Lowering of glacial atmospheric CO\textsubscript{2} in response to changes in oceanic circulation and marine biogeochemistry

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We use an Earth system model of intermediate complexity, CLIMBER-2, to investigate what recent improvements in the representation of the physics and biology of the glacial ocean imply for the atmospheric concentration. The coupled atmosphere-ocean model under the glacial boundary conditions is able to reproduce the deep, salty, stagnant water mass inferred from Antarctic deep pore water data and the changing temperature of the entire deep ocean. When carbonate compensation is included in the model, we find a CO\textsubscript{2} drawdown of 43 ppmv associated mainly with the shoaling of the Atlantic thermohaline circulation and an increased fraction of water masses of southern origin in the deep Atlantic. Fertilizing the Atlantic and Indian sectors of the Southern Ocean north of the polar front leads to a further drawdown of 37 ppmv. Other changes to the glacial carbon cycle include a decrease in the amount of carbon stored in the terrestrial biosphere (540 Pg C), which increases atmospheric CO\textsubscript{2} by 15 ppmv, and a change in ocean salinity resulting from a drop in sea level, which elevates CO\textsubscript{2} by another 12 ppmv. A decrease in shallow water CaCO\textsubscript{3} deposition draws down CO\textsubscript{2} by 12 ppmv. In total, the model is able to explain more than two thirds (65 ppmv) of the glacial to interglacial CO\textsubscript{2} change, based only on mechanisms that are clearly documented in the proxy data. A good match between simulated and reconstructed distribution of $\delta^{13}$C changes in the deep Atlantic suggests that the model captures the mechanisms of reorganization of biogeochemistry in the Atlantic Ocean reasonably well. Additional, poorly constrained mechanisms to explain the rest of the observed drawdown include changes in the organic carbon:CaCO\textsubscript{3} ratio of sediment rain reaching the seafloor, iron fertilization in the subantarctic Pacific Ocean, and changes in terrestrial weathering.

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1. Introduction

More than 2 decades after their discovery, the cause of the glacial CO\textsubscript{2} cycles remains a scientific challenge [Broecker and Peng, 1982; Archer et al., 2000; Sigman and Boyle, 2000]. The rise and fall of atmospheric CO\textsubscript{2} concentration is an integral part of the glacial cycles, accounting for about one third of changes in globally averaged temperature over the glacial cycle [Schneider von Deimling et al., 2006b]. In spite of identification of numerous potentially relevant processes, there is still no widespread agreement on mechanisms behind this important internal feedback in the Earth system. Our inability to explain the CO\textsubscript{2} variations of the past undermines our ability to predict the future response of the carbon cycle and climate to anthropogenic carbon emissions [Schefler et al., 2006; Torn and Harte, 2006].

Ice core records show a remarkable correlation between atmospheric CO\textsubscript{2} and Antarctic temperature. This correlation was evident in the original Vostok record to 400 kyr B.P. [Petit et al., 1999] and has recently been extended to 650 kyr B.P. [Siegenthaler et al., 2005]. Paleoclimate records rarely correlate with this degree of reliability and simplicity, spurring a search for a single dominant mechanism, based on or related to the temperature of Antarctica, to explain the glacial CO\textsubscript{2} cycles.

In a number of recent publications it was proposed that a large portion of glacial-interglacial CO\textsubscript{2} variations can be explained by changes in the South Ocean sea ice extent [Stephens and Keeling, 2000], Southern Ocean stratification [Gildor et al., 2002], or the Antarctic Bottom Water (AABW) production [Paillard and Parrenin, 2004]. The correlation between Antarctic temperature and CO\textsubscript{2} is indeed compelling, but ultimately we need a quantitative mechanism which can explain the atmospheric CO\textsubscript{2} drawdown, a detailed, realistic model that achieves the observed CO\textsubscript{2} drawdown, which has thus far been elusive. The most straightforward mechanism for explaining lower glacial CO\textsubscript{2} is the cooling of the world ocean. CO\textsubscript{2} solubility in seawater increases with decreasing temperature, and the temperature of the deep sea is an obvious place to look. Paleotemperature records of the deep Pacific Ocean based on Mg/Ca ratios of foraminifera show temperature changes of 2°–4°C over glacial cycles [Martin et al., 2002]. Abyssal temperatures have also been reconstructed from pore water $\delta^{18}$O data by Adkins et al. [2002], likewise showing a glacial-interglacial temperature range of 2° to 4°C. A positive feedback between ocean temperature and CO\textsubscript{2}
would amplify any initial temperature change due to growth of the ice sheets forced by orbital changes [Archer et al., 2004]. This feedback could also ultimately amplify CO$_2$ concentration changes from fossil fuel combustion [Schefver et al., 2006; Torn and Harte, 2006].

[5] A single core from the Southern Ocean in the Adkins et al. [2002] data set revealed an abyssal water mass of extremely high salinity, higher than the increase in ocean mean salinity that resulted from the lowering of sea level. Subsequent reconstruction of the $^{14}$C content of deep water from analysis of benthic corals [Robinson et al., 2005], shows that this deep salty water mass was extremely isolated from the atmosphere. Some coupled climate models of the glacial ocean circulation are able to reproduce this stagnant salty water mass by altering the fresh water balance of the Southern Ocean [Schmitter, 2003; Shin et al., 2003]. With the colder glacial climate, sea ice is able to survive the trip north to cross the latitude of the polar front. Fresh water carried by the ice leaves behind a polar convection region of higher salinity. This mechanism has been included in recent box model studies [Köhler et al., 2005; Peacock et al., 2006; Watson and Garabato, 2006] and a study using an oceanic general circulation model [Toggweiler et al., 2006]. With our model, we are able to simulate explicitly changes in atmospheric and oceanic circulations in response to glacial changes in radiative forcing. Our results in this paper represent the first quantitative assessment of the impact of interactive circulation structure on the atmospheric CO$_2$, in particular the deep salt stratification in the Southern Ocean [Adkins et al., 2002], including the effects of the sediment CaCO$_3$ feedback to atmospheric CO$_2$ by the carbonate compensation mechanism via changes in ocean pH [Broecker and Peng, 1987; Archer et al., 2000].

[6] Another recent publication by Kohfeld et al. [2005] explored evidence for enhanced glacial marine productivity, i.e., the “biological pump,” another potential mechanism of lowering glacial CO$_2$. They found that proxy data support increased marine production during Last Glacial Maximum (LGM) in the Atlantic and Indian sectors of the sub-Antarctic Ocean, north of the Antarctic polar front. South of the front, the proxy data are interpreted as decreased productivity. Data from the Pacific sector of the Southern Ocean are sparse and equivocal. An enhancement of global oceanic marine productivity during the LGM is supported by model studies [e.g., Bopp et al., 2003].

[7] We will show our finding that neither of these mechanisms is sufficient to explain the entire magnitude of the atmospheric CO$_2$ changes between glacial and interglacial times. The high correlation between CO$_2$ and temperature spurs a search for a single mechanism, but we are forced to consider the possibility that glacial CO$_2$ changes are driven not by a single mechanism but a combination of different factors. This strategy is also necessitated by the existence of changes in the carbon cycle in which we are fairly confident, such as changes in shallow water CaCO$_3$ deposition, or the amount of carbon stored in the terrestrial biosphere. In the absence of a single plausible explanation for the CO$_2$ cycles, we search for a “CO$_2$ stew” of mechanisms (Table 1) which, combined, can account for the entire CO$_2$ change, without violation of proxy paleoceanographic constraints on the glacial carbon cycle. This idea is certainly not new [Archer et al., 2000; Sigman and Boyle, 2000; Köhler et al., 2005; Peacock et al., 2006], but our challenge is to pursue it with a geographically explicit model, which includes numerous dynamical mechanisms of interactions between climate and biogeochemistry.

2. Methods

[8] Our investigation is based on an Earth system model of intermediate complexity, CLIMBER-2, that is fast enough to carry out simulations on a scale of 10,000 years while its geographical explicitness allows accounting for spatial changes in atmospheric, oceanic, and biogeochemical dynamics. The CLIMBER-2 model includes a 2.5-dimensional statistical-dynamical atmospheric model with $10^6 \times 51^2$ horizontal resolution [Petoukhov et al., 2000], a zonally averaged three-basin dynamic oceanic model based on the work of Stocker et al. [1992] with $2.5^\circ$ meridional resolution and 20 unevenly distributed vertical levels, a one-layer thermodynamic sea ice model with a simple parameterization for horizontal ice transport, a terrestrial biosphere model, an oceanic biogeochemistry model, and a phosphate-limited model for marine biota [Petoukhov et al., 2000; Ganopolski et al., 2001; Brovkin et al., 2002a]. Atmospheric and oceanic components of the version of the CLIMBER-2 model we are using here are essentially the same as in the version used in our early simulation of the LGM climate [Ganopolski et al., 1998; Petoukhov et al., 2000] and the

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Table 1. Description of Simulations

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Glacial Circulation/SST Changes</th>
<th>Nutrient Utilization in Sub-Antarctic Ocean</th>
<th>Atlantic and Indian Sectors</th>
<th>Land Carbon Reduction</th>
<th>Sea Level Change</th>
<th>50% Reduction in Shallow Water Sedimentation</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td>G</td>
<td>yes</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td>GN</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
</tr>
<tr>
<td>GNL</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td>GNLS</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>no</td>
</tr>
<tr>
<td>GNLSG</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
</tr>
</tbody>
</table>

*Imitated by 50% increase in weathering.*
LGM biogeochemistry [Brovkin et al., 2002b]. However, an important modification has been made in the sea ice components, which explains some differences between results presented here and our early LGM simulations. The early version of CLIMBER-2 includes a purely thermodynamic sea ice model, while the version used here includes a simple parameterization of horizontal sea ice transport. This makes sea ice, especially under the glacial climate conditions, an important contributor to the surface freshwater flux and explained a much large change in the ocean salinity as compared to that reported by Ganopolski et al. [1998]. Mobility of sea ice is a crucial part of many theories of the glacial carbon cycle [Keeling and Stephens, 2001; Shin et al., 2003].

Carbonte compensation is simulated by a model of deep sea sediment diagenesis [Archer, 1991] coupled to CLIMBER-2 (see Figure 1). The sediment model operates on the CLIMBER-2 spatial resolution with one extension: While the ocean model has vertical basin boundaries, the sediment model accounts for realistic oceanic hypsometry by calculating the fractional area of the grid cell floor at every intermediate model depth (14 layers from 500 to 5000 m). Surface carbonate production ceases with sea surface temperatures (SSTs) lower than 10°C [see, e.g., Iglesias-Rodrıgüez et al., 2002]. This is captured in the model by a reduction in CaCO₃ production by a factor \( f \) which is a semiempirical function of SST, \( T_s \):}

\[
f(T_s) = \frac{e^{T_s - T_0}}{5 + e^{T_s - T_0}},
\]

where \( T_0 \) is equal to 3.5°C [Maier-Reimer and Bacastow, 1990].

[10] The flux of dust to the seafloor, \( r_{\text{dust}} \), is estimated by a correlation with organic production, \( r_{\text{org}} \), based on the present-day detrital rain rate to the seafloor [Archer et al., 2002]:

\[
r_{\text{dust}} = 1.8 \cdot 10^{-15} \cdot (r_{\text{org}})^{1.41}.
\]

The production ratio of CaCO₃ to organic carbon in the surface ocean is a product of a reference “rain ratio,” \( R_r \), and a temperature-dependent factor \( f(T_s) \) (equation (1)). A value of the \( R_r \) parameter, 0.12, is tuned to best reproduce observed fields of dissolved inorganic carbon (DIC) and alkalinity (ALK) in the preindustrial simulation CTRL. In general, the coupled model tends to overestimate near-surface concentration of phosphate in the tropical regions, but in the intermediate to deep ocean, distributions of phosphate, DIC and ALK are in good agreement with the data (for details, see Brovkin et al. [2002a]). Simulated global CaCO₃ export flux, 0.85 Pg C/yr (Table 2), was within a wide range of CaCO₃ export estimates from 0.5 to 2.0 Pg C/yr [Iglesias-Rodrıgüez et al., 2002; Feely et al., 2004; Jin et al., 2006]. It is also within a more narrow range of estimates based on alkalinity observations (0.6–1.4 Pg C/yr [Jin et al., 2006]). The global mean value of the CaCO₃ to organic carbon ratio was 0.11 (0.85:7.7, see Table 2). This value is slightly higher than a value of 0.1 used in the reference CLIMBER-2 version without sedimentation [Brovkin et al., 2002a], but it is close to a range of 0.07 to 0.1 found in the model experiments by Jin et al. [2006].
[11] Shallow water carbonate sedimentation, organic carbon weathering and burial were neglected in the CTRL simulation. Surface alkalinity and DIC fields were permanently updated to compensate for the carbonate buried in the deep sea. In the equilibrium simulation integrated for 50,000 years, deep sea carbonate burial was 0.11 Pg C/yr (9.4 Tmol/yr). This value is slightly higher than 8.3 Tmol/yr of present-day CaCO$_3$ burial in deep sea estimated by Catubig et al. [1998] but well within a range of 0.1 to 0.14 Pg C/yr suggested recently by Feely et al. [2004]. It is also very close to a value of 10 Tmol/yr obtained in the experiments with carbonate sediment model by Ridgwell and Hargreaves [2007]. Because shallow water sedimentation was ignored in our simulations, we prescribed a part of the weathering which balances the deep sea burial, or deep-sea influx [Archer and Maier-Reimer, 1994] to the deep sea burial of CaCO$_3$ predicted by the sediment model (9.4 Tmol/yr). Prescribing weathering flux to the burial flux is opposite to what found in reality [Broecker and Peng, 1987; Ridgwell and Zeebe, 2005], but this method guarantees that the model is staying at equilibrium with alkalinity and DIC inventories close to observations. This approach is often used in sediment modeling studies which do not account for the shallow water carbonate sedimentation [e.g., Ridgwell and Hargreaves, 2007].

[12] A list of simulations is shown in Table 1. We performed five glacial simulations with five factors added consecutively, one factor after another. We started from a glacial climate simulation (G), then added enhanced nutrient utilization in the subpolar Southern Ocean (GN), land carbon release (GNL), sea level change (GNLS), and reduction in shallow water carbonate sedimentation (GNLSC). Details of these simulations are explained in section 3.

[13] In all glacial simulations G to GNLSC, the radiative effect of the CO$_2$ concentration was imposed (constant) in order to exclude the complicating effect of the climate to CO$_2$ feedback. Since a radiative scheme of CLIMBER-2 accounts only for one well mixed greenhouse gas, CO$_2$, the impacts of low LGM CH$_4$ and N$_2$O concentrations are accounted for by using in the radiative scheme the so-called “equivalent” CO$_2$ concentration instead of the actual CO$_2$ concentration. Assuming LGM CO$_2$ concentration of 200 ppmv and the combined radiative forcing of glacial-interglacial changes in CH$_4$ and N$_2$O of −0.7 W/m$^2$, the “equivalent” LGM CO$_2$ concentration was set to 180 ppmv. The radiative effect of the elevated dust concentration in the glacial atmosphere was not accounted for (according to simulations by Schneider von Dettingen et al. [2006b], this forcing could lead to an additional 1°C global cooling). The distribution of ice sheets and changes in elevation and land area were prescribed following Peltier [1994], and orbital forcing was taken as of 21,000 years B.P. [Berger, 1978]. The ice sheets were assumed to be in equilibrium, hence amount of freshwater corresponding to accumulated snow over the ice sheets was released in the nearest oceanic grid cell. The same river routing scheme was used for control and glacial simulations.

[14] Silicate weathering, volcanic CO$_2$ outgassing, shallow water carbonate sedimentation, as well as organic carbon weathering and burial were neglected. The sedimentary feedback to ocean pH, calcium carbonate compensation, was active in all glacial simulations. For comparison with the other studies, we also performed simulations G, GN, GNL, and GNLS without carbonate compensation and reported equilibrium CO$_2$ concentrations after 10,000 years in the Table 2. Dust deposition input to the sediment model for glacial time was doubled from the control simulation to account for increased dust load during the LGM [Rea, 1994; Mahowald et al., 1999]. Simulations with carbonate compensation were integrated for 50,000 years to allow the weathering and sedimentary CaCO$_3$ cycle to approach equilibrium. Each simulation took about 100 hours of single processor time on the IBM p655 cluster.

### 3. Results

#### 3.1. Simulated Glacial Climate

[15] The climate model under a complete set of glacial forcings (except dust, see above) simulates a global surface air temperature cooling by 5°C, and globally averaged SST and volume averaged ocean temperature decreased by 3°C. The main changes in oceanic circulation are associated with a weakening (by ~20%) and a considerable shoaling of the Atlantic meridional overturning cell and intensification of Antarctic bottom water (AABW) formation, with a much stronger penetration of AABW into the Atlantic Ocean. While in the present-day CTRL simulation AABW is present only in the southern Atlantic (Figure 2a), in the

### Table 2. Model Results

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Atmospheric CO$_2$ After Carbonate Compensation, ppmv</th>
<th>Atmospheric CO$_2$ Without Carbonate Compensation, ppmv</th>
<th>Organic Export Flux at 100 m Depth, Pg C/yr</th>
<th>Export CaCO$_3$ Flux at 100 m Depth, Pg C/yr</th>
<th>ALK$_{ave}$, µmol/kg</th>
<th>DIC$_{ave}$, µmol/kg</th>
<th>Atmospheric δ13C, %o</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL</td>
<td>280</td>
<td>-</td>
<td>7.7</td>
<td>0.85</td>
<td>2373</td>
<td>2233</td>
<td>−6.50</td>
</tr>
<tr>
<td>G (circulation and SST)</td>
<td>237</td>
<td>246</td>
<td>7.4</td>
<td>0.81</td>
<td>2410</td>
<td>2262</td>
<td>−6.48</td>
</tr>
<tr>
<td>GN (G plus nutrients)</td>
<td>200</td>
<td>211</td>
<td>9.2</td>
<td>0.85</td>
<td>2437</td>
<td>2284</td>
<td>−6.20</td>
</tr>
<tr>
<td>GNL (GN plus land)</td>
<td>215</td>
<td>243</td>
<td>9.2</td>
<td>0.85</td>
<td>2502</td>
<td>2352</td>
<td>−6.61</td>
</tr>
<tr>
<td>GNLS (GNL plus sea level)</td>
<td>227</td>
<td>251</td>
<td>9.3</td>
<td>0.86</td>
<td>2573</td>
<td>2423</td>
<td>−6.60</td>
</tr>
<tr>
<td>GNLSC (GNLS plus coral reefs)</td>
<td>215</td>
<td>-</td>
<td>9.3</td>
<td>0.86</td>
<td>2630</td>
<td>2459</td>
<td>−6.64</td>
</tr>
</tbody>
</table>

*In the CTRL simulation, weathering equals sedimentation; therefore atmospheric CO$_2$ is 280 ppmv with or without compensation. The GNLSC simulation without compensation was not performed because changes in shallow water sedimentation affect atmospheric CO$_2$ through the carbonate compensation mechanism.*
glacial simulations AABW fills the whole Atlantic basin below 2500 m (Figure 2b).

[16] Enhanced formation and increased density of AABW water masses is primarily caused by more extensive sea ice formation with associated brine rejection around Antarctica. The simulated maximum Southern Hemisphere sea ice area under LGM condition is twice as large as at present (Figure 3) which agrees well with recent glacial sea ice reconstruction [Gersonde et al., 2005]. In particular, in the Atlantic sector of the Southern Ocean the wintertime sea ice edge at LGM is located around 45°C176, about 1000 km northward of its present-day position, consistent with the paleoceanographic data. This large sea ice expansion leads to an increase of the sea ice transport out of the area of AABW formation, which increases the salinity of AABW by 0.5 psu (in addition to the 1 psu of global increase of salinity due to sea level drop) and decreases surface salinity in the Atlantic Ocean (Figure 2b). The simulated increase of AABW salinity is consistent although somewhat smaller than the deep pore water estimates by Adkins et al. [2002]. Although North Atlantic deep water (NADW) cools more than AABW under the glacial conditions, the relative increase of AABW salinity and a relative decrease of NADW salinity overwhelms the temperature effect and the density of glacial AABW becomes higher than the density of NADW. This allows AABW to occupy a much larger portion of the deep Atlantic, and causes a shoaling and a weakening of the Atlantic meridional circulation. In the Pacific, there is little change in the circulation field, but changes in temperature are comparable to the Atlantic and similar to reconstructions by Martin et al. [2002] (Figure 4a).

3.2. Modeled Glacial Biogeochemistry
3.2.1. Simulation G: Effect of Changes in Circulation and Sea Surface Temperatures
[17] Temporal dynamics of atmospheric CO₂ in the G simulation (Figure 5) demonstrates several stages of the model response to the imposed glacial boundary conditions. During the first 500 years, CO₂ drops by 18 ppmv in response to the surface ocean cooling. The CO₂ drawdown by additional 15 ppmv during the next 2000 years is a response to the reorganization of the oceanic circulation. Acidification of deep waters leads to a temporary reduction in carbonate sediment preservation until a balance between weathering and sedimentation is achieved. This process results in increased total oceanic alkalinity which decreases atmospheric CO₂ further. A further slow decrease by about 10 ppmv is associated with carbonate compensation. The full drawdown of atmospheric CO₂ after carbonate compensation is 43 ppmv (Table 2). The average temperature

Figure 2. Simulated circulation and salinity in the Atlantic basin for (a) present-day simulation CTRL and (b) glacial simulation G. (left) Overturning (Sv), orange and blue colors are for clockwise and counterclockwise circulations, respectively, and (right) salinity (psu) are shown.

Figure 3. Seasonal variations of sea ice area (10⁶ km²) in the Southern Hemisphere. Thin solid line shows simulated modern climate, thick solid line shows simulated LGM climate, dashed line shows present-day observations based on the Hadley Centre Sea Ice and Sea Surface Temperature data set [Rayner et al., 2003].
decrease of the ocean was \(3^\circ\text{C}\), therefore the equilibrium sensitivity of atmospheric \(\text{CO}_2\) to the oceanic temperature change amounts to 14 ppmv/\(^\circ\text{C}\). This is higher than the 6–10 ppmv/\(^\circ\text{C}\) estimate from a set of solubility experiments using several oceanic general circulation models (GCM) [Martin et al., 2005], in part because of the increase in volume of the high-carbon AABW water mass which was not accounted for in those models. The maximum increase in the DIC concentration, 100 \(\mu\text{mol/kg}\), is found in the deep North Atlantic, where the boundary between AABW and NADW has changed (Figure 4b). Without circulation changes and carbonate compensation, simulated atmospheric \(\text{CO}_2\) sensitivity to the SST change would be only 18 ppmv per \(3^\circ\text{C}\) or 6 ppmv/\(^\circ\text{C}\) as seen on Figure 5, in line with results by Martin et al. [2005].

### 3.2.2. Simulation GN: G Plus Iron Fertilization

In the GN simulation, the biological pump in the Atlantic and Indian sectors of the sub-Antarctic Ocean (30° to 50°S) was stimulated to complete drawdown of nutrient (\(\text{PO}_4\)) in the surface ocean imitating a productivity increase during LGM in these regions detected by Kohfeld et al. [2005], probably the result of iron fertilization from Patagonian dust [Mahowald et al., 1999]. Since the model does not include an iron cycle, the phytoplankton productivity in these regions was calculated assuming complete utilization of \(\text{PO}_4\) at the surface. The globally averaged export flux at the surface increased by 1.8 Pg C/yr (24%)

![Figure 4](image1.png)

**Figure 4.** Glacial oceanic changes: (a) temperature (\(^\circ\text{C}\)) and (b) DIC distribution (\(\mu\text{mol/kg}\)). Shown is a difference in these tracers between simulations G and CTRL.

![Figure 5](image2.png)

**Figure 5.** Dynamics of atmospheric \(\text{CO}_2\) (ppmv) in the G simulation. Initial conditions were from the CTRL simulation. Imposed glacial boundary conditions are explained in text. Arrows indicate time periods of dominant mechanisms.
and atmospheric CO$_2$ dropped by 37 ppmv in comparison with the G simulation (Table 2). This value is higher than the 15-ppmv effect found by Bopp et al. [2003] in an ocean-biogeochemistry model run driven by the LGM boundary conditions and dust flux, but that simulation had only 6% increase in global export production. Bopp et al. [2003] did not include the effect of carbonate compensation, but this effect is very small in the GN simulation: Without compensation, the atmospheric CO$_2$ difference between the GN and the G simulations is 35 ppmv (Table 2), or only 2 ppmv less than with the compensation. If CaCO$_3$ production were increased proportionally with organic carbon production, the resulting increase in CaCO$_3$ burial would act to acidify the ocean as weathering/burial steady state is restored, increasing $p$CO$_2$ by the CaCO$_3$ compensation mechanism [Archer et al., 2000]. However, because an increase in productivity was limited to subantarctic regions with low SSTs, an increase in CaCO$_3$ export flux was less pronounced than an increase in organic carbon export and globally averaged rain ratio decreased to 0.09 (0.85:9.2, Table 2).

In comparison with the G simulation (Figure 4b), the DIC concentration mainly increased in Southern Ocean and deep ocean (Figure 6a). This increase is pronounced not only in the Atlantic and Indian oceans where the forcing was applied, but also in the Pacific Ocean, which is linked to the Atlantic and Indian through the Southern Ocean.

### 3.2.3. Simulations GNL and GNLS: GN Plus Land Carbon Plus Sea Level Change

While the GN simulation explains almost the complete magnitude of glacial CO$_2$ changes, we should not neglect several other factors that worked in the opposite direction. Terrestrial pollen records reveal that during the LGM, forest cover was strongly reduced: Boreal forests were diminished because of ice sheet expansion and increased aridity of continental interiors, while tropical forests had a more open canopy due to increased aridity [Crowley, 1995; Prentice and Jolly, 2000]. Reduced atmospheric CO$_2$ levels should have a direct negative effect on the plant productivity through reduced water use efficiency [Harrison and Prentice, 2003]. Much of the land surface in the high northern latitudes was covered by ice sheets, although this reduction in forested area was partly compensated for by exposed tropical shelf areas in Southeast Asia, likely populated by forest [e.g., Otto et al., 2002].

The glacial reduction in forest cover simulated by the CLIMBER-2 model leads to a cooling of about 1°C because of increased surface albedo and reduced transpiration [Jahn et al., 2005]. These biogeophysical effects are accounted for in the glacial simulations reported here. Several model studies that account for climate, CO$_2$, and sea level changes during the LGM suggest a net reduction of terrestrial carbon storage by 440–720 Pg C [Francois et al., 1999], 820 to 850 Pg C [Joos et al., 2004], and 820 to 1100 Pg C [Otto et al., 2002]. Low glacial CO$_2$ concentration is the main factor affecting carbon storage in these simulations. Pollen-based reconstruction by Crowley [1995] suggests glacial decrease in land carbon in the range of 760 to 1040 Pg C. Independently, glacial changes in deep ocean $\delta^{13}$C have been interpreted as a reduction in land carbon storage by about 300–700 Pg C [Shackleton, 1977; Duplessy et al., 1988;...

![Figure 6. Glacial changes in oceanic DIC distribution (µmol/kg): (a) simulation GN, (b) simulation GNLS, and (c) simulation GNLSC. Shown is a difference in these tracers between the given simulation and simulation CTRL.](image-url)
Bird et al., 1994], although glacial-interglacial changes in the deep sea δ¹³C could be also interpreted as changes in global storage of methane hydrates [Maslin and Thomas, 2003]. The interactive terrestrial biosphere model in CLIMBER-2 predicts a net reduction in terrestrial carbon storage by 540 Pg C using prescribed glacial CO₂ of 185 ppmv, glacial changes in sea-land mask, and climate from the G simulation [Jahn et al., 2005]. Biomass and soil storages are lower by 190 Pg C and 350 Pg C, respectively, and about 50% of the total carbon decline is due to lowering of CO₂. Accounting for the land carbon release in the GNL simulation results in an atmospheric CO₂ increase by 15 ppmv after carbonate compensation (Table 2). This value is in line with estimates of 15 ppmv by Sigman and Boyle [2000] and 17 ppmv by Archer et al. [2000] for 500 Pg C land carbon release. Without carbonate compensation, the predicted effect of the land carbon on CO₂ is twice as large (32 ppmv, see Table 2).

[22] The buildup of ice sheets during glacial time led to reduced sea level that reached its lowest point during the LGM. In the GNLS simulation, we assumed that a 3% decrease in total ocean volume was accompanied by a 3% increase in salinity, alkalinity, and PO₄ concentrations. The initial DIC concentration was unchanged and total oceanic DIC inventory was reduced by 3%. The excessive CO₂ was initially allocated to the atmosphere in order to conserve carbon in the land-atmosphere-ocean system. In equilibrium response to these changes, atmospheric CO₂ increased by 12 ppmv after compensation, very similar to the 14 ppmv increase suggested by Broecker and Peng [1987]. Before compensation, the CO₂ increase was less (8 ppmv, Table 2). The effect of carbonate compensation on atmospheric CO₂ might be explained by a slight but globally uniform increase in carbonate production in this simulation (Table 2), see discussion of effect of iron fertilization by Archer et al. [2000].

[23] Together, these land carbon and sea level effects lead to an increase in atmospheric CO₂ by 27 ppmv after carbonate compensation. The oceanic DIC concentration is considerably increased (by about 150 µmol/kg) in comparison with the GN simulation (Figure 6b). This is mostly a sign of dissolved land carbon amplified by the carbonate compensation process, with a further contribution from the drop in sea level (Table 2).

3.2.4. Simulation GNLSC: GNLS Plus Reduction in Shallow Water Sedimentation

[24] During the glacial period, coral reef area in tropical and subtropical shelf regions was greatly reduced. Simulations of the ReefHab model by Kleypas [1997] suggested an almost fourfold reduction in coral reef production during the LGM. Our more mild assumption here is that carbonate sedimentation in shallow waters was reduced by 50% during the glacial time. We assume that shallow water sedimentation is approximately equal to deep-sea carbonate sedimentation at present [Milliman and Droxler, 1996], although a recent review by Feely et al. [2004] suggests relatively higher accumulation of CaCO₃ in continental shelf sediments (0.13 to 0.17 Pg C/yr) than along the continental margins or in the deep sea (0.1 to 0.14 Pg C/yr).

[25] Our model does not include shallow water sedimentation. In the real ocean, a decrease in shallow water CaCO₃ burial driven by a change in shelf area would manifest itself as an increasing weathering load which deep sea CaCO₃ burial must accommodate. Assuming that shallow water CaCO₃ burial is unresponsive to changes in ocean chemistry, we can impose the impact of changes in shelf area by changing the net weathering rate to the deep sea. An increase in terrestrial weathering flux ultimately leads to increase in the deep sea burial. Carbonate burial in the deep sea in equilibrium-present-day simulation CTRL is compensated by a terrestrial weathering flux of the same value (0.11 Pg C/yr). A 50% decrease in shallow water sedimentation manifested itself as a net weathering flux 50% higher in the GNLSC simulation. This results in further enhancement of the DIC and alkalinity concentrations and a consequent atmospheric CO₂ drop by 12 ppmv (Figure 6c and Table 2). In the steady state, ocean carbonate ion increase makes the deep ocean more basic, until deep ocean CaCO₃ burial balances the weathering flux. Ultimately, this scenario leads to an increase in deep sea carbonate ion concentration and a deepening of the lysocline or carbonate compensation depth (CCD).

[26] Figure 7 presents a comparison of the fraction of CaCO₃ (wt %) in the upper sediment layer. While a comparison of present-day observations averaged to the CLIMBER-2 grid (Figure 7a) with model results in the CTRL simulation (Figure 7b) shows clearly that the model performance is far from perfect, it is able to capture several important features of the carbonate distribution on the ocean floor. First, the present-day CCD is located deeper in the North Atlantic than in the South Atlantic, and this is reproduced by the model. Secondly, the difference between Atlantic and Pacific basins is captured quite well, as the CCD in the Pacific is shallower than in the Atlantic. These CCD differences are explained by less corrosive deep waters in North Atlantic relatively to the South Atlantic and Pacific [see Archer, 1996; Ridgwell and Zeebe, 2005]. The strong CaCO₃ burial in the North Pacific is due to overestimation of the biological production there (see discussion of Brovkin et al. [2002a]).

[27] The CCD depth is not substantially changed in the simulation GNLS (Figure 7c) relatively to the CTRL simulation (Figure 7b). This is expected because carbonate compensation restores the deep-sea carbonate ion concentration to something close to the CTRL value. In the GNLSC simulation the CCD deepens by about 500 m, resulting in a sedimentation shift to the model ocean floor of 5000 m in the Southern Ocean (Figure 7d). This is, perhaps, a stronger shift than would be supported by the glacial CCD reconstruction [Catubig et al., 1998; Anderson and Archer, 2002].

3.2.5. Other Mechanisms: Changes in Rain Ratio, Weathering, Nutrients Utilization in Pacific

[28] Another mechanism that may have contributed to the lower glacial atmospheric CO₂ is a decrease in carbonate to organic flux ratio [Archer and Maier-Reimer, 1994]. This mechanism involves a shift from calcite-to silicate-producing plankton during glacial periods, for example via the silica leakage mechanism [Matsumoto et al., 2002].
CLIMBER-2 does not simulate phytoplankton structure, but we tested the mechanism by reducing the rain ratio from 0.12 to 0.1 in simulation GN. The ocean carbonate burial rate initially decreases, driving the ocean toward the basic. The change in the equilibrium ocean chemistry reduces CO$_2$ by an additional 15 ppmv (Table 3).

During the glacial maximum, river runoff was different from the Holocene because of the large area covered by glaciers.

Table 3. Mechanism Contributions to Glacial Atmospheric CO$_2$ Changes

<table>
<thead>
<tr>
<th>Mechanism</th>
<th>Simulated $\Delta$CO$_2$ After Carbonate Compensation, ppmv</th>
<th>Proxy Data in Support of the Model Results</th>
</tr>
</thead>
<tbody>
<tr>
<td>Circulation and SST</td>
<td>-43</td>
<td>oceanic $\delta^{13}$C [Duplessy et al., 1988; Curry and Oppo, 2005]</td>
</tr>
<tr>
<td>Sea level</td>
<td>12</td>
<td>oceanic $\delta^{18}$O, corals [Lambeck and Chappell, 2001; Waelbroeck et al., 2002]</td>
</tr>
<tr>
<td>Nutrient utilization in sub-Antarctic, Atlantic and Indian oceans</td>
<td>-37</td>
<td>Cd/P, $^{15}$N, $^{13}$Si, [Elderfield and Rickaby, 2000; Kohfeld et al., 2005]</td>
</tr>
<tr>
<td>Land carbon</td>
<td>15</td>
<td>pollen records and oceanic $\delta^{13}$C [Bird et al., 1994; Crowley, 1995; Joos et al., 2004]</td>
</tr>
<tr>
<td>Shallow water carbonate sedimentation</td>
<td>-12</td>
<td>carbonate sediments and corals [Milliman, 1993; Milliman and Droxler, 1996]</td>
</tr>
<tr>
<td>Subtotal</td>
<td>-65</td>
<td></td>
</tr>
</tbody>
</table>

Less Certain Mechanisms

<p>| | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>10% decrease in weathering</td>
<td>2</td>
</tr>
<tr>
<td>20% decrease in rain ratio</td>
<td>-15</td>
</tr>
<tr>
<td>Nutrients utilization in subantarctic Pacific Ocean</td>
<td>-7</td>
</tr>
<tr>
<td>Total</td>
<td>-85</td>
</tr>
</tbody>
</table>
with ice sheets, exposed shelf areas, and drier climate. Glacial changes in terrestrial weathering (even direction of these changes) are uncertain, but they were unlikely to be high [Munhoven, 2002; Foster and Vance, 2006]. As a pure sensitivity test, we investigated a scenario of a 10% reduction in the terrestrial weathering flux and found an atmospheric CO\textsubscript{2} increase by 2 ppmv in response (Table 3).

Kohfeld et al. [2005] did not detect large glacial changes in productivity in the Pacific sub-Antarctic Ocean, but this region is poorly populated with data. We performed an additional model simulation with complete nutrient utilization in this region. The CO\textsubscript{2} drawdown is 7 ppmv, much less than 37 ppmv in case of the Atlantic and Indian sub-Antarctic Ocean (Table 3). This could be explained by the very small changes in glacial Pacific circulation as compared to the glacial circulation changes in the Atlantic. Additionally, surface nutrient concentrations in the Pacific sub-Antarctic Ocean were already drawn down by complete utilization of surface nutrients in the Atlantic and Indian sectors of the sub-Antarctic Ocean.

### 3.2.6. Redistribution of δ\textsuperscript{13}C

The distribution of δ\textsuperscript{13}C in the ocean is driven by several factors simultaneously: changes in ocean circulation, biological pump, and land carbon storage. In the Atlantic, a change in the water mass distribution in the G simulation results in a decrease in δ\textsuperscript{13}C below 2000 m, with a maximum decrease in the deep North Atlantic, while δ\textsuperscript{13}C in the upper ocean increases (Figure 8a). The subantarctic biological pump enhancement in the GN simulation leads to a further increase in the δ\textsuperscript{13}C gradient between upper and deeper waters (Figure 8b), and a stronger decrease in δ\textsuperscript{13}C in the Antarctic Ocean. These changes are qualitatively in line with reconstructions of δ\textsuperscript{13}C distribution in the Atlantic by Duplessy et al. [1988] and more recent synthesis of glacial δ\textsuperscript{13}C changes in West Antarctic [Curry and Oppo, 2005]. Release of terrestrial carbon in the GNL, GNLS, and GNLSC simulations causes an average δ\textsuperscript{13}C change in the ocean of −0.4‰ (Figure 8c), a bit more than the usual estimate of −0.35‰ [Shackleton, 1977].

When compared with the reconstruction of glacial-interglacial δ\textsuperscript{13}C changes by Duplessy et al. [1988], decrease in δ\textsuperscript{13}C in the deep North Atlantic is reproduced quite well by the model (Figure 9) although a shoaling of NADW in our simulation is not as substantial as in the reconstruction. The same is true for the deep Southern Ocean, although the data suggest where much stronger decrease in δ\textsuperscript{13}C. In the Indian and Pacific basins, δ\textsuperscript{13}C increases above 2000 m depth (not shown).

Ice core measurements of atmospheric δ\textsuperscript{13}CO\textsubscript{2} during the LGM [Smith et al., 1999] reveal little change between LGM and Holocene. The average LGM value of δ\textsuperscript{13}CO\textsubscript{2} is about −6.6‰, which is only by 0.1‰ smaller than the Holocene value (−6.5‰). If δ\textsuperscript{13}CO\textsubscript{2} were influenced by changes in land carbon only, changes in δ\textsuperscript{13}CO\textsubscript{2} should be about 0.3–0.4‰. This suggests that atmospheric δ\textsuperscript{13}CO\textsubscript{2} is controlled not only by the land carbon balance but also other factors such as changes in oceanic circulation and marine biology [Brovkin et al., 2002b]. In the G simulation with circulation changes only, atmospheric δ\textsuperscript{13}CO\textsubscript{2} was not affected at all (Table 3), despite a decrease in δ\textsuperscript{13}C in the

![Figure 8](image-url)
deep Atlantic. This is caused by a redistribution of ocean $^{13}$C, with an increase in the surface Atlantic and Pacific (Figure 8a). In the GN simulation, an enhanced biological pump amplified the $\delta^{13}$C gradient between deep and surface waters and led to an increase of atmospheric $^{13}$CO$_2$ by 0.3%o. Land carbon release in the GNL simulation draws $^{13}$CO$_2$ down by 0.4%o to a level of −6.6%o, in line with ice core data. Other factors, such as sea level change, did not significantly contribute.

4. Discussion

[34] Coupled model simulations of the glacial climate published to date show no consistency in changes of the Atlantic ocean circulation. Some models show a pronounced weakening and shoaling of the Atlantic meridional overturning circulation [e.g., Kim et al., 2003; Shin et al., 2003], and others show the opposite [e.g., Hewitt et al., 2001; Kitoh et al., 2001]. The latest set of LGM experiments performed with several GCMs and models of intermediate complexity, within the PMIP-2 project using identical glacial boundary conditions, shows the same discrepancy between different models [Weber et al., 2007]. Results of the CLIMBER-2 model are most similar to simulations performed with the CCSM climate model by Shin et al. [2003], who also found a shoaling and a weakening of the Atlantic meridional circulation, a pronounced increase of salinity and an intensification of AABW formation. This type of glacial ocean circulation changes is supported by proxy data for Atlantic meridional overturning [McManus et al., 2004; Lynch-Stieglitz et al., 2007], and a dominance of low $^{13}$C and $^{14}$C Antarctic water in the glacial Atlantic ocean below 2–2.5 km depth [Duplessy et al., 1988; Curry and Oppo, 2005; Robinson et al., 2005].

[35] These changes in AABW, and the associated increase in DIC in the deep North Atlantic, explain about 40 ppmv of the glacial CO$_2$ drawdown. An additional 40 ppmv decrease in atmospheric CO$_2$ is a consequence of increased nutrient utilization in the in Atlantic and Indian sectors of the sub-Antarctic Ocean, as reconstructed from proxy data [Kohfeld et al., 2005]. Although the total ocean export flux is enhanced by only 24%, the proximity to the AABW formation region increases the sensitivity of atmospheric CO$_2$ to the biological pump here [Marinov et al., 2006]. Without increased nutrient utilization, we are not able to explain the absence of a glacial to interglacial $^{13}$CO$_2$ change recorded in the ice cores [Smith et al., 1999].

[36] In total, five mechanisms well supported by proxy data and represented in the G to GNLSC simulations are together able to explain a 65 ppmv drop in CO$_2$ (Table 3), as compared to 80–90 ppmv recorded in the ice cores. The shortfall could be due to other, less well documented mechanisms, or possibly due a quantitative underestimation of the mechanisms as captured in our model. For example, simulated decrease in global mean annual surface air temperature in this series of experiments is 5°C and thus in the lower part of the best estimate range of 5.8° ± 1.4°C found by Schneider von Deimling et al. [2006a] by constraining a large model ensemble with proxy data. A greater...
cooling would have enhanced several of the mechanisms considered here. We have not accounted for changes in methane hydrate storages which may partly contribute to glacial-interglacial changes in atmospheric CO$_2$ and $\delta^{13}$CO$_2$ [Maslin and Thomas, 2003]. A record of diatom resting spore abundance in the Antarctic analyzed recently by Abelmann et al. [2006] indicate extensive phytoplankton blooms during the LGM across the entire Atlantic sector of the Antarctic Circumpolar Current. Accounting for complete nutrients utilization in the glacial Antarctic Ocean would definitely enhance CO$_2$ drawdown in the model, but this it is not supported by the other proxies such as Cd/P ratio [e.g., Elderfield and Rickaby, 2000]. Increased export production in the northwest Pacific and equatorial regions of the Atlantic and Indian Oceans during the LGM [Kohfeld et al., 2005] is another mechanism of lowering glacial CO$_2$ not considered in our study.

[37] CO$_2$ changes from the different mechanisms probably follow a distinct sequence around the glacial cycle. The CO$_2$ drawdown based on physical changes in ocean circulation is probably most important in the early stages of glaciation. An ice sheet nucleates in response to insolation changes toward the end of an interglacial climate state, with interglacial CO$_2$ levels. The ice sheet modifies the climate of the North Atlantic, but would initially have had minor climate impacts elsewhere. Both cooling and retention of freshwater in the ice sheet would make NADW denser. The cooling of North Atlantic would lead to initial CO$_2$ drawdown, an increase of the Southern Hemisphere sea ice cover, and consequent intensification of AABW which gradually getting dense and starts to replace NADW in the abyssal Atlantic. These circulation changes would result in cooling of the deep ocean and additional lowering of the atmospheric CO$_2$. Fertilization of the sub-Antarctic Ocean probably kicks in only later during a glacial stage [Kohfeld et al., 2005]. Dust fluxes in Antarctica only increase from interglacial values when the $\delta^{18}$O is at its negative extreme, in the coldest of the cold climates [Ridgwell, 2003; Wolff et al., 2006].

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References


Adkins, J. F., K. McIntyre, and D. P. Schrag (2002). The salinity, temperature, and $\delta^{18}$O of the glacial deep ocean, Science, 298, 1769 –1773.


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