the surface. The latter distributes the freshening over time, while the former distributes it in depth. Neither is obviously more realistic, but instantaneous freshening is more computationally efficient for studying THC recovery. We conducted seven freshened runs for a minimum of 20 yr each (three modern, three LGM, and one 4XCO₂). One member from each climate is longer (75 yr for modern and 4XCO₂ and 135 yr for LGM) to capture the decade to century time-scale recovery. Additional information about the model and experiments is given in the auxiliary materials.¹

3. Results

[6] Freshening in the North Atlantic causes the THC to immediately collapse with a cessation of North Atlantic Deep Water (NADW) formation in all seven runs as shown in Figure 1. In no case is the collapse permanent, but the

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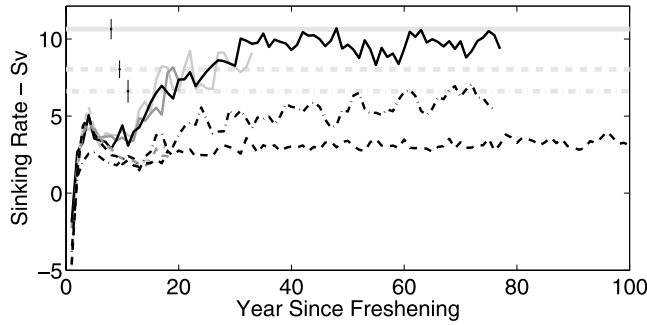


Figure 1. Annual mean thermohaline circulation index (solid lines = modern, dashed = LGM, dot-dash = 4XCO₂). The light gray lines show the mean index for the corresponding control runs with error bar indicating plus/minus one standard deviation. The index is the sinking flux across 1022 m depth from 60–65°N, which emphasizes changes in NADW formation rate, as suggested by *Gent* [2001]. Positive (negative) values indicate sinking (upwelling). An index of the THC using the maximum of the Atlantic meridional overturning streamfunction yields the same general conclusions (see auxiliary materials).

recovery rates are strikingly different depending on the background climate state. The range of recovery rates within each three-member ensemble is clearly much smaller. After a complete collapse and even reversal in year one, every run exhibits an initial rapid increase to 2–5 Sv by year 4, and then weakens again in years 5–10 (see Figure 1). After year 10 the modern THC rapidly recovers at a rate of ~ 0.2 Sv yr⁻¹. In contrast, the LGM THC recovery rate is about 10 times slower, or ~ 0.02 Sv yr⁻¹. The 4XCO₂ recover rate is ~ 0.07 Sv yr⁻¹ — intermediate to the other two. According to Figure 1, the THC recovery takes about ~ 40 yr in the modern case and about ~ 70 yr in the 4XCO₂ case, while the LGM case appears as though it could take several centuries. This dependence of the THC recovery rate on the climate state is the main finding of this study. In the rest of this section we evaluate several possible explanations.

[7] One might expect the recovery rate to depend on the degree to which the THC is controlled by thermal versus haline forcing in the different climates. Using a simple box model, *Rahmstorf* [1996] argued that the direction of freshwater transport by the meridional overturning circulation, W_{THC} , at 33°S indicates whether the THC at equilibrium is exclusively thermally driven (southward W_{THC}) or both thermally and haline driven (northward W_{THC}). W_{THC} is computed from

$$W_{\text{THC}} = -\frac{1}{S_o} \int \bar{v}(\bar{S} - S_o) dz, \quad (1)$$

where \bar{v} is the zonally integrated northward velocity, \bar{S} is the zonal mean salinity, and S_o is the reference salinity, taken here as the global mean salinity. *Rahmstorf* [1996] explained that a southward W_{THC} salinifies the Atlantic, so at equilibrium the sum of all other Atlantic freshwater contributors must brake the THC, rather than drive it. In this case, a THC shutdown eliminates a salinity source to the Atlantic, which sustains the shutdown by providing a

positive feedback. If the shutdown is only temporary, then positive feedbacks slow the recovery. A parallel argument suggests a northward W_{THC} yields a negative feedback, which aids recovery. *de Vries and Weber* [2005] found this reasoning worked well to describe a series of experiments in an intermediate complexity model where they varied W_{THC} at 33°S by applying surface freshwater flux adjustments.

[8] In our runs, only the 4XCO₂ control has a southward W_{THC} (see Table 1). Because freshening the 4XCO₂ climate does not cause a permanent THC collapse, other negative feedbacks must be operative. However, if we assume these other feedbacks are about the same strength for all three climate states, so that W_{THC} determines the relative rate of recovery, then the LGM case should recover fastest, with the modern case next, and the 4XCO₂ case last. Because the LGM case is actually the slowest to recover, we can conclude that other feedbacks must control the LGM recovery.

[9] *Vellinga et al.* [2002] argued that the THC recovery after freshening was controlled by anomalous salt advection from the North Atlantic subtropical gyre. The salinity flux by the gyre is typically expressed in terms of a freshwater flux:

$$W_{\text{gyre}} = -\frac{1}{S_o} \int \overline{v'S'} dz, \quad (2)$$

where S' and v' are deviations from the zonal means and the overbar indicates a zonal integral. After freshening, *Vellinga et al.* [2002] found W_{gyre} at 25°N decreased owing to a southward shift of the Atlantic ITCZ. The nature of the ITCZ shift after freshening depends on the mean climate state [*Cheng et al.*, 2007]. The ITCZ shift serves as a negative feedback on the THC, such that a larger decrease in W_{gyre} in the subtropical North Atlantic promotes a faster recovery. However, W_{gyre} at 33°N in the Atlantic decreases more after freshening the LGM climate than the modern or 4XCO₂ climates (see Table 1), and thus cannot explain the slow LGM recovery.

[10] The role of surface buoyancy forcing on NADW formation is another potential factor influencing the THC recovery rate. The presence of the sea ice strongly alters the surface buoyancy forcing in the subpolar seas, so we compare the sea ice cover in relation to the surface density in the freshened region for the three control climate states. Figure 2a shows that in the LGM control sea ice of 30–80% concentration covers the densest outcroppings, where NADW forms (also roughly where the mixed layer reaches its maximum depth, see auxiliary material.), suggesting a larger haline influence on the THC in the LGM climate. This is in agreement with paleoclimate evidence of brine rejection contributing to glacial NADW formation [*Vidal et al.*, 1998]. In contrast, NADW formation occurs south of the sea ice edge in the modern and 4xCO2 controls (Figures 2b

Table 1. THC Recovery Rates and Atlantic Northward Freshwater Fluxes

Background Climate	Recovery Rate, Sv yr ⁻¹	Control W_{THC} 33°S, Sv	Perturbed ^a ΔW_{gyre} 33°N, Sv
LGM	0.02	0.26	-0.47
Modern	0.2	0.03	-0.33
4XCO ₂	0.07	-0.02	-0.30

^a2nd decade after freshening.

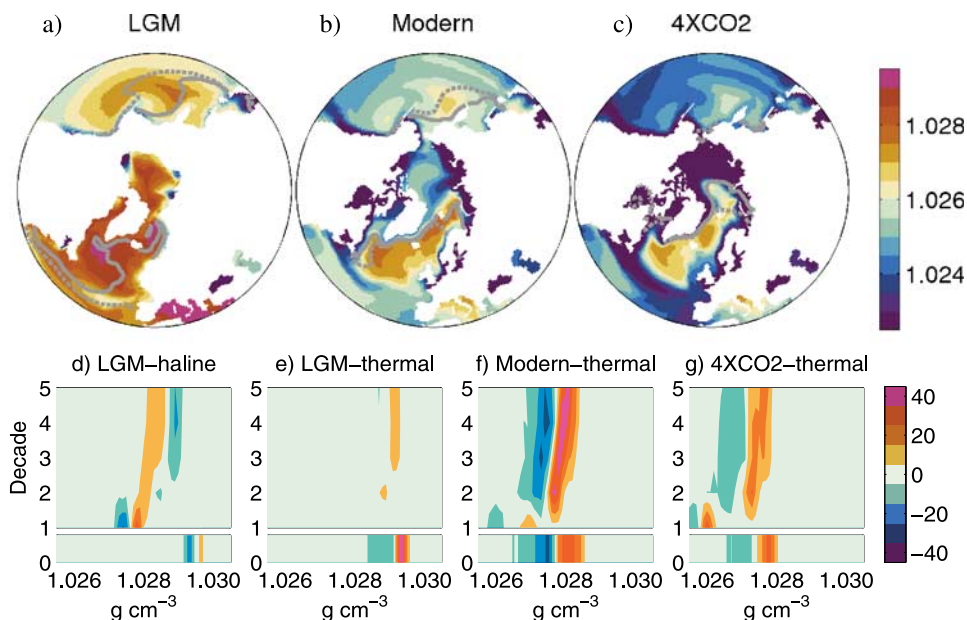


Figure 2. (a–c) Annual mean surface density in the controls in g cm^{-3} with the 15% sea ice concentration contour in the controls (solid lines) and in the second decade after freshening (dashed lines). (d–g) Watermass formation rate in $\text{Sv (kg/m}^3)^{-1}$ for the North Atlantic freshened region in the control (decade 0) and after freshening (decades 1–5). Haline components for the modern and 4XCO₂ are negligible and not shown.

and 2c). Freshening causes the sea ice cover to expand in all cases, but little sea ice reaches the NADW formation sites in the modern and 4XCO₂ climates. This disparity in sea ice cover among the climates changes the nature of feedbacks involving the surface fluxes that drive the THC recovery through deep-water formation.

[11] To quantify the effect of surface buoyancy flux forcing on watermass formation [see *Speer and Tziperman, 1992*], we first compute components of the surface density flux, $-\rho\alpha F_T$ and $\rho\beta F_S$, where ρ is the seawater density, α and β are seawater thermal and haline contraction coefficients, and F_T and F_S are surface heat and freshwater fluxes. The density flux components are integrated over density outcroppings in the Atlantic north of 55°N, south of the Arctic Ocean, and west of the Barents Sea (henceforth called the “North Atlantic freshened region”) to arrive at a watermass transformation function $\tau_{S,T}$. Finally we take the derivative, $\partial\tau_{S,T}/\partial\rho$, to give the surface watermass formation rate (WFR) as a function of ρ (with units $\text{Sv (kg/m}^3)^{-1}$) for each component. As such, the only other sources of buoyancy forcing to a given density range are from diapycnal mixing and boundary flow. If we imagine placing a bottom “boundary” on isopycnals at 1022 m depth in the sinking region, then the flow across this boundary (i.e., our index of the THC in Figure 1) is equal to the WFR from surface fluxes summed over all isopycnals that cross this boundary plus the diapycnal mixing into these isopycnals between the surface and 1 km depth.

[12] Figures 2d–2g shows WFR for the North Atlantic freshened region for all non-negligible thermal and haline components in our runs. The rate that surface fluxes further densify the heaviest outcroppings (creating NADW) is indicated by the magnitude of the rightmost maxima. The thermal component dominates NADW formation in all three controls and also after freshening in the modern and 4XCO₂ climates. Only the LGM case develops a substantial haline

component after freshening. Freshening shifts the whole pattern to the left, as outcroppings are made lighter. Then as the freshwater anomaly erodes, the pattern travel rightward, back towards its position in the control. (However, the physical location of the heaviest outcroppings and deepest mixed layer sites are not altered by freshening, see auxiliary material.) In the modern case, surface heat fluxes create dense watermasses at an even higher rate than in the control within two decades after freshening, despite the shift in density (Figure 2f). In the 4XCO₂ case, surface heat fluxes create dense watermasses after freshening at about 70% of the control rate (Figure 2g). In contrast, the LGM case has greatly reduced dense water formation after freshening (Figures 2d and 2e) owing to expanded sea ice cover over deep-water outcroppings. Sea ice in the LGM climate is transported further south and east, where it reduces heat loss and provides a new source of surface freshwater after freshening that destroys outcroppings with $\rho \sim 1.029 \text{ g cm}^{-3}$. There is some compensating increased brine rejection along with reduced precipitation (discussed in the auxiliary material) over less dense waters, with $\rho \sim 1.028 \text{ g cm}^{-3}$ (Figure 2d). Such a mechanism involving sea ice was proposed by *Stocker et al. [2001]*. The reversal of the effective salt flux over the densest water outcroppings, counteracting the thermal forcing after freshening, is the signature of a fundamental difference between the LGM climate and the modern and 4XCO₂ climates.

[13] The WFR we computed only depends on surface buoyancy forcing, yet the THC strength and structure also depends on the internal ocean stratification. Among our control simulations, the modern ocean is the least stable below about 900 m (see Figure 3a), consistent with its higher sinking rate in Figure 1 and the greater sinking depths reached by its NADW compared to the other controls [*Otto-Bliesner et al., 2006; Bryan et al., 2006*]. Comparably greater stratification in the LGM control is

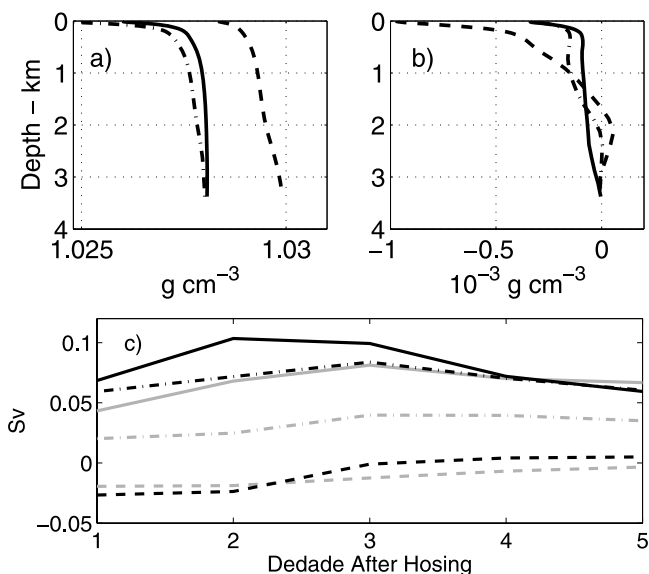


Figure 3. Density profiles in the (a) controls and (b) anomalies five decade after freshening and (c) anomalous vertical freshwater flux through 1022 m (black) and 1501 m (grey). All quantities are averaged over the North Atlantic freshened region. The reference salinity used to compute the freshwater flux in Figure 3c is the average salinity above 3.125 km depth in the region. (Solid line = modern, dashed = LGM, dot-dash = 4XCO₂.)

due mainly to salinity, in broad agreement with paleo evidence [Labeyrie, 1992], while in the 4XCO₂ control it is due mainly to temperature (not shown). Weaker stratification below the freshened layer in the modern case allows faster vertical dissipation of the freshwater anomaly, as can be seen in the potential density anomaly five decades after freshening in Figure 3b. The downward anomalous freshwater flux at 1022 m and 1501 m depths is largest in the modern case and smallest in the LGM case (Figure 3c). Anomalous downward freshwater transport erodes the high stratification that was imposed by freshening and assists in the THC recovery. This salinity homogenization after freshening is most effective in the modern case and least effective in the LGM case. After five decades, LGM and 4XCO₂ anomaly profiles below 1 km depth are clearly more stabilizing than the modern anomaly profile (Figure 3b). The strong implication is that the originally more stable density profiles below the surface layer in the LGM and 4XCO₂ controls inhibit dissipation of the salinity anomaly by transport (as noted by Tziperman *et al.* [1994]) and by convection (as noted by Lenderink and Haarsma [1993]).

[14] We also examined the horizontal freshwater transport into and out of the northern North Atlantic (integrated over the entire water column in the freshened region), as well as the surface freshwater flux. The total rate at which the freshwater anomaly dissipates from the water column depends little on the climate state (see auxiliary material). Hu *et al.* [2007] found the THC recovers more rapidly when the Bering Strait is open (compared to closed) in a modern climate in CCSM version 2 after adding freshwater at the rate of 1 Sv for 100 yr. Because their freshwater was added at the surface, the anomaly remains more surface bound. In turn, a sizeable fraction escapes through the 50 m deep

Bering Strait when it is open. In contrast, when we prescribe less than 1/6 of the total freshwater anomaly that was used by Hu *et al.* and we distribute it over the upper 970 m, we find only about 1/16 of this anomaly escapes through Bering Strait in our modern case. Thus Freshwater need not escape through Bering Strait in CCSM3 for the THC recovery to be rapid. It is also noteworthy that the THC recovery in our 4XCO₂ freshened case is much less rapid than in the modern case, despite Bering Strait being open in both.

4. Summary and Conclusions

[15] We have shown that the recovery rate from a temporary collapse of the thermohaline circulation after freshening in the north Atlantic depends on the background climate state in CCSM3. Clearly, the recovery rate from a given freshwater pulse in models with present day conditions is not a good indication of the recovery rate from a pulse of the same magnitude and location in a 4XCO₂ or glacial climate.

[16] We analyzed recovery mechanism that are influenced by horizontal freshwater transport by the mean meridional circulation at 33°S and by the subtropical gyre in the North Atlantic. If these were the dominant feedback mechanism in our model, then the LGM case would have the fastest recovery followed by the modern, with the 4XCO₂ case being the slowest.

[17] A mechanism that slows the recovery in the LGM case is the expansion of sea ice in the NADW formation regions. NADW formation via surface fluxes in the LGM freshened case alone is inhibited by sea ice expansion, which effectively reduces surface heat loss and supplies meltwater over deep-water outcroppings. In contrast, surface heat loss drives NADW formation in the modern and 4XCO₂ climates within two decades after freshening. Sea ice is poised to play a greater role in the LGM climate because *NADW production sites are partially covered with sea ice even before freshening.*

[18] After freshening the combination of the anomalous horizontal freshwater transport and surface freshwater flux in the North Atlantic freshened region are roughly the same in all three cases. Therefore the dissipation rate of the total freshwater anomaly through the surface and lateral boundaries cannot explain the differing recover rates. Instead, we find that the anomalous downward freshwater transport across ~1 km depth is effective at weakening the stratification in the modern case after freshening, while the same flux is smaller in the 4XCO₂ and nearly zero in the LGM case.

[19] **Acknowledgments.** We gratefully acknowledge support from the National Science Foundation through grants ATM-0502204 and ATM-0438201 and the Comer Science and Educational Foundation. Computational facilities were provided by the National Center for Atmospheric Research (NCAR).

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