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A lowering effect of reconstructed Holocene changes in sea surface temperatures on the atmospheric CO_2 concentration

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[1] One of the mechanisms proposed to explain the roughly 20 ppmv increase in atmospheric CO_2 concentration from the early to late Holocene is a warming of the ocean surface that reduces CO_2 solubility (Indermühle et al., 1999). Here we show that this hypothesis is not supported by reconstructed changes in sea surface temperatures (SSTs) because of an inhomogeneous distribution of SST changes across the globe during the course of Holocene. While alkenone-based SST reconstructions compiled in the GHOST database (Kim et al., 2004; Kim and Schneider, 2004) suggest a net warming of the surface ocean on a global scale by 0.2 ± 0.2°C, both data and model results support a significant cooling trend for the North Atlantic during the last 8000 years. In response to the reconstructed cooling of the North Atlantic by 1.1 ± 0.2°C, a zonally averaged model of oceanic biogeochemistry simulates a drawdown of atmospheric CO_2 by 7 ± 0.8 ppmv, while a reconstructed warming of the Pacific Ocean by 0.6 ± 0.4°C counterbalances this effect by about 1 ppmv. On a global scale, this model simulates a lowering of atmospheric CO_2 from the Holocene to pre-industrial times by 6 ± 2 ppmv due to changes in SSTs, while more complex, three-dimensional biogeochemistry model indicates a moderate decrease by 1 ppmv after 300 years of the model integration. Our study suggests that changes in SSTs may have altered atmospheric CO_2 in a direction opposite to the observed trend and that other mechanisms, presumably related to the changes in carbonate chemistry, could be responsible for the CO_2 increase during the Holocene.

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1. Introduction

[2] Understanding the carbon cycle in the past is a key for a successful projection of the dynamics of atmospheric CO_2 into the future. What were the mechanisms that caused the increase in atmospheric CO_2 during the Holocene of about 20 parts per million by volume (ppmv), recorded in the Taylor Dome ice core [*Indermühle et al.*, 1999]? Interest in this question has been heated up by the hypothesis of *Ruddiman* [2003, 2005, 2006] that this CO_2 growth manifests the beginning of the profound influence that humans have had on climate. There is no doubt that anthropogenic land cover changes started long before the industrial era, but the magnitude of these changes and their climatic consequences is quite uncertain [*Houghton et al.*, 1983; *Klein Goldewijk*, 2001; *Brovkin et al.*, 2004; *Olofsson and Hickler*, 2008]. Meanwhile, there is a rationale to check the need to invoke an anthropogenic hypothesis by first testing hypotheses about natural forcings in the Holocene, especially ones that might be tested against available proxy data [*Claussen et al.*, 2005; *Crucifix et al.*, 2005; *Broecker and Stocker*, 2006; *Schurgers et al.*, 2006]. Several natural explanations for the Holocene CO_2 trend have been suggested, including release of terrestrial carbon and changes in SSTs [*Indermühle et al.*, 1999], carbonate compensation in the ocean [*Broecker et al.*, 1999a; *Broecker and Clark*, 2003], and coral reef growth [*Ridgwell et al.*, 2003].

[3] Here we focus on the effect of changes in SSTs on the atmospheric CO_2. *Indermühle et al.* [1999] applied inverse modeling and estimated an effect of a global SST increase of 0.5°C on atmospheric CO_2 of about 4–5 ppmv. *Brovkin et al.* [2002] have found no substantial effect of SSTs on the CO_2 in Holocene simulation, while model results by *Joos et al.* [2004] suggest that 5 ppmv of CO_2 growth during the Holocene could be caused by increasing SSTs. *Wang et al.* [2005] suggested some increase in CO_2 due to a small increase in SSTs (0.2°C on a global scale) during the Holocene as simulated by their climate model. This uncertainty regarding the effect of SSTs on CO_2 in simulations discussed above is due to differences in physical and biogeochemical models, as well as differences in boundary conditions. For example, *Brovkin et al.* [2002] did not account for the cooling effect of the Laurentide ice sheet on climate that completely disappeared only around 7000

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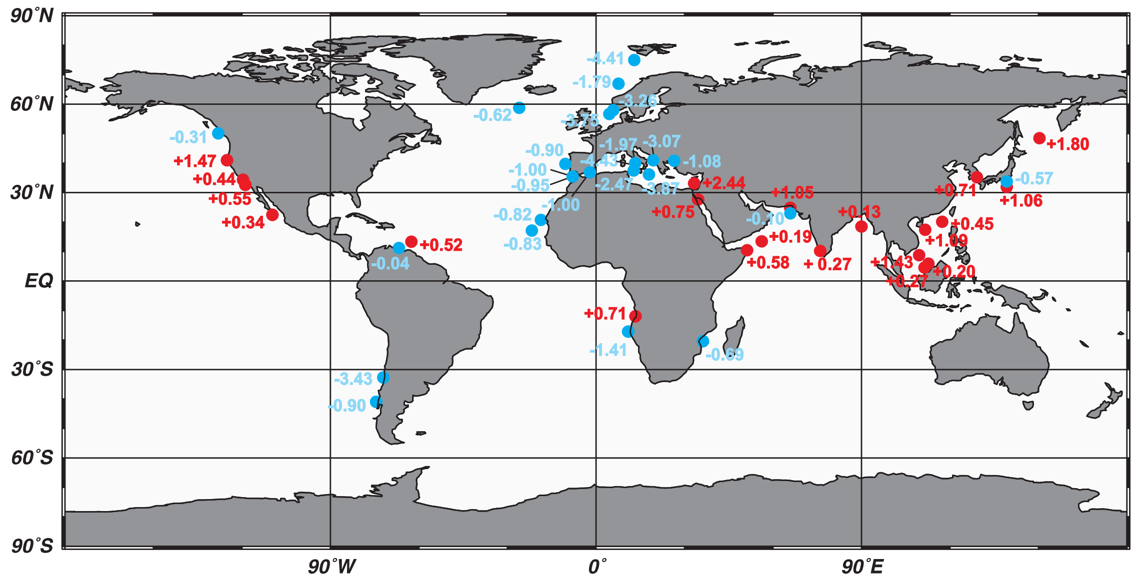

Figure 1. The spatial distribution of the linear trend in SST (°C/8000 years) reconstructed from the alkenone data [*Kim and Schneider*, 2004]. The blue and red colors of the labels indicate cooling and warming trends from 8,000 yr BP to the pre-industrial period, respectively.

years BP, while this factor was included in simulations by *Joos et al.* [2004] and *Wang et al.* [2005]. *Wang et al.* [2005] have not quantified the effect of ice sheet retreat on the atmospheric CO_2 because of the absence of an ocean carbon cycle model in their study. *Joos et al.* [2004] reported a 6 ppmv increase in atmospheric CO_2 in their sensitivity simulation with an ocean carbon model forced by an SST increase of 0.6°C between 7 and 6 ka BP in response to the ice sheet retreat. However, they noted that this result might be biased toward high values due to the simplicity of the model they used and concluded that the contribution of the SST changes to the Holocene CO_2 rise requires further investigations. Spatial resolution might be an important factor which affects the model response: the impulse response model of biogeochemistry applied by *Joos et al.* [2004] is driven by globally averaged SSTs changes, while the CLIMBER-2 model used by *Brovkin et al.* [2002] includes zonally averaged ocean model with 2.5° latitudinal resolution. The latter model accounts for difference in precessional forcing between tropical and extratropical regions that is pronounced for the Holocene [*Berger and Loutre*, 2004].

[4] Recently, *Kim et al.* [2004] reconstructed changes in SSTs during the Holocene based on an alkenone biomarker. This analysis reveals a rather complex pattern of SST changes during the Holocene. North Atlantic sites have a pronounced trend toward cooling, while North Pacific and northern subtropical and tropical sites indicate a warming trend. For the North Atlantic, their conclusions are in line with results by *Marchal et al.* [2002] on long-term cooling of the sea surface in the northeast Atlantic and Mediterranean during the Holocene. The assemblage of alkenone SST data in the GHOST data set [*Kim and Schneider*, 2004] provides globally averaged changes of 0.2 ± 0.2°C for the last 8000 years (Figure 1).

2. Methods

[5] Are these opposite trends between the North Atlantic and the tropics revealed in the proxy data supported by model simulations? What is the effect of reconstructed SST changes on atmospheric CO_2 concentration? To address these questions, we have used two Earth system models developed in PIK. The CLIMBER-2 model that includes interactive atmospheric and land surface modules, a zonally averaged model of ocean dynamics, and oceanic biogeochemistry [*Petoukhov et al.*, 2000; *Ganopolski et al.*, 2001; *Brovkin et al.*, 2002] has been used for transient simulations through the Holocene and in the sensitivity analysis. The coarse spatial resolution of the model (51° in longitude and 10° in latitude for the atmosphere, 2.5° meridional resolution and 20 unevenly distributed vertical levels for 3 ocean basins) allows integration of about 500 years per one hour of single processor time on the IBM p655 cluster. A more sophisticated model, CLIMBER-3α, consisting of interactive atmosphere, land surface, and 3-dimensional ocean models [*Montoya et al.*, 2005] has been applied in sensitivity experiments to evaluate the effect of reconstructed SSTs on atmospheric CO_2. While the horizontal resolution of the atmospheric model of 22.5° in longitude and 7.5° in latitude is rather coarse, the coupled sea-ice ocean model has a resolution of 3.75° × 3.75°. The ocean model consists of 24 vertical layers of 25 m in thickness at the top and 513 m at the bottom, continuously increasing with depth. CLIMBER-3α was coupled with the same type of NPZD model [*Six and Maier-Reimer*, 1996] as CLIMBER-2, however, in contrast to the latter, air-sea gas-exchange of

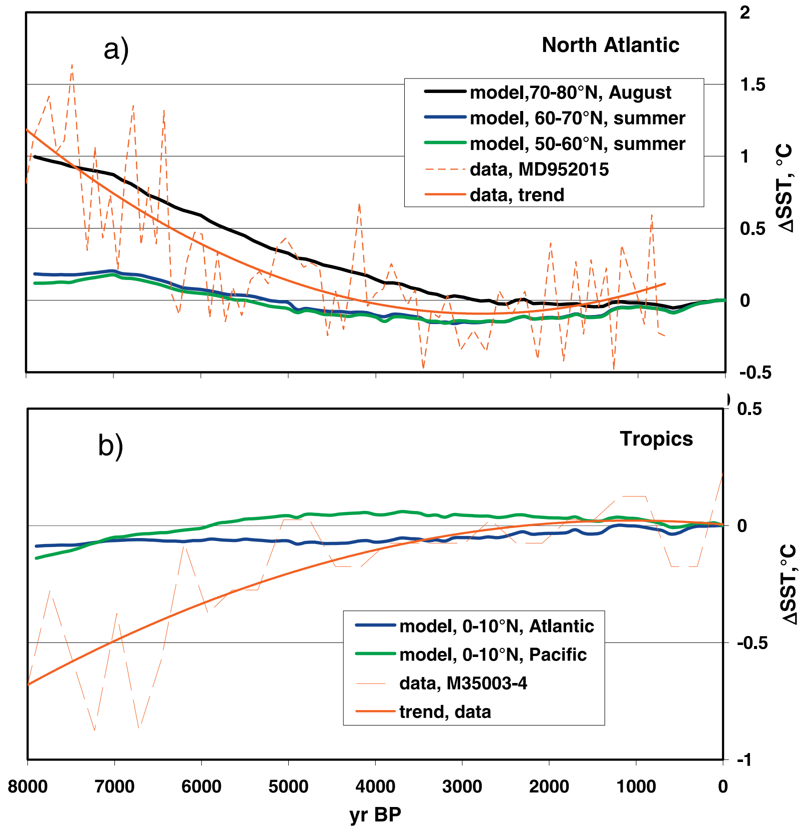

Figure 2. Model-data comparison of SST changes (°C) through the Holocene relative to the pre-industrial period. Model results are from the transient CLIMBER-2 simulation: (a) North Atlantic and (b) Tropical Ocean. The geographical coordinates of marine cores MD952015 and M35003-4 are (59°N, 26°W) and (12°N, 61°W), respectively.

CO_2 was treated as a quadratic function of the 10 m height wind speed [*Wanninkhof*, 1992] provided by the NCEP climatology [*Kalnay et al.*, 1996]. Since the CLIMBER-3α ocean model employs an advection scheme which is nearly free of numerical diffusion, it was used with low diapycnal background diffusivities of the order of 0.1 cm^2s^{-1}, notably within the permanent thermocline [*Hofmann and Maqueda*, 2006]. Computational efficiency of the CLIMBER-3α model with ocean carbon cycle is about 1 year per one hour of single processor time on the IBM p655 cluster. In this model version, inclusion of the oceanic biogeochemical tracers considerably (by a factor of 5) slowed down the performance of the physical ocean model, but it was necessary for calculation of response of oceanic biogeochemistry to the SST change.

3. Results

3.1. CLIMBER-2 Transient Simulation

[6] In the transient simulation TRAN of climate from 8000 years BP to pre-industrial times, the CLIMBER-2 model was driven by changes in orbital forcing [*Berger and Loutre*, 1991], ice sheet retreat [*Peltier*, 1994; *Joos et al.*, 2004], and observed atmospheric CO_2 concentration [*Indermühle et al.*, 1999]. The initial state was achieved after 10,000 years of model integration with boundary conditions for 8000 years BP. On a global scale, simulated SST increased from 8000 years BP to the pre-industrial period by 0.1°C; most of this warming occurred from 8000 to 7000 years BP in response to the decay of the Laurentide ice sheet. During the later period, a warming effect from increasing atmospheric CO_2 concentration counteracts cooling due to declining summer insolation in the northern high latitudes. Both CO_2 and precessional forcings are amplified by vegetation-albedo and sea ice-albedo feedbacks [*Claussen et al.*, 2006]. In the high latitude regions, the biomarkers such as alkenones are proxies for temperature during the summer when organic production is taking place. In general, CLIMBER-2 shows changes in SSTs of a smaller scale than in the alkenone SST reconstructions. In the North Atlantic, simulated temperatures in August (mid-summer) for the region 70–80°N decline by 1°C; the greatest change we identified in the period 8000 to 3000 years BP (Figure 2a). At latitudes 50–70°N, a cooling trend is visible but the amplitude of changes is smaller (from 0.2°C at 8000 years BP to −0.1°C at 3000 years BP). Data for marine core MD952015 (59°N, 26°W) from the open ocean support a decrease of SSTs by about 1.2°C from 8000 to 3500 years BP and some tendency for later warming (Figure 2a). In the tropical region of 0–10°N, the model simulates a warming trend in SST by about 0.1°C and 0.2°C for the Atlantic and

Table 1. Changes in SSTs Used to Force the Biogeochemistry Models[a]

Latitude	ΔSST, °C		
	Atlantic	Indian	Pacific
70–80°N	−1.3 ± 0.1		
60–70°N	−1.5 ± 0.1		
50–60°N	−1.4 ± 0.1		0.7 ± 0.4
40–50°N	−0.9 ± 0.5		0.7 ± 0.4
30–40°N	−0.8 ± 0.2		0.6 ± 0.7
20–30°N	−0.9 ± 0.1	0.6 ± 0.2	0.4 ± 0.2
10–20°N	0.2 ± 0.2	0.3 ± 0.1	0.7 ± 1.0
0–10°N	0.2 ± 0.2	0.3 ± 0.1	0.9 ± 0.1
0–10°S	0.2 ± 0.2	0.3 ± 0.1	0.9 ± 0.1
10–20°S	−0.2 ± 0.1	0.3 ± 0.1	0.9 ± 0.1
20–30°S	−0.2 ± 0.1	−0.1 ± 0.3	0.9 ± 0.1

[a]GHOST data [*Kim and Schneider*, 2004], shown in Figure 1, are seasonally adjusted for latitudes 50–80°N in the Atlantic and substituted with nearest basin data in cases where no data is available for the latitudinal belt. SST changes in the Arctic basin and to the south of 30°S were neglected. The measurement error is for a 95% confidence interval.

the Pacific, respectively (Figure 2b). In the other latitudes in the zone 30°S to 30°N, the behavior of SSTs is similar (not shown). A marine core M35003-4 (12°N, 61°W) from the Caribbean basin demonstrates warming of 0.7°C, which is stronger than changes simulated by the model but goes in the same direction.

[7] Simulated SSTs in the North Atlantic decline during the Holocene, and this cooling trend is reversed toward warming in the tropical region. As already discussed by *Marchal et al.* [2002] and *Lorenz and Lohmann* [2004] these trends could be explained by the orbital forcing, and an increase in atmospheric CO_2 concentration may play an additional role that counterbalanced orbital forcing in the North Atlantic and amplified it in the tropics. In the North Pacific, CLIMBER-2 simulations do not confirm the warming trend reconstructed from alkenone records by *Kim et al.* [2004]. This could be explained either by limitations of the model or by data limitations. First, the CLIMBER-2 model does not simulate planetary waves and zonal oscillation patterns that might be responsible for the Northeast Atlantic-Northeast Pacific dipole (see discussion below). Second, alkenone records analyzed by *Kim et al.* [2004] were collected for coastal regions that might not be representative for the open ocean areas.

[8] *Kim et al.* [2004] suggested that a dipole in temperature changes between the Northeast Atlantic (cooling) and the Northeast Pacific (warming) could be caused by an interaction between positive Pacific North American (PNA) and negative North Atlantic Oscillation (NAO) phases of the atmospheric circulation. This hypothesis was supported by simulations performed with the ECHO-G model [*Lorenz and Lohmann*, 2004]. ECHO-G is a general circulation model that in particular simulates atmospheric planetary waves responsible for oscillation patterns and teleconnections, while CLIMBER-2 has a statistical-dynamical atmospheric module that simulates average atmospheric dynamics and neglects planetary waves. Therefore the hypothesis on the suggested mechanism of teleconnection cannot be tested with CLIMBER-2, which simulates a cooling trend in the North Pacific similar to the North Atlantic.

3.2. CLIMBER-2 Simulations With SST Forcing

[9] The SST changes presented in Figure 1 are site-dependent, while for using as a forcing in numerical experiments, they should be spatially and temporally aggregated. To account for the coarse resolution of the CLIMBER-2 model, the values were averaged over 10° latitudinal belts for the Atlantic, Indian, and Pacific basins. The Arctic region and the Southern Ocean to the south of 30°S were excluded from the analysis because of the lack of data. The values for the other areas with absent data, e.g., the tropical Pacific, were taken from the adjusted neighboring sectors for the same basin. To correct for the seasonality in sensitivity simulations, we reduced the magnitude of SST changes for the Atlantic region 50–80°N in accordance with the ratio of summer to annual temperature changes in the transient simulation TRAN. Because of this seasonality adjustment, we assume that estimates of SST changes for the region and their effect on SSTs are conservative. Resulting SST changes with a 95% confidence interval are listed in Table 1. Let us note that this confidence interval reflects only the technical measurement error and not the uncertainty of the spatial extrapolation of the site data to the large areas. In fact, Table 1 gives only a very first approximation of the global SST changes because almost all reconstructed summer SSTs are confined to the coastal regions.

[10] What are the consequences of the reconstructed SST changes for the atmospheric CO_2? In the global simulation GLOB, changes in SSTs listed in Table 1 were added as anomalies to the SSTs simulated by the CLIMBER-2 physical model used to force the CLIMBER-2 oceanic biogeochemical model while physical climate model components were driven by pre-industrial initial condition so that neither changes in SSTs nor CO_2 influence the climate. Atmospheric CO_2 was interactive between ocean and atmosphere and the model was integrated for 2000 years to achieve a new steady state. As a result, the atmospheric CO_2 concentration was lowered by 6 ± 2 ppmv. Assuming that circulation patterns and marine biology did not change substantially during the Holocene, this CO_2 decrease could be interpreted as the effect of the reconstructed SST trend on atmospheric CO_2 during the Holocene. To get a deeper insight into the effect of regional SST changes on CO_2, additional simulations have been performed in which SSTs were changed only in the North Atlantic (30°–80°N), North Pacific (30°–60°N), Tropical Atlantic (30°S–30°N), and Tropical Pacific (30°S–30°N). A spatial distribution of SST forcing averaged over these regions and the response of atmospheric CO_2 is presented in Figure 3 in schematic form. Driven by SST changes for the North Atlantic, the model reveals a decrease in atmospheric CO_2 by 7 ± 0.8 ppmv. The SST changes in the other ocean regions had a much less pronounced effect on atmospheric CO_2: it increased by 0.3 ± 0.2 and 0.4 ± 0.2 ppmv in simulations for the North Pacific and North Atlantic, respectively. The effect of SST changes in the tropical Atlantic was insignificant (0 ± 0.2 ppmv). This suggests that regarding the effect on

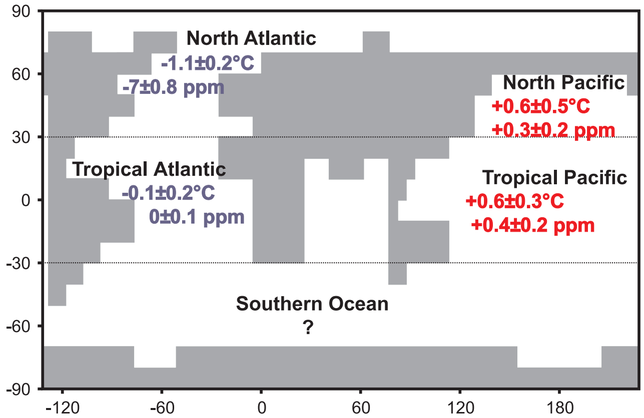

Figure 3. A sketch showing the effect of SST changes in Table 1 on atmospheric CO_2 for different ocean regions during the last 8000 years obtained in CLIMBER-2 sensitivity simulations (see text for details).

atmospheric CO_2, the cooling in the North Atlantic dominates over the warming in the other oceanic regions.

[11] SSTs in the Southern Ocean were unchanged in the GLOB simulation. Proxy data for Holocene SST changes to the south of 30°S are very rare. Two alkenone-based reconstructions off Southern Chile [*Lamy et al.*, 2002; *Kaiser et al.*, 2005] suggest a cooling trend of 0.8°C during the last 8000 years (see Figure 1). Analysis of δD and δ^{18}O changes in the Antarctic ice cores reveals a widespread Antarctic early Holocene optimum between 11,500 and 9000 years BP, a secondary optimum later in the Holocene, and consequent cooling which was pronounced not only for the Antarctic but also for the surrounding Subantarctic Ocean [*Masson et al.*, 2000; *Masson-Delmotte et al.*, 2004]. Analyzing ice-rafted debris and microfossils in a piston core at 53°S in the South Atlantic, *Hodell et al.* [2001] found that Antarctic surface waters were warm and ice free between 10,000 and 5000 years BP, and that this warm period was followed by abrupt sea surface cooling and sea ice advancing after 5000 years BP. In contrast to these studies, a diatom-based SST reconstruction by *Nielsen et al.* [2004] suggested much less pronounced early Holocene warming (12,500–6200 years BP) but a late Holocene warming (2900 years BP to the present) of about 2°C at the site located at the Polar Front (50°S, 6°E). In the transient Holocene simulation of a climate model of intermediate complexity, *Renssen et al.* [2005] found that over the Southern Ocean, the early and mid-Holocene temperatures were higher than at present in all seasons. These results are in line with the modeling study by *Lorenz et al.* [2006] who reported continuous cooling in the southern high latitudes throughout the Holocene, albeit accompanied by some warming in the southern midlatitudes.

[12] Proxy data for the Southern Ocean are sparse and equivocal; however, the Southern Ocean plays a substantial role in the global carbon cycle. To test the sensitivity of atmospheric CO_2 to the changes in the Southern Ocean, additional CLIMBER-2 equilibrium simulations were performed with changes of the surface ocean temperatures to the south of 30°S. The atmospheric CO_2 increased (decreased) by 2 ppmv in response to warming (cooling) of SSTs by 0.5°C, respectively. This is a significant sensitivity of about 4 ppmv/°C, but it is less than the 7 ppmv/°C sensitivity for the North Atlantic (see Figure 3). Ambiguity in the SST reconstructions for the Southern Ocean reported above does not allow a quantification of the CO_2 response to the SST changes in the region.

[13] Because reconstructed SST data are unevenly distributed and mostly confined to the coastal regions, it is rationale to estimate an effect of simulated SST changes on the atmospheric CO_2. The differences in the SSTs between the end and the beginning of the TRAN simulation were added as anomalies to the present-day SSTs and used to force the biogeochemical model of CLIMBER-2 as in the GLOB simulation. After 2000 years of the model integration, the atmospheric CO_2 remained unchanged, in line with the previous CLIMBER-2 Holocene simulations [*Brovkin et al.*, 2002]. This is explained by much lower amplitude of simulated SST changes in comparison with the reconstructions. For example, simulated annually averaged changes in the North Atlantic for the zone 60–70° are −0.2°C while data reveal much stronger cooling of 1.5 ± 0.1°C. This highlights an importance of the North Atlantic temperature reconstructions for evaluation of the effect of Holocene SSTs on the atmospheric CO_2.

3.3. CLIMBER-3α Simulations With SST Forcing

[14] CLIMBER-2 model possesses a simplified, zonally averaged oceanic module. How different are its results from the response of a three-dimensional model? We performed simulation GLOB with the CLIMBER-3α model. For comparison with CLIMBER-2, the SST forcing was kept in the same, basin-averaged zonal form and the mean SST values given in Table 1 were applied simultaneously to the all oceanic regions. After a 1500-year spinup of the ocean physics, the oceanic biogeochemistry was switched on in CLIMBER-3α and integrated for a further 500 years. During this spinup, the atmospheric CO_2 rose to a level of 288.7 ppmv, with an upward trend of about 0.3 ppmv per decade. The carbon cycle spinup was followed by a further integration over 300 years using the SST forcing applied in the carbon cycle model at the year 35 of the simulation. To account for the CO_2 drift, a 300-year control simulation without Holocene SST forcing was performed as well. The CO_2 drift from the carbon cycle spinup continued during both experiment and control simulations, but a divergence of two trends after 150 years is clearly visible on Figure 4a. CLIMBER-3α does not possess year-to-year variability [*Montoya et al.*, 2005], and that makes even relatively small changes in sensitivity simulations statistically significant. The difference in atmospheric CO_2 between the experiment and control is shown in Figure 4b. In response to the SST changes, the atmospheric CO_2 difference rose abruptly by 0.3 ppmv, presumably in response to a warming in the large Pacific and Indian basins, and then started to decline due to cooling in the Atlantic. By the end of the experiment, the

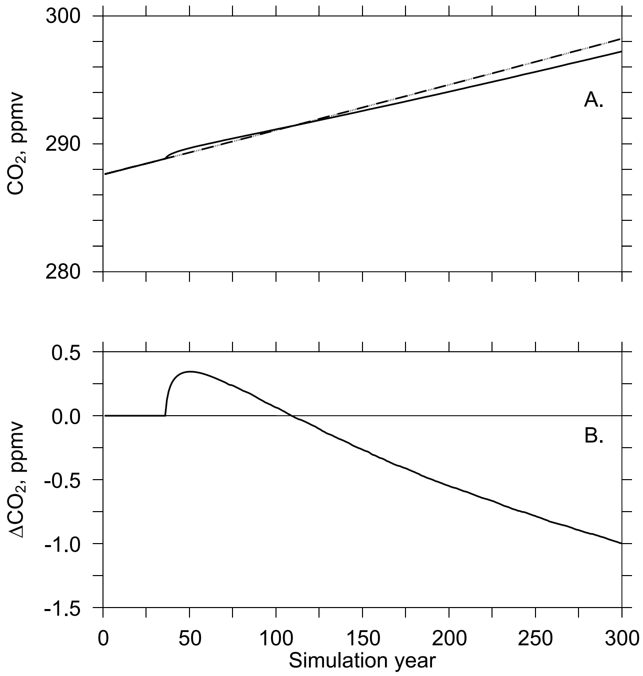

Figure 4. Dynamics of annual mean sea surface CO_2 (ppmv) simulated by CLIMBER-3α. (A) CO_2 dynamics in sensitivity experiment with SST changes (solid line) and control simulation (dashed line). (B) A difference in CO_2 between sensitivity and control simulations. The SST forcing was applied after first 35 simulation years.

atmospheric CO_2 difference was 1 ppmv and continued to grow, indicating a further possible decrease in the SST experiment. The spatial distribution of annual mean sea-surface CO_2 difference between the sensitivity and control simulations at the end of integration (year 300) is shown in Figure 5. While CO_2 in the Pacific and Indian basins increased by 5 to 15 ppmv, it decreased by 10 to 20 ppmv in the North Atlantic, in particular in the regions where North Atlantic Deep Waters (NADW) are formed. The latter explains why – although regions with CO_2 increase have a larger area than areas with CO_2 decrease – the model reveals an averaged decrease in atmospheric CO_2, in line with the CLIMBER-2 simulations. Although the amplitude of the increase of the annual mean sea-surface pCO_2 over the Pacific and Indian basin is much higher in the simulation using CLIMBER-3α than in CLIMBER-2, the global net effect in applying reconstructed SST anomalies is very small and tends in the same direction.

[15] Additional CLIMBER-3α simulations were performed to evaluate the response of atmospheric CO_2 to changes of the surface ocean temperatures to the south of 30°S. After 120 years of model integration, atmospheric CO_2 increased (decreased) by 2 ppmv in response to warming (cooling) of SSTs by 0.5°C, respectively (not shown). Because CO_2 changes were still growing after 120 years (although with a tendency toward saturation), these results indicate a stronger sensitivity of CLIMBER-3α to SST changes in the Southern ocean in comparison with CLIMBER-2. However, we were not able to quantify the effect of the Southern ocean SSTs on atmospheric CO_2 because of uncertainty in the direction and magnitude of

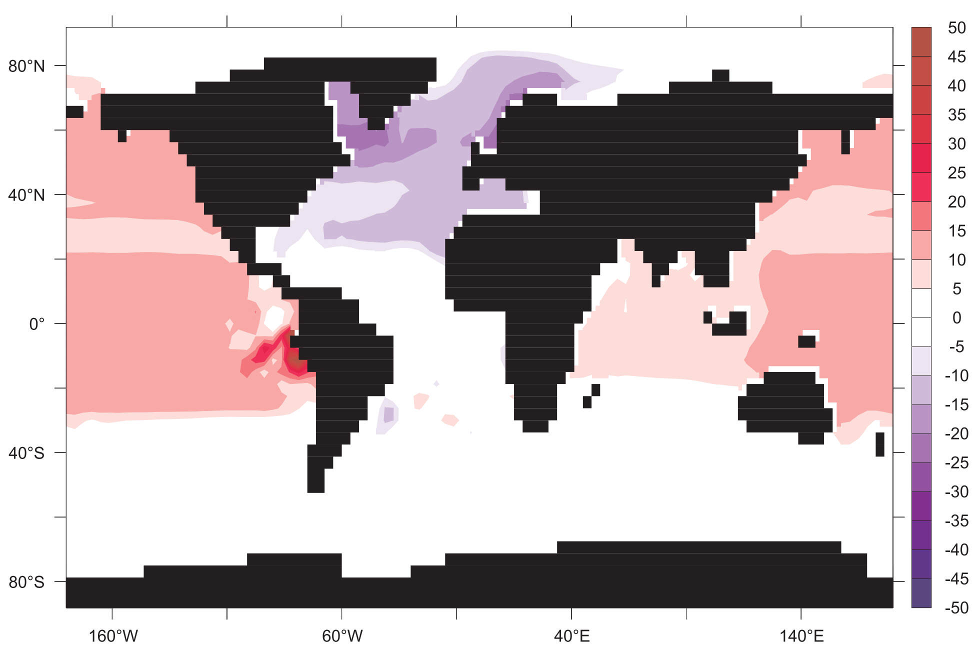

Figure 5. Annual mean sea-surface CO_2 difference (ppmv) simulated by CLIMBER-3α in response to the SST changes in Table 1. Shown is the difference between sensitivity and control simulations after 300-year integration.

reconstructed SST changes (see discussion of the CLIMBER-2 results above).

4. Discussion and Conclusions

[16] We have performed several additional tests with CLIMBER-2. The biogeochemistry model of CLIMBER-2 has a sensitivity of atmospheric CO_2 of 11 ppmv/°C for present-day circulation (neglecting changes in land carbon and carbonate compensation). This is close to the 9–10 ppmv/°C equilibrium sensitivity suggested by *Bacastow* [1996] and *Archer et al.* [2004]. A Harvardton-Bear Equilibration Index (HBEI) has been suggested to measure the effect of low latitudes on atmospheric CO_2. The index is a ratio of equilibrium changes to instantaneous changes in CO_2 after perturbation of the warm low-latitude surface ocean. *Broecker et al.* [1999b] have showed that box models typically possess low values of HBEI (0.1–0.2), corresponding to a low significance of warm ocean for atmospheric CO_2, while three-dimensional models reveal higher values on the index (0.2–0.4), suggesting a greater exchange between warm and cold waters than in box models. To evaluate the HBEI, the CLIMBER-2 model performed a 2000-year simulation of a 6°C cooling of the surface ocean between 40°S and 40°N and calculated changes in the surface CO_2. The resulting HBEI value of 0.33 suggests that the CLIMBER-2 model, as a 2-dimensional model that explicitly simulated thermohaline circulation, is more similar to the 3-dimensional ocean general circulation models than to the box models. *Broecker et al.* [1999b] reported that the Bern 2-D model that is similar to CLIMBER-2 has a HBEI value of 0.14, while *Ridgwell* [2001] has found a HBEI value of 0.39 for a similar 2-D model. As suggested by *Ridgwell* [2001], the representation of convective mixing is a primary factor responsible for the variability in values reported by *Broecker et al.* [1999b].

[17] Transient Holocene simulations with the CLIMBER-2 model suggested that the SSTs in the North Atlantic declined during the Holocene, and that this cooling trend was reversed toward warming in the tropical region. These results support the views of *Marchal et al.* [2002] and *Kim et al.* [2004] that the Holocene SST dynamics could be explained by orbital forcing changes, and an increase in atmospheric CO_2 concentration may play an additional role that counterbalanced orbital forcing in northern temperate and subtropical latitudes (see comments above). In the North Pacific, CLIMBER-2 simulations do not reveal the warming trend observed in alkenone records. Presumably, the simplicity of the atmospheric dynamics module prevents the model from simulating a dipole in temperature changes between the Atlantic and Pacific basins due to the interaction between positive PNA and negative NAO phases of the atmospheric circulation [*Kim et al.*, 2004; *Lorenz and Lohmann*, 2004]. *Lorenz et al.* [2006] argued that the warming in the tropics is due to the winter insolation increase in the tropics while the cooling in the North Atlantic is due to the summer insolation decrease. The increase of CO_2 might have increased SST everywhere. Therefore a CO_2-induced warming in the North Atlantic is overprinted by a stronger cooling by orbital forcing but amplified by a warming in the tropics.

[18] The CO_2 lowering by 1 ppmv in CLIMBER-3α is much less than the decrease by 6 ± 2 ppmv obtained in the CLIMBER-2 simulations. One of the reasons for this difference is that a 300-year simulation with CLIMBER-3α is not long enough to come to equilibrium with the new boundary conditions. In the CLIMBER-2 simulations, the decrease in CO_2 after 300 years (3 ppmv) is only about a half of the equilibrium response after 2000 years. The same temporal dynamics could be valid for CLIMBER-3α. As opposed to CLIMBER-2, CLIMBER-3α uses a variable air-sea gas-exchange parameterization [*Wanninkhof*, 1992] and low vertical diffusivities [*Hofmann and Maqueda*, 2006], which might also lead to a less sensitive model behavior.

[19] Reconstructed SST changes used to force the CLIMBER-2 model are mostly based on the data from the coastal regions (Figure 1). This is a strong limitation to our approach and a possible source of significant uncertainty which is difficult to quantify. Besides, the Southern Ocean is excluded from our study because of lack of data. Nonetheless, the qualitative agreement between SSTs simulated by CLIMBER-2 and the ECHO-G model [*Lorenz et al.*, 2006] is remarkable. A cooling trend in the North Atlantic and tropical warming are general features of both reconstructed and simulated SST changes. These two trends have counteracting effects on the atmospheric CO_2.

[20] Finally, simulations of oceanic biogeochemistry with the CLIMBER-2 and CLIMBER-3α models suggested that the Holocene SST changes reconstructed by *Kim and Schneider* [2004] led to a decrease in atmospheric CO_2. The difference in the magnitude of CO_2 decrease, 6 ± 2 ppmv and 1 ppmv for the CLIMBER-2 and CLIMBER-3α models respectively, could be explained by differences in model parameterization and simulation length. Our study suggests that changes in SSTs may have altered atmospheric CO_2 in a direction opposite to the observed trend and that other mechanisms, presumably related to the changes in carbonate chemistry, could be responsible for the CO_2 increase during the Holocene.

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