Response of deep-sea CaCO₃ sedimentation to Atlantic meridional overturning circulation shutdown

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[1] Changes in the distribution of the preservation and burial of calcium carbonate (CaCO₃) in deep ocean sediments and associated atmospheric pCO₂ response to the shutdown of the Atlantic meridional overturning circulation (MOC) are examined using an Earth system model. We find that shutdown of the Atlantic MOC forced by the freshwater inflow significantly decreases the CaCO₃ content in North Atlantic sediments. This is a consequence of a decrease in bottom-water carbonate ion concentrations and reduction in sea-surface CaCO₃ production. The main sedimentary impacts of these two effects are separated in time, however, with reduced CaCO₃ production dominating the decrease in CaCO₃ burial during the first 1000 years after the forcing is applied. In the absence of significant overturning circulation in the Atlantic, atmospheric pCO₂ increases by 11 ppm, largely due to a decrease in POC export and a weakening biological pump. The change in pCO₂ induced by reorganization of CaCO₃ burial in deep-sea sediments is small, only 1 ppm, because increased preservation of CaCO₃ in the Pacific largely efficiently buffers decreased preservation in the Atlantic, leaving the global burial and ocean alkalinity minimally changed at equilibrium.


1. Introduction

[2] The abundance of calcium carbonate (CaCO₃) preserved and buried in ocean sediments provides valuable information about the past state of the ocean and carbon cycle. The CaCO₃ content of sediments partly reflects deep ocean circulation and is negatively affected by the increasing acidity along the path of the deep water flow, so that CaCO₃ preservation is generally better in the Atlantic than in the Pacific [e.g., Broecker and Peng, 1982; Ridgwell and Zeebe, 2005]. At the same time, CaCO₃ burial is affected by CaCO₃ production at the surface: for the same bottom water conditions, CaCO₃ content would increase with increasing production [e.g., Brummer and van Eijden, 1992]. In addition to deep-water carbonate ion concentration and sea-surface carbonate production, changes in the flux of organic matter to deep-sea sediments may also play an important role [e.g., Archer and Maier-Reimer, 1994].

[3] The superimposition of production and carbonate ion concentration effects on the CaCO₃ content of deep-sea sediments complicates the interpretation of the paleoceanographic record in terms of ocean circulation changes. During the last glaciation and deglaciation, prominent cooling events such as the Younger Dryas (YD) and Heinrich events [e.g., Rooth, 1982; Broecker et al., 1989] have been postulated to be associated with significant changes in Atlantic meridional overturning circulation (MOC). Sediment core analyses suggest that the Atlantic MOC was shut down during the drastic Heinrich event (17,500 year BP) [Barker et al., 2004]. However, there is no consensus on the exact behavior of the Atlantic MOC during the YD event because regional consistency in change in existing sediment proxies is lacking [Jansen and Veum, 1990; Veum et al., 1992; McManus et al., 2004].

[4] One way to overcome this difficulty is to understand and evaluate the response of sediment composition content to changes in the Atlantic MOC. Broecker and Clark [2001] estimate the difference in carbonate ion concentration in the Atlantic bottom water between the glacial and Holocene periods; they suggest that the enhanced CaCO₃ dissolution in the glacial sediments was due to lowered carbonate ion concentration in the water column. Heinze et al. [1999] examine model sensitivity of CaCO₃ content to the altered Atlantic MOC under the Last Glacial Maximum (LGM) condition. Since both the sedimentary CaCO₃ content and the carbonate ion concentration in the North Atlantic appear to be reduced, they too argue that the lowered CaCO₃ saturation state in the bottom water increased CaCO₃ dissolution in the sediments. On the other hand, sea-surface production may be changed by the altered Atlantic MOC [Schmittner, 2005; Marchal et al., 1998; Obata, 2007]. For instance, Schmittner [2005] found in an intermediate complexity Earth system model that the collapse of Atlantic MOC decreases the export production in Atlantic and Pacific basins. A decrease in export production in the North

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Atlantic has also been found in a coupled to a global Atmosphere-Ocean General Circulation Model (AOGCM) and land and ocean carbon cycle model [Obata, 2007]. Reduced export production would promote changes in CaCO₃ deposition and burial independently of any change in deep-sea carbonate ion concentration.

[8] Biological export of organic matter from the surface ocean in the model is based on nitrate uptake according to a Michaelis-Menten kinetic formulation. We used exactly the same biological model as Ridgwell et al. [2007a], but replacing phosphate with nitrate in a 1:17 ratio [Anderson and Sarmiento, 1994]. The uptake kinetics based on nitrate is more consistent with most of the modern ocean being limited more by nitrate than phosphate. We allow denitrification to take place in the anoxic waters, although this hardly occurs in our model due to a high rate of ventilation that maintains oxic waters in the ocean interior [Ridgwell et al., 2007a]. We replace any loss of nitrate due to denitrification, so as to maintain the whole ocean nitrate inventory, by distribution of N*, a quasi-conservate, diagnostic tracer of denitrification and nitrogen fixation [Gruber and Sarmiento, 1997] to determine the nitrogen fixation.

[9] The production rate of CaCO₃ is proportional to the production rate of POC together with the saturation state with respect to CaCO₃ at the sea surface, as formulated by Ridgwell et al. [2007a]. This means that both the POC flux as well as the CaCO₃ to POC rain ratio (the ratio of CaCO₃ to POC in export production) can change in response to ocean surface changes in nutrients and carbonate ion concentrations. Export of POC and CaCO₃ from the surface ocean can thus change independently, affecting both the rate of CaCO₃ supply to the ocean floor as well as the CaCO₃/POC rain rate.

[10] The export production of opal from the euphotic zone \((F_{\text{opal}}^{\text{he}})\) is described by

\[
F_{\text{opal}}^{\text{he}} = F_0 \text{POC}_0 \times \frac{[\text{SiO}_2]}{[\text{SiO}_2] + K_{\text{SiO}_2}^{\text{half}}}
\]

where \(F_0\text{POC}_0\) is the export production of POC, \([\text{SiO}_2]\) is the concentration of dissolved silica, \(K_{\text{SiO}_2}^{\text{half}}\) is the half-saturation constant of silica, and \(F_0\) is the opal/POC ratio. \(K_{\text{SiO}_2}^{\text{half}}\) and \(F_0\) are set to 10 \(\mu\)mol kg\(^{-1}\) and 1.0, respectively. We consider the opal production affected by the ambient dissolved silica concentration. For simplicity, the constitution of the nutrient limitation of CaCO₃ productivity is neglected. Dissolution of opal in the ocean interior follows Ridgwell et al. [2002].

[11] A 1-D sediment geochemistry model is coupled at each grid of ocean floor to represent the accumulations of calcium carbonate, opal and clay [Ridgwell and Hargreaves, 2007; Ridgwell et al., 2002]. The ocean model with water-column vertical resolution of eight levels (Figure 1a) is insufficient to capture the spatial variability in hydrostatic pressure at the sediment surface that exerts an important control on carbonate preservation. Ocean floor bathymetry is therefore obtained from ETOPO 2 (2-minute gridded relief data, 2001, http://www.ngdc.noaa.gov/mgg/fliers/01mgg04.html), not directly from the water column depth at each grid point in the ocean circulation model, as described by Ridgwell and Hargreaves [2007]. The model enables us to explicitly calculate the oceanic inventories of alkalinity and dissolved silica determined by the balance between inflow of riverine flux and outflow of sediment burial flux. We chose a riverine inflow of dissolved silica of 3.7 Tmol Si yr\(^{-1}\), consistent with observed global mean dissolved silica concentration but slightly lower than the model estimate of Ridgwell et al. [2002] of 5.1 Tmol Si yr\(^{-1}\). The export production of opal of 260 Tmol Si yr\(^{-1}\) is consistent with previous model estimate of 137–275 Tmol Si yr\(^{-1}\) [Archer et al., 2000; Ridgwell et al., 2002]. The riverine inflow of alkalinity necessary to balance the burial of CaCO₃ in the deep-sea is based on the estimation of 10 Tmol yr\(^{-1}\)
Ridgwell and Hargreaves, 2007]. The e-folding time of alkalinity and dissolved silica in the ocean, which are estimated from time series of the deviations from these initial and steady-state values, are 12 and 54 kyr, respectively.

2.2. Control Condition and Experiment Design

[12] Our control run (CTL) uses the boundary forcings of the present mean annual wind stress and a modern orbit and solar radiation. The production rate of North Atlantic Deep Water (NADW) and the cross-equatorial southward volume transport of the Atlantic deep circulation are 16.5 Sv and 9.7 Sv, respectively (Figure 2a). The distribution of CaCO3 content at the sediment surface is shown in Figure 1b, which is consistent with the core-top database [Archer, 1996]. Opal is most abundant in the sediments of Southern Ocean and North Pacific (Figure 1c) but also exhibits notable preservation in the equatorial Pacific, Indian Ocean, and North Atlantic. This pattern is similar to the global distribution of opal content [Seiter et al., 2004].

[13] In the perturbation experiment, we force a shutdown of the Atlantic MOC by applying freshwater discharge in the northern Atlantic - a ‘water-hosing’ experiment (Figure 1d). The freshwater flux of 0.3 Sv is added under the control condition and is applied continually for 10-kyr to obtain a quasi-steady state of the marine carbon cycle, including sedimentation. Under water-hosing (WF) experiments the production rate of NADW and the cross-equatorial southward volume transport are reduced to 0 Sv and −7.6 Sv (northward), respectively, shutting down the Atlantic MOC and inducing Antarctic Deep Water to flow in the opposite direction (Figure 2c). In the Pacific, the change in the circulation due to the freshwater discharge is relatively small (Figures 2b and 2d).

[14] In the following sections we describe model experiments CTL and WF with and without sediments. We focus on changes in export production of POC and CaCO3, dissolution of CaCO3 in the surface sediments, and atmospheric \( p\text{CO}_2 \).

3. CaCO3 Burial Change in the North Atlantic

[15] The CaCO3 content anomaly at the sediment surface between the WF and CTL runs is shown in Figure 3. The anomalies in the WF run are defined as departures from the CTL run. With the freshwater hosing, the CaCO3 burial flux in the WF run is reduced by 86% in the North Atlantic and increased by 56% in the Pacific. The prominent decrease in the North Atlantic content is similar to the result of the last glacial simulation obtained by Heinze et al. [1999]. The opposite burial changes between Atlantic and Pacific Oceans are similar to sediment core analyses of the glaciation [Keigwin and Jones, 1994; Catubig et al., 1998]. However, the comparison between the model result and observation must be viewed with caution, because our simulation that is forced by the freshwater flux significantly longer than the past freshwater events is highly idealized. The amplitude of change in the North Atlantic is larger than that in the Pacific, showing perhaps as expected that the impact is largest in the North Atlantic where the deep limb of the global MOC begins and that the Pacific basin is rather
larger than the Atlantic. We examine these two processes in closer detail.

[16] First, the component of bulk compositional change resulting from a change in CaCO$_3$ dissolution flux is indicated by the carbonate ion concentration anomaly in the bottom water (Figure 4). The averaged carbonate ion concentration in the North Atlantic is reduced by 36.9 mol kg$^{-1}$, which enhances sedimentary CaCO$_3$ dissolution and therefore reduces CaCO$_3$ accumulation. The pattern of reduced carbonate ion concentration is to a first order similar to that of the decreased CaCO$_3$ content in the North Atlantic, except in the polar North Atlantic where sea ice and topographic highs such as the mid Atlantic Ridge are present (Figures 3 and 4). In the Pacific, the averaged carbonate ion concentration is increased by 10.7 mol kg$^{-1}$. This weakens CaCO$_3$ dissolution, thereby preserving more CaCO$_3$ particles in the sediments. The amplitudes of the carbonate ion concentration change in Atlantic and Pacific Oceans resemble estimates of the change of $-23$ mol kg$^{-1}$ in the pelagic Atlantic and of $+18$ mol kg$^{-1}$ in equatorial Pacific Oceans from the glacial to the Holocene period [Broecker and Clark, 2001].

[17] Next, we examine the effect of altered POC and CaCO$_3$ export fluxes between the WF and CTL runs. Figure 5 shows time series of anomalies of sea-surface averaged nitrate concentration and POC export flux in the North Atlantic. The nitrate concentration in the North Atlantic is decreased during the first 680 years after the forcing is applied (Figure 5a). This in part reflects a weakened convection. The nitrate decrease causes the POC export to decrease (Figure 5b), which is consistent with the response on timescales of hundreds of years in the freshwater hosing experiment [Schmittner, 2005]. After the first 680 years, nitrate concentration and export production
both recover and eventually reach new highs in the North Atlantic. However, the global export production is reduced by 5.2% at equilibrium, because this increase in the export flux of POC in the Atlantic is more than offset by a decrease in the Pacific and Indian Oceans (Figure 6a). The associated changes in the sea-surface averaged nitrate concentrations are $+1.34 \mu\text{mol kg}^{-1}$ in the Atlantic and $-1.26 \mu\text{mol kg}^{-1}$ in the Pacific and Indian Oceans. Interestingly, despite the increase in the POC flux in the North Atlantic, the CaCO$_3$ export flux is decreased (Figure 6b), which is caused by a reduction in CaCO$_3$ production relative to POC (Figure 6c). This change in the CaCO$_3$ production is driven by a decrease in sea-surface carbonate ion concentration in the North Atlantic (not shown).

The carbonate ion concentration at the sea surface is reduced in the freshwater hosing experiment for two reasons. The first is the dilution of all dissolved tracers at the sea surface by the freshwater inflow. Since total alkalinity (ALK) $>$ dissolved inorganic carbon (DIC), the same fractional dilution would reduce ALK more than DIC in absolute concentration. This diminishes the difference between ALK and DIC, thereby lowering the carbonate ion (CO$_3^{2-}$) concentration because $[\text{CO}_3^{2-}] \approx [\text{ALK}] - [\text{DIC}]$ [e.g., Broecker and Peng, 1982]. The second is the increased nitrate concentration at the sea surface. Since a change in nitrate concentration is inversely related to alkalinity, the nitrate increase reduces total alkalinity and therefore decreases carbonate ion concentration at the sea surface. Our formulation of CaCO$_3$ production is positively related to carbonate ion concentration [Ridgwell et al., 2007a], so that CaCO$_3$ production is reduced in the North Atlantic. However, it should be noted that considerable uncertainties remain in observation and experimental data regarding the calcification response to changes in carbonate ion concentration [Ridgwell et al., 2007b]. Should calcification and carbonate production not negatively correlate with carbonate ion concentration, our model results will somewhat overestimate the impact of Atlantic MOC shutdown on North Atlantic carbonate accumulation.

Next, we focus on the time series of the anomalies of CaCO$_3$ burial flux, deposition flux, and dissolution flux in the North Atlantic sediments (Figure 7). The burial flux anomaly in the North Atlantic is $-2.5 \times 10^{12}$ mol yr$^{-1}$ around 1,500 years after the freshwater forcing is applied (Figure 7a). This change is explained by the deposition flux anomaly minus the dissolution flux anomaly. The deposition flux anomaly initially decreases to $-2.7 \times 10^{12}$ mol yr$^{-1}$ before reaching $-0.5 \times 10^{12}$ mol yr$^{-1}$ (Figure 7b) at steady-state. On the other hand, the dissolution flux anomaly initially decreases to only $-1.6 \times 10^{12}$ mol yr$^{-1}$ before increasing to such that it is greater in WF compared to CTL (Figure 7b). The dissolution change during the first

![Figure 4](G03017)

**Figure 4.** The carbonate ion concentration anomaly ($\mu\text{mol kg}^{-1}$) in the bottom water in the WF run from the CTL run. Positive (negative) contours are black (white), and zero contour is omitted.

![Figure 5](G03017)

**Figure 5.** The anomalies of (a) the averaged nitrate concentration ($\mu\text{mol kg}^{-1}$) and (b) the POC export flux at the sea-surface in the North Atlantic in the WF run from the CTL run. The dotted lines are zero line.
500 years conversely contributes to the increase in burial. Finally, the reduced burial flux corresponds to a decrease in deposition flux and an increase in dissolution flux. After 10,000 years, the deposition and dissolution changes reach about 26% and 74% of the burial change, respectively. Our result shows that the burial change depends not only on the dissolution change but also on the deposition change, when the freshwater is discharged into the northern Atlantic. It is further notable that the main controlling factor of the burial change during the first 1000 years is the deposition change rather than the dissolution change. The YD and Heinrich events are triggered by the freshwater discharge within a millennial timescale [Johnson and McClure, 1976; Bond et al., 1992]. During these events, the North Atlantic CaCO3 sedimentation might thus be impacted more by a change in CaCO3 deposition than by a change in CaCO3 dissolution. Reconstruction of paleoproductivity is therefore necessary to understand the observed variability in CaCO3 sedimentation.

4. Response of Atmospheric $p$CO$_2$ to the Freshwater Discharge

[20] In order to isolate the effect from those of the solubility and biological pumps on atmospheric $p$CO$_2$, we compare the result of the CTL and WF experiment includ-
Changes in biological and solubility pumps. In the absence of sediments, these closed system responses are virtually identical with those in the closed system (Figure 8a). Changes in export production and SST in the open system are referred to weathering inputs. On the other hand, the closed system uses the GENIE-1 model without the sediment model and the alkalinity and silica inventories of the ocean are invariant. The CTL and WF runs of closed system are referred to as the closed-CTL and closed-WF runs, respectively. Time series of atmospheric \( p\ CO_2 \) anomalies between the WF and CTL runs and between the closed-WF and closed-CTL runs are shown in Figure 8a.

First, we examine the \( p\ CO_2 \) response to the collapse of Atlantic MOC in the closed system. The atmospheric \( p\ CO_2 \) anomaly in the closed-WF run from the closed-CTL run is increased by 11 ppm after the first 500 years and eventually reaches +12 ppm after 10,000 years (dashed line in Figure 8a). The collapse of Atlantic MOC reduces the global biological production of 0.63 GtC yr\(^{-1}\) (Figure 8b) and increases the global averaged SST of 0.19\(^\circ\)C (Figure 8c). In the absence of sediments, these closed system responses indicate changes in biological and solubility pumps.

The overall \( p\ CO_2 \) change in the open system is only 1 ppm lower than that in the closed system (solid line in Figure 8a). Changes in export production and SST in the open system are virtually identical with those in the closed system, because the meridional circulation is not substantial difference between two simulations. The small difference reflects the carbon cycle-climate positive feedbacks (not shown), i.e., because of the slight difference in atmospheric \( p\ CO_2 \) and thus radiative forcing of climate. In the open system, the \( CaCO_3 \) burial changes after the collapse of Atlantic MOC (Figure 8d). This causes a slight increase in alkalinity in the whole ocean (Figure 8e), which contributes the decrease in atmospheric \( p\ CO_2 \). The \( p\ CO_2 \) anomaly of \(-1\) ppm in the open system is attributable to carbonate compensation through altering \( CaCO_3 \) burial. The effect on \( p\ CO_2 \) of carbonate compensation is small, because the increase in \( CaCO_3 \) burial in the Indo-Pacific Oceans (not shown) compensates at steady state for the reduction in \( CaCO_3 \) burial in the Atlantic. The consequences of burial of \( CaCO_3 \) in deep-sea sediments being very sensitive to changes in ocean chemistry are that only a relatively small increase in ocean alkalinity is required (Figure 8e) in order to rebalance global burial with weathering input.

5. Sensitivity of \( CaCO_3 \) Burial and Atmospheric \( p\ CO_2 \) to Freshwater Flux

The influence of the magnitude of freshwater flux to the northern Atlantic on \( CaCO_3 \) burial and atmospheric \( p\ CO_2 \) is now evaluated. The following forcing fluxes are chosen: 0.05, 0.075, 0.1, 0.2, 0.5, 0.75 and 1.0 Sv (Table 1). When the fluxes are 0.05 and 0.075 Sv, the maxima in Atlantic MOC are weakened to 14.3 and 11.6 Sv, respectively, which indicate the weakening circulation. For freshwater perturbations of 0.1 and 0.2 Sv, the circulation is further weakened over the first 1000 years and completely ceases with freshwater flux of more than 0.3 Sv.

The response of sedimentary \( CaCO_3 \) content in the North Atlantic to this range in the Atlantic MOC is quite large. In all cases the \( CaCO_3 \) accumulation is reduced relative to CTL (Table 1, column 3). The contribution of reduction in \( CaCO_3 \) deposition to reduction in \( CaCO_3 \) burial is over 85% when Atlantic MOC is active (Table 1, column 4). The same contribution increases from 18% to 68% as the freshwater forcing increases from 0.1 and 1.0 Sv when the Atlantic MOC is completely collapsed. Interestingly, the response of atmospheric \( p\ CO_2 \) only reaches 14 ppm in the most extreme case (Table 1, column 5). This suggests that the imprint of freshwater-induced collapse in Atlantic MOC is most evident in the North Atlantic sediments rather than in atmospheric \( p\ CO_2 \).

6. Discussion

We have applied freshwater forcing on the time scale of 10 kyr to obtain the \( p\ CO_2 \) response due to the change in \( CaCO_3 \) content and carbonate compensation. Since the timescale of the forcing is likely longer than the timescale of freshwater perturbation during past cooling events, our simulations are highly idealized. However, the interpretation of how the \( CaCO_3 \) content changes during the first 1000 years can be applied to past fresh water discharge events occurring on the millennial timescales, such as the YD and Heinrich events.

Our WF simulations capture the substantial regional cooling in the North Atlantic and the warming in the South.
Atlantic in response to freshwater discharge, similar to coupled-AOGCM simulations [e.g., Zhang and Delworth, 2005; Dahl et al., 2005]. This pattern is consistent with a proxy data during the YD period that is shown as the SST cooling of 6°C in the North Atlantic [Bard et al., 2000] and the warming of 2°C in the South Atlantic [Arz et al., 1999]. However, we obtain the weakening response of SST in North Pacific to the freshwater forcing by the WF simulation, which is different with the simulation using a coupled-AOGCM [e.g., Zhang and Delworth, 2005]. Our model excludes any wind stress change associated with the altered MOC, which may affect the SST sensitivity in the North Pacific to the circulation collapse and explain this disparity. Further analysis of global SST and biological production changes in a coupled-AOGCM would help shed light on the disparity. We have also applied freshwater fluxes to a modern rather than glacial boundary configuration of the model, so we would not a priori expect to see the same climatic anomalies as contained in the data.

[27] The freshwater flux induced collapse of the Atlantic MOC leads to a recovery of the nitrate concentration in the North Atlantic after several hundreds of years, which is

**Figure 8.** The anomalies of (a) atmospheric pCO₂, (b) export flux of POC, and (c) global averaged SST between the WF and CTL runs (solid lines) and between closed-WF and closed-CTL runs (dashed lines). Figures 8d and 8e show the anomalies of burial flux of CaCO₃ and the total oceanic alkalinity in the WF run from the CTL run (solid lines). The thin dotted line in Figure 8d is zero line.
similar to the response found by Schmittner [2005]. This is perhaps surprising considering the low vertical resolution of the ocean circulation model we have employed [Edwards and Marsh, 2005] because Müller et al. [2006] concluded that the excess mixing and convection can occur at low vertical resolution. However, our choice of vertical resolution does not appear to unduly affect global productivity estimates, nor the CO₂ uptake response on timescales of hundreds of years and beyond [Ridgwell et al., 2007a]. Thus, only short-term (<1000 years) responses we have found should be treated with particular caution. The low model resolution does affect the initial (CTL) distribution of CaCO₃ in the sediments, both from any inaccuracy in predicting deep-ocean carbonate ion concentrations [Ridgwell and Hargreaves, 2007] as well as the smoothing effect on the continental shelf and the spreading ridges (and other topographic highs). However, in experiments contrasting the effectiveness of deep-sea sediments in buffering atmospheric CO₂ against perturbation (not shown) we have found that achieving a higher sediment spatial resolution by ‘subgridding’ the sediments [Ridgwell and Hargreaves, 2007] is not critical to the dynamical response of the system.

[28] We evaluate the effects on atmospheric pCO₂ of solubility and biological pump changes using two factorial model experiments. One is the ‘prescribed-SST’ simulation that evaluates the solubility pump kinetics in year 1000 in the closed-WF and WF runs and all other parameters in the closed-CTL and CTL runs, respectively. The other is the ‘prescribed nitrate’ simulation that uses the sea-surface nitrate concentration at 1000 years in the closed-WF and WF runs and all other parameters in the closed-CTL and CTL runs, respectively. The former and later simulations are identical to the prescribed experiments, the sea ice change forced by the freshwater perturbation is not considered. The extent of sea ice after the water hosing affects the increase of atmospheric pCO₂ through reducing the gas exchange and weakening the biological pump. There are thus other factors to the control of atmospheric pCO₂ associated with fresh water discharge that we cannot quantify here, and await assessment using fully coupled carbon-climate models.

[30] Previous model simulations show a significant decrease in atmospheric pCO₂ through carbonate compensation [Archer and Maier-Reimeer, 1994; Sigman et al., 1998]. The responses are effectively caused by the low rain ratio in the low-latitude or global ocean, but how the ratio could be lowered sufficiently on a global scale has not been identified. In this study, the ceased Atlantic MOC reduces the rain ratio only in the northern Atlantic (Figure 6c), implying that the rain ratio change due to a reorganization of the Atlantic MOC needs not extend to the global ocean. Sea level decrease of 100 m during glacial times would shift the deposition of CaCO₃ and opal from shallower to deeper basin and might help change the rain ratio globally as a result of changes in alkalinity and dissolved silica in the whole ocean. Model analysis has suggested that perturbations of the ocean dissolved silica inventory may play a role in the atmospheric CO₂ increase during deglaciation, but perhaps restricted to just 3 ppm [Ridgwell et al., 2002]. Regardless, it would be valuable to consider the response of the ocean carbon cycle under the LGM condition rather than under modern boundary conditions.

7. Summary

[31] The shutdown of the Atlantic MOC in the GENIE-1 Earth system model, forced by freshwater inflow in the northern Atlantic, promotes a decrease in CaCO₃ burial flux by 86% in the North Atlantic sediments. This decrease is mainly controlled by two factors. One is the reduction in carbonate ion concentration in the bottom water, which enhances CaCO₃ dissolution in the sediments and then reduces the burial flux. The other is the weakening of CaCO₃ production at the sea surface, which decreases CaCO₃ deposition and burial in the sediments. The effects of the dissolution and sedimentation change on the burial flux change are 26% and 74% after 10,000 years after the freshwater forcing is applied. We find that the burial flux almost depends completely on the lowered CaCO₃ production during the first 1000 years. This implies that the Younger Dryas and Heinrich events, triggered by freshwater discharge, may drive the decrease in CaCO₃ production through weakened CaCO₃ production, rather than through lowered bottom-water carbonate concentration which acts on a timescale longer than the actual duration of the freshwater events.

[32] The depleted Atlantic MOC increases atmospheric pCO₂ by 11 ppm in an open system: a response that is similar to that in a closed system. Both systems exhibit weakened biological pump. The reduction of atmospheric pCO₂ changes +4 ppm in both systems. This implies that the biological pump weakens the ocean carbon uptake and contributes to the increase in atmospheric pCO₂, which is different response pattern to the circulation shutdown under global warming [Joos et al., 1999; Plattner et al., 2001]. In the prescribed experiments, the sea ice change forced by the freshwater perturbation is not considered.

| Table 1. Summary of Correlations Between Freshwater Flux and Indices* |
|--------------------------|------------------|------------------|------------------|------------------|
| Freshwater Flux (Sv) | MOC (Sv) | CaCO₃ Anomaly (%) | Deposition Effect (%) | pCO₂ Anomaly (ppm) |
| 0.05 | 14.3 | −16.1 | 95.5 | 0.9 |
| 0.075 | 11.6 | −42.5 | 85.1 | 1.3 |
| 0.1 | 0.0 | −79.2 | 18.5 | 9.0 |
| 0.2 | 0.0 | −82.3 | 17.6 | 10.2 |
| 0.3 | 0.0 | −85.6 | 26.4 | 11.0 |
| 0.5 | 0.0 | −90.6 | 41.8 | 11.8 |
| 0.75 | 0.0 | −94.8 | 57.7 | 13.2 |
| 1.0 | 0.0 | −99.0 | 68.0 | 13.9 |

*The first, second, third, fourth, and fifth columns are freshwater forcing flux (Sv), maximum Atlantic MOC below 1000 m (Sv), CaCO₃ burial anomaly (%) in the North Atlantic sediments in the WF run from the CTL run, effect of deposition flux change on burial flux change (%), and atmospheric pCO₂ anomaly (ppm) in the WF run from the CTL run, respectively.

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$p_{\text{CO}_2}$ through the carbonate compensation is only 1 ppm, highlighting the strength of feedback with carbonate preservation in deep-sea sediments in regulating atmospheric CO$_2$ and climate.

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