Holocene climate and carbon cycle dynamics: Experiments with the “green” McGill Paleoclimate Model

Yi Wang
Earth System Modelling Group, Department of Atmospheric and Oceanic Sciences, McGill University, Montreal, Quebec, Canada, H3A 2K6

Lawrence A. Mysak
Earth System Modelling Group, Department of Atmospheric and Oceanic Sciences, McGill University, Montreal, Quebec, Canada, H3A 2K6

Nigel T. Roulet
Department of Geography and McGill School of Environment, McGill University, Montreal, Quebec, Canada H3A 2A7

Abstract. An inverse method is used to investigate the global carbon cycle from the early Holocene (8 kyr BP) to the end of the pre-industrial period (0 kyr BP) in an improved version of the “green” McGill Paleoclimate Model (MPM). In this paper, we now take into account the vegetation-precipitation feedback and evaluate the terrestrial carbon cycle for the pre-industrial equilibrium. From our coupled transient simulation under orbital forcing, reconstructed (Taylor Dome) atmospheric $\text{CO}_2$ forcing and a prescribed retreating Laurentide Ice Sheet (LIS), we find a decrease of 70 PgC in total carbon storage in the Sahara region ($15^\circ$N to $30^\circ$N and $15^\circ$W to $50^\circ$E), which is caused by the desertification simulated in the green MPM. The above decrease is partially compensated by an increase of 40 PgC in total carbon storage in the Southern Hemisphere from 8 to 2 kyr BP. From an analysis of the total carbon stored under the ice sheet, we can infer that this carbon has negligible impact on atmospheric $\text{CO}_2$ after 8 kyr BP. From our model results, we further conclude that the retreating LIS, together with the vegetation-albedo feedback, cause the global terrestrial carbon to increase from 8 to 6 kyr BP. The application of the inverse method suggests that the first 10 ppmv increase in atmospheric $\text{CO}_2$ from 8 to 6 kyr BP comes from the ocean. Finally, in the model simulations, the total terrestrial carbon release from 6 to 0 kyr BP is about 68 to 95 PgC, which would produce about a 5 to 7 ppmv atmospheric $\text{CO}_2$ increase, based on the calculation of Joos et al. [2004]. Due to our model limitation (there is no ocean carbon cycle), we cannot conclude whether the overall oceanic $\text{CO}_2$ release from 8 to 0 kyr BP is due to outgassing related to SST changes or to calcite compensation as proposed by Broecker et al. [2001].

1. Introduction

In a recent paper [Wang et al., 2005b; hereafter referred to as WMWB], the “green” McGill Paleoclimate Model (MPM) was used to investigate the biogeophysical roles of vegetation dynamics and a receding Laurentian Ice Sheet (LIS) on the evolution of the Holocene climate up to eight thousand years before present (8 kyr BP), under variable Milankovitch forcing according to Berger [1978] and a constant atmospheric $\text{CO}_2$ concentration of 280 ppmv. The main goal of this paper is to extend the work of WMWB on the simulation of the Holocene climate by including: (1) the vegetation-precipitation feedback, where evapotranspiration from the terrestrial vegetation is the critical link; and (2) a prescribed but variable atmospheric $\text{CO}_2$ forcing on ecosystem production based on the Taylor Dome record [Indermühle et al., 1999], as shown in Fig. 1.

The green MPM, apart from the vegetation module,
(VECODE, see Brovkin et al., 2002, and the references therein), consists of a coupled atmosphere-ocean-sea ice
land surface Earth system Model of Intermediate Complexity (EMIC, see Claussen et al., 2002) that is secto-
rially averaged in the zonal direction (Wang and Mysak, 2000). The ocean component simulates the global ther-
mosaline circulation, and the atmosphere is represented by an energy-moisture balance model with prescribed
winds. In VECODE, the carbon stored in live vegetation is partitioned into two pools, and two pools are used to
represent soil organic matter (SOM). We modify VECODE to allow for a slowly accumulating store of recalc-
itrant carbon in the soil. Also in this paper we evaluate, for the first time, the pre-industrial terrestrial carbon cy-
cle in the green MPM by comparing our simulated stores with the estimates of Schimel et al. [1994], Schimel [1995]
and Falloon et al. [1998]. The novel features of this paper are: (1) prescription of a slow retreat of the LIS from 8
to 6 kyr BP (see Figure 14-2 in Ruddiman, 2000), which has an impact on the Holocene vegetation dynamics and
terrestrial carbon cycle, especially in northern high lat-
itudes; (2) consideration of both vegetation-albedo and
vegetation-precipitation feedbacks (see Box 2-6 in Rud-
diman, 2000) in our transient simulation, which play im-
portant roles in the desertification of northern Africa;
and (3) use of a different physical climate model (the
green MPM) than in Brovkin et al. [2002] and Joos et
al. [2004]. Other special aspects of the paper are: (1)
introduction of an inverse method (see Sect. 2.3 for more
details) to close the global carbon cycle by employing the
conservation of global carbon according to the assump-
tions in Brovkin et al. [2002] and Joos et al. [2004]; (2)
disregard of the one-way transport of carbon from land
to ocean; and (3) neglect of the buildup and decay of
peatlands (wetlands) carbon. These aspects allow us to
analyze the atmospheric carbon sources throughout the
Holocene up to 8 kyr BP. Assumptions similar to the lat-
ter two above were also made in Brovkin et al. [2002] and
Joos et al. [2004].

Several explanations for the Holocene atmospheric
$\mathrm{CO}_2$ change ($\approx 20$ ppmv increase from 8 kyr to 390 yr
BP) have been hypothesized recently [Indermühle et al.,
1999; Broecker et al., 1999, 2001; Broecker and Clark,
2003; Ruddiman, 2003; Ridgwell et al., 2003]. Indermühle
et al. [1999] suggest that most of the variability in at-
mospheric $\mathrm{CO}_2$ concentration is caused by changes in
the amount of terrestrial biomass and sea surface tem-
perature (SST). From an inverse model, they estimate
a terrestrial uptake of 110 PgC in the early Holocene,
followed by a release of 195 PgC between 7 and 1 kyr
BP. They conclude that the global annual mean SST in-
creased about 0.5 °C between 9 and 6 kyr BP. Broecker
et al. [2001] propose that the 20 ppmv rise in atmos-
pheric $\mathrm{CO}_2$ concentration over the last 8 kyr was, in
part, due to the 500 PgC increase in terrestrial biomass
early in the present interglacial. Broecker et al. [2001]
argue that the $\mathrm{CO}_2$ released from the ocean-atmosphere
reservoir and used for the early regrowth of terrestrial
biosphere would have led to an increase in the carbon-
ate ion ($[\mathrm{CO}_3^{2-}]$) concentration in the deep ocean. As a
consequence, the lysocline would have deepened and the
oceanic $\mathrm{CaCO}_3$ budget would have become imbalanced,
i.e., the supply of $\mathrm{CaCO}_3$ by weathering discharge would
have been temporarily smaller than the loss by sedimen-
tation. This imbalance causes the surface ocean partial
pressure of $\mathrm{CO}_2$ to be larger than that of the atmospheric
$\mathrm{CO}_2$, until a new equilibrium is reached. The estimated
time for such an adjustment to occur is on the order of
5,000 years [Broecker and Peng, 1987; Sundquist, 1990;
More recently, Ruddiman [2003] proposes that the Holocene atmospheric CO$_2$ increase is due to a large release of carbon by changes in land use before the beginning of industrialization at about 1800 AD. His hypothesis is based on: (1) the existence of cyclic variations in atmospheric CO$_2$ driven by Earth orbital changes during the last 350,000 years that predict decreases, rather than increases, throughout the Holocene; and (2) a wide array of archaeological, cultural, historical and geologic evidence that points to possible sources due to anthropogenic land-use changes. Finally, Ridgwell et al. [2003], following the work of Berger [1982a, b], suggest that the buildup of coral reefs during the past 8 kyr could explain the atmospheric CO$_2$ increase.

The reconstructed CO$_2$ increase during the Holocene has also been studied by the paleoclimate modelling community. From an analysis of the transient runs of the CLIMBER-2 model that includes modules for the oceanic and terrestrial carbon cycles, Broeckin et al. [2002] find a higher land carbon store of 90 PgC at 8 kyr BP, compared to the pre-industrial value. To explain the atmospheric CO$_2$ rise, they need an additional assumption of excessive CaCO$_3$ sedimentation in the ocean. Joos et al. [2004] recently couple the carbon component of the Bern Carbon Cycle Climate (Bern CC) model with the Lund-Potsdam-Jena (LPJ) Dynamic Global Vegetation Model (DGVM). These model components are driven by climate fields from time-slice simulations of the past 21 kyr carried out using (1) the Hadley Centre Unified Model [Pope et al., 2000], or (2) the National Center for Atmospheric Research (NCAR) Climate System Model [Boville and Gent, 1998]. Their results suggest that a range of mechanisms, including calcite compensation in response to earlier terrestrial uptake (during the last glacial-interglacial transition), terrestrial carbon uptake and release (through the Holocene), SST changes, and coral reef buildup contributed to the Holocene atmospheric CO$_2$ increase.

The remainder of this paper is structured as follows. In Sect. 2, the model extension, experimental design, and the inverse method are described. Four scenarios for the carbon stored beneath an ice sheet are proposed and evaluated in Sect. 3. In Sect. 4, five snapshot (equilibrium) simulations are presented and the distribution of carbon store is evaluated using published estimates. In Sect. 5, the results of four transient simulations are shown and compared with those in other model studies. A summary and some concluding remarks are given in Sect. 6.

### 2. Model extension and experimental design

#### 2.1. Model extension and description

The land surface component in the green MPM as described in Wang et al. [2005a] has been extended to include the vegetation-precipitation feedback by taking into account the evapotranspiration processes associated with the terrestrial vegetation. The model we use for these processes is structurally similar to MOSES 2 [Cox et al., 1999, Eqs. (1)-(4)] and SLand [Hales et al., 2004, Eq. (6)]. In a region without vegetation cover, evapotranspiration ($E_L$) is given by Eq. (17) in Wang and Mysak [2000]. In a region with vegetation, evapotranspiration is given by the following equation:

Archer et al., 1997].
where $w_l$ is the soil moisture content, $w_k$ is a critical value given by $w_k = 0.75 w_m$, in which $w_m$ is the field capacity of moisture (0.15 m) [Manabe 1969], $q$ is the specific humidity, and $q_{sat}(T_l)$ is the saturated specific humidity at the temperature of the land surface ($T_l$). The quantities $E_l$ (evaporation) and $E_v$ (transpiration) in (1) are given by the following two equations:

\[ E_l = \min(1, w_l/w_k)(E_l + E_v)[q_{sat}(T_l) - q], \]  

(1)

where $\rho_a$ is the density of air, $C_E$ is the Dalton number ($1.3 \times 10^{-3}$) and $U_l$ is the zonally averaged wind over land from Oberhuber [1988], and

\[ E_v = Z_v S_{max} Z_w^{1/4} (1 - e^{-C_k Z_{lai}})/C_k, \]  

(3)

where $Z_v$ is the zonally averaged total vegetation fraction, $S_{max}$ is the maximum value of surface conductance [Hales et al., 2004], $Z_w$ is the zonally averaged soil moisture content, $C_k = 0.75$ is the nondimensional extinction coefficient, and $Z_{lai}$ is the zonally averaged leaf area index.

After the implementation of evapotranspiration in the green MPM, the precipitation in the equatorial region increases slightly. The vegetation distribution and surface air temperature (SAT) field in this region are also improved (figures not shown). These improvements are not only due to changes in the hydrological cycle, but also due to the commensurate changes in the latent heat fluxes between the land and the overlying atmosphere, which yield a better surface energy balance.

Since the green MPM is fully described in Wang et al. [2005a,b], here we only briefly describe the terrestrial carbon cycle module. In VECODE, the carbon in the live vegetation is partitioned into two compartments: a “fast” pool of green biomass (B1) that includes mainly leaves; and a “slow” pool of structural biomass (B2) that includes stems and roots. Dead organic matter in the soil is also divided into a “fast” compartment (B3) that consists mainly of woody residue; and a “slow” compartment (B4) that consists of humus (see Brovkin et al. [2002] for more details). We add recalcitrant organic carbon into the latter “slow” SOM carbon pool (B4), which becomes an important compartment in SOM due to its long turnover time (thousands of years). We do this by increasing the residence time of the “slow” SOM carbon pool in VECODE. Net primary productivity (NPP) is simulated annually, based on the semi-empirical parameterization of Lieth [1975]. The fertilization effect of atmospheric CO$_2$ concentration on plant growth is taken into account by the method of den Elzen et al. [1995, as cited in Brovkin et al., 2002].

2.2. Details of experimental design

The atmosphere, improved land surface, and sea ice components of the MPM together with VECODE are first spun up under the forcing of: (1) present-day monthly mean insolation radiation at the top of the atmosphere as derived from Berger [1978]; (2) the zonally averaged monthly SST climatology from Levitus [1982]; and (3) a fixed atmospheric CO$_2$ concentration of 280 ppmv. This set-up is intended to represent the pre-industrial
forcing conditions at approximately 1800 AD. A quasi-equilibrium state for these four components is attained after a 60-year integration.

The “pre-industrial” equilibrium run (EQ-0k in Table 1) is next performed by coupling the ocean component to the aforementioned four components using flux adjustments [Wang and Mysak, 2000]. The equilibrium state is reached after a 5-kyr integration. Two other equilibrium runs (EQ-6k and EQ-8k in Table 1) are carried out as above, but with solar forcing and a prescribed residue of the LIS (see Fig. 1 in WMWB) set at the values valid for 6 and 8 kyr BP. The reconstructed atmospheric CO$_2$ concentration from Taylor Dome at 6 and 8 kyr BP (Fig. 1) are used for these simulations. Another two equilibrium runs (EQ-0k-CO$_2$ and EQ-8k-clm, see Table 1) are also carried out, in which the atmospheric CO$_2$ concentrations for the terrestrial carbon cycle module are prescribed at 260 and 280 ppmv, respectively. We did these latter two experiments to test the sensitivity of the terrestrial carbon cycle to the atmospheric CO$_2$ concentration (i.e., (EQ-0k-CO$_2$)-(EQ-0k) = impact of CO$_2$) and the climate (i.e., (EQ-8k-clm)-(EQ-0k) = impact of climate) in the green MPM.

To set up the model for the four transient runs (see last four lines in Table 1), the green MPM is first spun up for 5060 years to reach an equilibrium state as in the EQ-8k run described above, except in the experiment AOV$_0$, in which there is no LIS. The model is then integrated from 8 to 0 kyr BP with varying solar forcing as prescribed in Berger [1978], the reconstructed atmospheric CO$_2$ concentration from Taylor Dome (Fig. 1), and for different LIS cases and a suppressed “fertilization” case: (1) with no LIS, but with interactive vegetation (AOV$_0$ run in Table 1); (2) with LIS fixed at 8 kyr BP, but with interactive vegetation (AOV$_8$ run in Table 1); (3) with interactive vegetation and the prescribed retreat of the LIS as in WMWB (AOVI run in Table 1); and (4) same as in the AOV1 run, except that the CO$_2$ fertilization is suppressed for the terrestrial carbon cycle module by fixing the CO$_2$ content at 260 ppmv (AOVI$_*$ run in Table 1). We carry out the AOV$_0$ run to compare our model results with those of Brovkin et al. [2002] and Joos et al. [2004]. Note that: (1) the ice sheet component is not interactive in any of these simulations; (2) the Greenland Ice Sheet (GIS) is prescribed at the most eastern part of northern North America, and it is fixed in all these simulations; and (3) the vegetation component is interactive in all these simulations. We fixed the GIS because the changes are much smaller during the pre-industrial Holocene than during the glacial period [Huybrechts, 2002].

2.3. An inverse method

Since we do not yet have an oceanic carbon cycle module in the green MPM, we used an inverse method to close the Holocene global carbon cycle in our model. On the timescales of interest, there are three major reservoirs in the global carbon cycle: (1) terrestrial biosphere (including recalcitrant carbon), (2) atmosphere and (3) ocean (including sedimentation). Two-way exchanges of carbon can occur between the atmosphere and terrestrial biosphere, and between the atmosphere and ocean; however, the only exchange of carbon between the terrestrial biosphere and ocean is the one-way transport from land to ocean through river discharge of DIC and dissolved organic carbon (DOC). Examples of how we use the inverse method are as follows: If both terrestrial and atmospheric carbon storages increase over a certain period, we can conclude that the source for these carbon increases
must come from the ocean. On the other hand, if the atmospheric carbon increases and the terrestrial carbon decreases, we cannot conclude that the source of the increased carbon in the atmosphere comes from the terrestrial biosphere only. Some of the atmospheric increase could also come from the ocean.

As in Brovkin et al. [2002] and Joos et al. [2004], our goal is to have the global carbon storage conserved in our transient simulations. To ensure this conservation in their models, Brovkin et al. [2002] and Joos et al. [2004] added an external sedimentation process in the oceanic carbon cycle module. In our simulations, the oceanic carbon pool (including the sedimentation process) is considered as a box whose content is varied to compensate for the changes of (1) the atmospheric carbon storage calculated from Taylor Dome, and (2) the terrestrial carbon storage simulated by VECODE. From Fig. 1 and our simulations, we can calculate the carbon storage for the atmosphere and land each year. Assuming conservation of total carbon in the system, we can infer the required carbon changes in the ocean.

As shown in Fig. 1 of Joos et al. [2004], the ocean is roughly in equilibrium with the atmosphere on a millennial timescale, and about 85% of the carbon emission from the land into the atmosphere must have been removed by the ocean. If we can estimate the total amount of carbon release from the terrestrial biosphere, we can calculate the contribution of the land carbon emission to the atmospheric \( \text{CO}_2 \) increase by the following equation:

\[
A_{\text{CO}_2} = C_{\text{source}} P_f R_a, \tag{4}
\]

where \( A_{\text{CO}_2} \) is the atmospheric \( \text{CO}_2 \) concentration increase in ppmv, \( C_{\text{source}} \) is the total land carbon source in PgC, \( P_f = 0.15 \) is the ratio of the total land carbon source that has remained in the atmosphere, and \( R_a = 0.47 \text{ppmv}/\text{PgC} \) is the well-known conversion constant for the atmosphere.

The value of \( P_f \) used on the right-hand side of Eq. 4 is an approximation for the Holocene case (pers. comm., F. Joos, 2004). At a background atmospheric \( \text{CO}_2 \) level of about 280 ppmv, only about 15% of the perturbation in the atmosphere-ocean system remains airborne. We note that this airborne fraction is not well constrained, as ocean carbonate chemistry and carbonate compensation will further reduce it.

3. Carbon storage underneath an ice sheet

Before we discuss the simulated terrestrial carbon storage for the equilibrium runs (EQ-0k, EQ-6k, and EQ-8k), we need to have a method for estimating the carbon storage underneath an ice sheet. Van Andel and Tzedakis [1996] suggested that before the Last Glacial Maximum (LGM), there were many areas of vegetation and soil that were later covered by the LIS. Recently, Zeng [2003] carried this one step further by proposing "The Glacial Burial Hypothesis" to explain the glacial-interglacial atmospheric \( \text{CO}_2 \) changes.

However, Adams and Faure [1998] argued that it is doubtful whether there was a major store of organic carbon underneath the continental ice sheets (or whether the carbon was frozen into permafrost in the periglacial zones) at the LGM. Most sedimentological and pedological studies (e.g., see West [1978] and Williams et al., [1993]) of both ancient and recently-produced glacial de-
bris indicate that this material is highly sterile and lacking in organic matter and weathered nutrients. Assuming the pre-industrial condition as a good proxy for the pre-LGM state, we propose four possible scenarios for carbon storage underneath an ice sheet. In scenario I, we neglect all the terrestrial carbon inventories, namely, B1, B2, B3 and B4 as defined in Sect. 2.1. In scenario II, we keep only the slow SOM carbon pool (B4), and neglect all other terrestrial carbon inventories. In scenario III, we neglect the green and structural biomass pools (B1 and B2) and keep the slow and fast SOM pools (B3 and B4). In scenario IV, we store half of the total carbon above ground (B1 and B2) in the fast SOM pool (B3), and half in the slow SOM (B4).

From Table 2, we note that for these four scenarios, the changes in a specific experiment (e.g., EQ-0k) are quite small, which means that the green MPM simulates negligible terrestrial carbon storage when the land surface is covered by the prescribed GIS in all four scenarios. In fact, the total terrestrial carbon storage neglected (scenario I) in the EQ-8k run is about 6 PgC, which is only 0.28% of the simulated global terrestrial carbon storage at 8 kyr BP. It is also interesting to note that in two sensitivity experiments (EQ-0k-CO\textsubscript{2} and EQ-8k-clm in Table 2), the results are quite different from the pre-industrial equilibrium (EQ-0k) run (see detailed discussion in Sect. 4.3). Since the neglected terrestrial carbon storage (scenario I) is at a maximum 0.41% (9 PgC) of the simulated global terrestrial carbon for the EQ-8k-clm run, we will choose scenario I. That means we neglect all the terrestrial carbon inventories (B1, B2, B3 and B4) in all our simulations afterwards, whenever the land surface is covered by an ice sheet (the LIS and/or the GIS).

4. Equilibrium simulations

The pre-industrial equilibrium simulation (EQ-0k) is important for evaluating the terrestrial carbon cycle module in the green MPM. Our model simulates a global NPP of 58.7 PgC/yr (Fig. 2C), which is consistent with estimates in IPCC TAR [Houghton et al., 2001] and falls within the ranges of other model simulations [Cramer et al., 1999, 2001]. The simulated global terrestrial carbon is 2095 PgC; this value derives from 598 PgC (green+structural) carbon plus 1497 PgC (fast and slow SOM) carbon. The above partitioning of the terrestrial carbon reservoir in the green MPM is consistent with the studies of Schimel et al. [1994], Schimel [1995], Foley [1995], and Cramer et al. [2001]. We added the recalcitrant carbon pool as estimated by Falloon et al. [1998] into the slow SOM pool in VECODE; thus, the turnover time for this pool has been increased from 1,000 years as in Brovkin et al. [2002], to 4,000 years. The addition of the recalcitrant carbon pool is important in our simulation, since its turnover time is of the same order as the length of our integration.

The general agreement between our simulated terrestrial carbon cycle and the published estimates for the pre-industrial carbon cycle gives us confidence in our model results. At steady state for the pre-industrial equilibrium, the simulated global NPP is partitioned into a green biomass flux of 18.8 PgC/yr and a structural biomass flux of 39.9 PgC/yr. Note that the green biomass flux includes only green leaf production in VECODE. The growth of roots and shoots is included in the structural biomass flux. The slowest decomposition rate of SOM is about 1 PgC/yr, which is much lower than the rate of litterfall from the green biomass (9 PgC/yr) and from
the structural biomass (33.7 PgC/yr) (Fig. 2C).

The simulated pre-industrial vegetation distribution is shown in Fig. 3. We note that northern Canada and Eurasia are covered mainly by boreal forest (Fig. 3A), except for the most eastern part of northern North America, where the Greenland Ice Sheet is prescribed. In the subtropical region, the green MPM simulates the location of the African and Middle-East deserts (Fig. 3C). In the equatorial region, the tropical rain forest is well located. With inclusion of the vegetation-precipitation feedback, the rain forest fraction has increased from 0.4-0.5, as seen in Fig. 2 of Wang et al. [2005a], to 0.6-0.7 (Fig. 3A). However, the green MPM still overestimates the Southern Hemisphere (SH) forest cover because it simulates too much precipitation in narrow-width continents (see the moisture transport parameterization scheme in Wang and Mysak, 2000). Generally, the pre-industrial vegetation distribution is in good agreement with the observation of Olson et al. [1983] and model simulations of Cramer et al. [2001].

4.1. Simulated terrestrial carbon storage changes

At 8 kyr BP, the simulated global terrestrial carbon is 2137 PgC (Fig. 2A), which is the sum of 599 PgC (green+structural biomass) and 1538 PgC (fast+slow SOM). This is within the range of the estimates of Peng et al. [1994, 1998], Foley [1994] and Adams and Faure [1998]. The model also simulates a green Sahara region (see WMWB), which is similar to the reconstruction of TEMPO Members [1996] and the study of Prentice and Jolly [2000]. From 8 to 0 kyr BP, the total SOM carbon decreases by 41 PgC (Figs. 2A and C), accounting for 98% of the global terrestrial carbon decrease over this period. This results from a small net change in vegetation (biomass) carbon. However, there are regional changes in vegetation carbon, which largely cancel out. For the same period, Brovkin et al. [2002] calculated a 90 PgC decrease for the terrestrial carbon store, while Joos et al. [2004] found increases ranging from 39 to 158 PgC depending on the climate models used. Interesting enough, like in Joos et al. [2004], Beerling [2000] also simulated mid-Holocene carbon storages that were 265 and 370 PgC lower than the pre-industrial value, when he coupled the Sheffield Dynamic Global Vegetation Model (SDGVM) with climate forcings from the UK Universities Global Atmospheric Modelling Programme (UGAMP, Hall and Valdes, 1997) and the NCAR GENESIS simulations [Kutzbach et al., 1998], respectively.

On the other hand, François et al. [1999] reconstructed the global terrestrial carbon storage using the carbon assimilation in the biosphere (CARAIB) model for the mid-Holocene and LGM. They estimated that the carbon storage changes from the mid-Holocene to the pre-industrial period ranged from a decrease of 132 PgC to an increase of 92 PgC. The CARAIB model uncertainties with respect to the CO2 fertilization contribute to the range of storage changes.

The NPP at 8 kyr is 2.1 PgC/yr higher than that at 0 kyr BP (Table 3), which is similar to the value of 3.1 PgC/yr in Brovkin et al. [2002]. The largest change of total carbon occurs in the Sahara region where there is a decrease of 70 PgC from 8 to 0 kyr BP (Table 3). This is partially compensated by the increased carbon in the northern boreal forest, the SH, and the equatorial areas (the total increase is 48 PgC from 8 to 0 kyr BP, Table 3). The main contributions to the SH carbon rise are increases in the “slow” pool of structural biomass (B2) and the “slow” compartment of SOM (B4) (figure not shown).
The increases in the B2 and B4 pools are associated with an increased tree fraction (about 0.1, figure not shown), caused by the relatively wetter condition (see Fig. 7 in WMWB) and the slightly warmer austral summer (see Fig. 6 in WMWB).

From Fig. 2B, we note that, at 6 kyr BP, the global terrestrial carbon is 2163 PgC. This value is the sum of 610 PgC (green+structural biomass) and 1553 PgC (fast+slow SOM). From 8 to 6 kyr BP the total carbon in northern boreal forest increases by 11 PgC (Table 3). This is 42% of the global carbon increase (26 PgC). Such a large contribution from this region can be attributed to the prescribed retreat of the LIS and the positive vegetation-albedo feedback. The SH total carbon increases by 25 PgC (Table 3) over the same period due to an increased tree fraction (about 0.05, figure not shown). The gradual desertification of the Sahara region results in a reduction of the total carbon by 25 PgC from 8 to 6 kyr BP (Table 3). The desert fraction in the Sahara region increases about 10% at 6 kyr BP (see Fig. 12 in WMWB). Since the above changes of total carbon in the SH and Sahara region offset each other, the remaining 58% of the global carbon increase from 8 to 6 kyr BP must come from the NH mid-latitude and equatorial regions.

Based on our inverse method (Sect. 2.3), the inferred carbon changes in the ocean for the periods from 8 to 6 kyr BP and from 6 to 0 kyr BP are -46 and 46 PgC, respectively (Table 4). The ocean is the only source for both the atmospheric and land carbon increases from 8 to 6 kyr BP. However, from 6 to 0 kyr BP, terrestrial carbon emission could contribute to the atmospheric CO$_2$ rise. The terrestrial carbon emission could produce an atmospheric CO$_2$ rise of about 5 ppmv from 6 to 0 kyr BP, according to (4).

4.2. Simulated terrestrial carbon dynamics

The simulated biomass changes slightly from 8 to 6 kyr BP (Fig. 4A) and from 6 to 0 kyr BP (Fig. 4B). From 8 to 6 kyr BP, the largest increase occurs in eastern North America, where the simulated biomass increase ranges from 1.5 to 5.5 kgCm$^{-2}$. The retreating LIS, together with the vegetation-albedo feedback, contributes to this large biomass increase. In contrast, the simulated biomass increase from 8 to 6 kyr BP in the SH is smaller (up to 1.5 kgCm$^{-2}$). On the other hand, the simulated biomass decreases from 8 to 6 kyr BP in mainly monsoon related regions (e.g., southeastern Asia and central America). As shown in Fig. 7 of WMWB, the monsoon precipitation decreases from 8 to 6 kyr BP in these regions. From 6 to 0 kyr BP, the simulated biomass (Fig. 4B) substantially decreases in the NH, in particular, in northern Africa, southeastern Asia and central America. This decrease is likely due to: (1) the decreasing summer insolation in the NH, caused by the orbital parameter changes; and (2) the desertification of the Sahara region, caused by the further reduction of monsoon precipitation there (see Fig. 12 of WMWB).

The simulated soil carbon increases substantially from 8 to 6 kyr BP in eastern North American high-latitudes (Fig. 5A), ranging from 5 to 15 kgCm$^{-2}$. The largest increase is mainly located in areas which have been covered by the prescribed LIS at 8 kyr BP. This is because we neglect all the soil carbon if the land surface is covered by an ice sheet (see Sect. 3). For the rest of the globe, the simulated soil carbon slightly increases from 8 to 6 kyr BP, except in subtropical NH (e.g., northern Africa and southeastern Asia) where it decreases due to
the simulated decrease of the NPP (figure not shown) and biomass (Fig. 4A) in the same region. From 6 to 0 kyr BP, the simulated soil carbon slightly decreases in North American high-latitudes, and dramatically decreases in northern Africa and southeastern Asia, with a maximum decrease of 3-5 kg C m$^{-2}$. The simulated conversion from vegetated to desert conditions in the Sahara region (see Fig. 12 in WMWB) could explain such a big decrease of soil carbon. In contrast, the simulated soil carbon generally increases from 8 to 0 kyr BP in the SH (Fig. 5A and B), which is related to an increased forest cover (see Fig. 10A).

### 4.3. Sensitivity of Holocene terrestrial carbon dynamics

We carry out two additional sensitivity experiments to determine the relative importance of atmospheric CO$_2$ and climate on the terrestrial carbon cycle. Our approach follows the method of Brovkin et al. [2002]. In the first experiment (EQ-0k-CO$_2$), the terrestrial carbon cycle is driven by the atmospheric CO$_2$ concentration at 8 kyr BP (260 ppmv), but the physical components of the model simulate the pre-industrial climate as in the EQ-0k run, with the CO$_2$ forcing fixed at 280 ppmv (Table 1). In the second experiment (EQ-8k-clm), the terrestrial carbon cycle responds to the pre-industrial atmospheric CO$_2$ concentration (280 ppmv), but the physical components of the model simulate the climate at 8 kyr BP as in the EQ-8k run, with the CO$_2$ forcing fixed at 260 ppmv (Table 1). In the EQ-0k-CO$_2$ run, lower atmospheric CO$_2$ leads to a decrease in the NPP of 0.9 Pg C/yr; consequently, the carbon stored in the biomass and SOM decrease by 10 Pg C and 24 Pg C, respectively (Table 5). In contrast, the EQ-8k-clm run reveals an increase of 3.1 Pg C/yr in the NPP and an increase of 71 Pg C in global terrestrial carbon, as compared to a decrease of 34 Pg C in the EQ-0k-CO$_2$ experiment (Table 5). We thus confirm the conclusion of Brovkin et al. [2002], that changes in climate are more important for the terrestrial carbon cycle than changes in atmospheric CO$_2$ concentration.

In summary, we simulate global terrestrial carbon storages at 8 kyr BP and 6 kyr BP, that are 42 Pg C and 68 Pg C above the pre-industrial value, respectively. We also note that in our simulation the terrestrial carbon increases from 8 to 6 kyr BP, and then decreases from 6 to 0 kyr BP. This nonlinear behavior is different from the results in the studies of Brovkin et al. [2002] (who obtained a monotonic decrease of terrestrial carbon from 8 to 0 kyr BP) and Joos et al. [2004] (who obtained a gradual increase of terrestrial carbon from 8 to 0 kyr BP). Table 6 gives a list of model estimates (including our own) and reconstructed changes of global terrestrial carbon in the Holocene.

### 5. Transient simulations

We now discuss the results for the transient simulations $AOV_0$, $AOV_8$, $AOV_I$, and $AOV_{I*}$ (Table 1). Our $AOV_0$ simulation corresponds to the runs of Brovkin et al. [2002] and Joos et al. [2004], since in their simulations there are no residues of the LIS from 8 to 6 kyr BP and from 7 to 6 kyr BP, respectively. The $AOV_8$ simulation was carried out to examine the impact of a fixed LIS on the Holocene climate and the terrestrial carbon cycle. The $AOV_I$ simulation investigates the Holocene climate and the terrestrial carbon cycle in the presence of a prescribed retreating LIS. Finally, the $AOV_{I*}$ simulation suppresses the CO$_2$ fertilization in $AOV_I$. 
5.1. Simulated Holocene climate changes

In the AOV I simulation, the global annual mean SST increases by about 0.2°C between 8 and 1 kyr BP (Fig. 6A). The first 0.1°C increase from 8 to 6 kyr BP occurs because of the warming caused by the retreating LIS, which overrides the orbital-induced cooling for the same period. However, we believe that the second 0.1°C increase is due to the long-term adjustment of the ocean circulation and global SST forced by the 15 ppmv increase in atmospheric CO\textsubscript{2} from 6 to 1.2 kyr BP. Without the LIS (AOV\textsubscript{0} run), the global SST is initially higher (Fig. 6A). After 5.8 kyr BP, the two SST curves from the AOV I and AOV\textsubscript{0} runs are similar. The global SST in the AOV\textsubscript{8} run is always smaller than the SST in the AOV I and AOV\textsubscript{0} runs. Presumably, the fixed LIS at 8 kyr BP induces additional overall cooling on the generally increasing trend of the global SST.

The global annual mean SAT in the AOV I run increases during the prescribed retreat of the LIS, and it reaches a local peak value of 16.3°C at around 5.5 kyr BP (Fig. 6B). It then oscillates around 16.2°C between 5.5 and 1.5 kyr BP, and reaches 16.35°C at 1.2 kyr BP. For the AOV\textsubscript{0} run, the global SAT is higher at the beginning than the AOV I run, due to the absence of the LIS. However, after 6.5 kyr BP, the global SAT is slightly lower than the AOV I run, presumably due to the expansion of northern boreal forest and the vegetation-albedo feedback. After 5 kyr BP, the two global SAT curves for the AOV I and AOV\textsubscript{0} runs are similar. The global SST and SAT curves in the AOV\textsubscript{8} run are lower than those in the AOV I and AOV\textsubscript{0} runs because of the presence of the LIS, which has been fixed at 8 kyr BP (Figs. 6A and B).

The NH and SH SAT changes in the three transient simulations (Fig. 6C and D) are quite different from those shown in Fig. 6B. The NH SAT in the AOV I run reaches a maximum at around 5.5 kyr BP (Fig. 6C) due to the prescribed retreat of the LIS and the associated northward shift in the treeline. The decreasing NH summer insolation produces an NH SAT decrease from 5.5 to 0 kyr BP. In contrast, the SH SAT increases from 8 to 1.2 kyr BP in the AOV I run because of the increased SH summer insolation and forest cover (see Fig. 10A). We also simulate an increased SH terrestrial carbon during the same period (see Fig. 9). The SH SAT increase is most likely due to the vegetation-albedo feedback [Wang et al., 2005a] and the vegetation-precipitation feedback (Sect. 2.1). Without the LIS (AOV\textsubscript{0} run), the patterns of NH and SH SAT are similar to those in the AOV I run. Apart for a period right before the total disappearance of the LIS at 6 kyr BP in the AOV I run, the NH and SH SAT in the AOV\textsubscript{0} run is higher than in the AOV I run (Fig. 6C and D). For the AOV\textsubscript{8} run, the overall cooling due to the fixed LIS at 8 kyr BP is most pronounced in the NH SAT (Fig. 6C). It is interesting to note that the green curves (AOV I) do not simply stay between the blue (AOV\textsubscript{8}) and red (AOV\textsubscript{0}) curves in Fig. 6, because of the positive vegetation-climate feedbacks under the prescribed retreat of the LIS.

We note that there was a significant drop in temperature from 1.2 to 0.5 kyr BP. We argue that this big drop comes from the radiative forcing (cooling of about 0.2 W/m\textsuperscript{2}) of the atmospheric CO\textsubscript{2}, because with fixed atmospheric CO\textsubscript{2}, there is no such drop in temperature (see Figure 13 in WMWB). We also note that the climate fields from the AOV I and AOV\textsubscript{8} simulations are similar (figure not shown), as only the CO\textsubscript{2} fertilization effect is suppressed in AOV\textsubscript{8}, and the simulated vegetation distribution changes slightly.
5.2. Simulated Holocene terrestrial carbon cycle

The global terrestrial carbon curves (Fig. 7B) follow the trends of the global NPP curves (Fig. 7A) for the four transient runs. However, during the first 2,000 years, the lowest global NPP value is obtained in the AOV₀ run, whereas the lowest global terrestrial carbon value is simulated in the AOV₈ run. In the AOV_I run, the global terrestrial carbon first increases by about 27 PgC from 8 to 6 kyr BP, and then decreases by about 74 PgC during the next 6 kyr. The increase during the first 2 kyr is partially associated with the prescribed retreat of the LIS and the subsequent northward shift of the boreal forest treeline. The decrease after 6 kyr BP can be attributed to the decreasing summer insolation in the NH. Without the prescribed retreat of the LIS (AOV₀ and AOV₈ runs), the global terrestrial carbon decreases on average monotonically from 8 to 0 kyr BP. Upon comparing the AOV_I and AOV_I* runs, we note that, with CO₂ fertilization suppressed, the global NPP and terrestrial carbon are lowered by 0.8 PgC/yr and 25 PgC at 0 kyr BP, respectively. Upon comparing the equilibrium runs with the AOV_I run, we also note that there are slight differences. These differences are presumably caused by the unequilibrium state in the AOV_I run.

The total variation in the SOM carbon (Fig. 7D) is much larger than that in the biomass carbon (Fig. 7C, note the different vertical scales) for the four transient runs. In fact, the peak to peak change in the biomass carbon for the AOV_I run is about 20 PgC. However, the same change in the SOM carbon for the same run is about 60 PgC, which shows that the response of the SOM carbon dominates that of the biomass carbon during the Holocene. This is because the response time of the biomass is faster than that of the SOM in VECODE. Without the retreating LIS (AOV₀ and AOV₈ runs), the global biomass carbon is stabilized between 585 PgC and 605 PgC. On the other hand, the SOM carbon for these runs decreases monotonically throughout the Holocene. The first increase from 8 to 6 kyr BP and the later decrease from 6 to 0 kyr BP patterns are only obtained when the retreating LIS is added in the AOV_I simulation. In the case of suppressed CO₂ fertilization (AOV_I*), its effect is almost equivalent to a fixed LIS in the AOV₈ run from 4 to 0 kyr BP. As noted above, the green curves (AOV_I) often overshoot or even stay above the red curves (AOV₀). Hence, we suggest that it is important to have a transient boundary condition in order to simulate a transient response of global carbon cycle.

We next analyze the total carbon (living biomass and SOM) stored in the different regions of the model (Fig. 8). Firstly, the total carbon changes in northern boreal forest (60° N to 75° N) for the AOV_I run (Fig. 8A) are consistent with the treeline shifts in this region [TEMPO Members, 1996; WMWB]. We also note that in the green MPM, the total carbon in the boreal forest (50° N to 75° N) is about 450 PgC at 0 kyr BP (figure not shown), which is consistent with observations (e.g., IPCC TAR, Houghton et al., 2001). During the period from 8 to 6 kyr BP, the total carbon increases as the boreal forest expands northward, and then it decreases from 6 to 0 kyr BP when the boreal forest retreats southward. The suppressed CO₂ fertilization (AOV_I* run) reduces the range of the variation but not its pattern. Secondly, the total carbon in the Sahara region decreases by about 70 PgC from 8 to 1 kyr BP in all runs (Fig. 8B). This is directly related to the desertification over this period, as simulated in WMWB. The suppressed CO₂ fertilization has negligible effect in this region. Thirdly, in Fig. 8C,
the total carbon increases about 10 PgC in the equatorial area for the three transient runs $AOV_1$, $AOV_2$ and $AOV_3$ (hereafter three runs). The total carbon stays almost constant for the suppressed CO$_2$ fertilization ($AOV_t$) in this area. Finally, due to the increasing summer insolation in the SH during the Holocene, there the total carbon in the three runs increases by about 40 PgC from 8 to 2 kyr BP (Fig. 8D). Suppressing the CO$_2$ fertilization ($AOV_1$) reduces the range of the increase by 7 to 8 PgC, as expected. Because of the small differences in the three runs for the Sahara, equatorial, and SH regions, it is clear that the receding LIS has little impact on the dynamics of the Holocene carbon cycle there. Upon comparing Figs. 8B and D, we note that from 8 to 6 kyr BP, the total carbon decrease in the Sahara region is balanced by the total carbon increase in the SH for the three runs. This was illustrated in our earlier equilibrium simulations (Table 3). The CO$_2$ fertilization has the most influence in the equatorial area, since the total carbon density there decreases by 0.36 kgCm$^{-2}$ (figure not shown).

To determine why the SH total carbon increases during the Holocene, we now consider the SH biomass and SOM developments (Fig. 9) and the variations in the SH forest, grass, and desert areas (Fig. 10). Firstly, the trends for both the total biomass and SOM curves are similar for the three runs. When the CO$_2$ fertilization is suppressed ($AOV_t$), we find reduced carbon increases of total biomass and SOM of 3 and 7 PgC at 0 kyr BP, respectively. Secondly, both the grassland and desert areas in the SH decrease over the period 8 to 2 kyr BP (Fig. 10B and C). During this period, the SH forest area increases (Fig. 10A), which accounts for the increased biomass and SOM carbon in the SH (Fig. 9A and B).

5.3. Comparison with other model studies

We now compare our main findings for the Holocene climate and terrestrial carbon cycle dynamics with the results from the simulations in Brovkin et al. [2002] and Joos et al. [2004].

Joos et al. [2004] showed that during the Holocene, the global SST increased about 0.6°C in their model. This increase occurred mainly between 7 and 6 kyr BP. They concluded that there was an additional oceanic outgassing of CO$_2$ (about 6 ppmv) during this period. On the other hand, Brovkin et al. [2002] simulated a much smaller increase in the global SST (about 0.05°C) between 8 and 6 kyr BP in their AOVC-T simulation. Our global SST increases about 0.2°C between 8 and 1 kyr BP (Fig. 6A), which is intermediate between the above two increases.

Fig. 10 of Brovkin et al. [2002] showed that their global SAT was nearly steady between 8 and 6 kyr BP. It then declined by 0.1°C during the next 2,000 years and remained steady after 4 kyr BP. Our global SAT (Fig. 6B), however, first increases by 0.3°C to a local peak at about 5.5 kyr BP. This SAT then oscillates around 16.2°C between 5.5 and 1.5 kyr BP, and finally reaches a maximum of 16.35°C at 1.2 kyr BP. In contrast to these patterns of change, Fig. 2 of Joos et al. [2004] showed that from 8 to 0 kyr BP their global SAT increased by 0.8°C in the UM climate model. However, both their SAT and SST changes might be overestimated due to the slab ocean used in the UM climate model.

Upon comparing our model results for the terrestrial carbon cycle dynamics with the above studies, we find three major differences: (1) For the $AOV_1$ run, the global terrestrial carbon increases from 8 to 6 kyr BP, and 42%
of this increase occurs in northern boreal forest. The global terrestrial carbon then decreases from 6 to 0 kyr BP. The rise and fall pattern can be explained by the retreating LIS and the vegetation-albedo feedback, accompanied by the decreasing summer insolation in the NH. This contrasts with Brovkin et al. [2002], who did not include the retreating LIS, and hence found that the global terrestrial carbon decreases steadily throughout the Holocene by about 90 PgC. Joos et al. [2004], on the other hand, obtained a gradual increase of the global terrestrial carbon of about 39 or 158 PgC throughout the Holocene. We argue that this latter difference from our results is due to (i) the lack of the biogeophysical feedbacks, and (ii) the different climate fields used in their simulations. (2) Joos et al. [2004] did not simulate a greening of the Sahara region as reconstructed for 6 kyr BP. The greening of the Sahara region was obtained in our simulation and that of Brovkin et al. [2002]. (3) We found that the SH terrestrial carbon increases by 40 PgC from 8 to 2 kyr BP. The $CO_2$ fertilization contributes less than a 10-PgC increase during this period. In contrast, Brovkin et al. [2002] simulated a 20-PgC increase in the SH from 8 to 3 kyr BP. They suggest that the $CO_2$ fertilization can cause their simulated SH carbon increase (V. Brovkin, pers. comm., 2004). On the other hand, Joos et al. [2004, Fig. 4] simulated a SH carbon increase of about 20 PgC from 6 to 0 kyr BP. We suggest that an increased forest area throughout the Holocene in the SH is the major factor that explains the SH terrestrial carbon increase. The $CO_2$ fertilization has a secondary effect in this increase.

5.4. An explanation of the 20 ppmv rise in atmospheric $CO_2$ during the Holocene

We have shown above that both the global terrestrial carbon and the atmospheric $CO_2$ increased from 8 to 6 kyr BP. Based on our inverse method and the conservation of global carbon, the source of these carbon increases must have come from the ocean. Thus, the total carbon content of the ocean must decrease to supply the simulated increases in both the atmosphere and the terrestrial biosphere (Fig. 11). Hence, we conclude that the first 10-ppmv rise in the atmospheric $CO_2$ during this period comes from the ocean only. However, because of the absence of an oceanic carbon cycle module in the green MPM, we cannot resolve nor confirm the detailed processes involved in this 10-ppmv rise. On the other hand, we note from Fig. 11 that from 6 to 0 kyr BP, the global terrestrial carbon decreased by 68 PgC, while the atmospheric $CO_2$ continued to increase. The 68-PgC release of carbon from the terrestrial biosphere can only explain about a 5-ppmv ($68 \times 0.15/2.13$) rise in the atmospheric $CO_2$ from 6 to 0 kyr BP. Note that for the suppressed $CO_2$ fertilization case ($AOV I$), the global terrestrial carbon decreased by 95 PgC (Fig. 7B), which can cause about a 7-ppmv ($95 \times 0.15/2.13$) atmospheric $CO_2$ rise. Thus, the source of the remaining 3 to 5 ppmv increase in the atmospheric $CO_2$ from 6 to 0 kyr BP must come from the ocean, as well. The model uncertainty surrounding the $CO_2$ fertilization contributes to the range in the global terrestrial carbon decrease (5 to 7 ppmv rise in atmospheric $CO_2$) from 6 to 0 kyr BP. We also note that the uncertainty associated with the airborne fraction ($P_f$ in Eq. 4) might change our results with respect to the period from 6 to 0 kyr BP.
6. Summary and concluding remarks

The terrestrial carbon cycle dynamics and pre-industrial Holocene climate were investigated within the framework of an EMIC. An inverse method is introduced to explain the atmospheric CO$_2$ rise during the pre-industrial Holocene up to 8 kyr BP. Our model setup is different from Brovkin et al. [2002] in that we include a prescribed retreating LIS from 8 to 6 kyr BP [Dyke and Presot, 1986; 1987]. Our model setup is also different from Joos et al. [2004] in that the green MPM’s components are interactively coupled and thus take into account the vegetation-albedo and vegetation-precipitation feedbacks. We first evaluate our terrestrial carbon cycle for the pre-industrial equilibrium by adding the recalcitrant carbon into the slow SOM pool. In addition, in all the experiments, we neglect the one-way transport of carbon from land to ocean (cf. Brovkin et al. [2002] and Joos et al. [2004]), and the freshwater forcing from ice sheet melt. We further note that the transient behavior of the Holocene climate and global carbon cycle dynamics depend critically on the transient boundary conditions applied (Fig. 7).

The fully coupled AOV transient run simulated a decrease of 70 PgC in total carbon in the Sahara region, which is caused by desertification in this region [WMWB]. This decrease is partially compensated by an increase of 40 PgC in the total carbon in the SH from 8 to 2 kyr BP. The increasing austral summer insolation, which causes the expansion of forest area in the SH, is the major reason for the above carbon increase. The CO$_2$ fertilization contributes to less than 10 PgC in this carbon increase. In NH high latitudes, we note that the simulated treeline shifts [WMWB] are consistent with the total carbon storage changes in northern boreal forest: (1) northward shift (increased carbon storage) from 8 to 6 kyr BP; and (2) southward shift (decreased carbon storage) from 6 to 0 kyr BP. The simulated global SST increases about 0.1°C from 8 to 6 kyr BP, and 0.1°C from 6 to 0 kyr BP.

There are a few processes which are not represented in our model, and these might also contribute to the Holocene CO$_2$ rise. (1) Anthropogenic land use changes are not included in our simulations; hence we cannot evaluate the “grand” hypothesis of Ruddiman [2003]. (2) Peatland/wetland development is neglected; this may play an important role throughout the Holocene as suggested by Gajewski et al. [2001]. In fact, the estimated peatland formation and carbon store suggest that peat is a major portion of the terrestrial carbon budget (as much as 500 PgC) and that major peat development occurred during the Holocene [Gorham, 1991]. However, the timing and magnitude of the peatland/wetland development need to be closely constrained before it can be included in terrestrial and global carbon studies of the Holocene. (3) Coral reef formation is not considered. However, without a 3-D sedimentation module, we cannot test the hypothesis of Ridgwell et al. [2003].

It should also be noted that we have not attempted to constrain our results with the δ$_{13}$C data obtained from the Taylor Dome ice core. We do this because: (1) Broecker et al. [2001] argued that the small signal (0.2‰) to measurement error (0.06‰) ratio and the scatter of the results (0.08‰) around the model curve offer no assurance to avoid the contamination introduced by the CO$_2$ absorbed onto surfaces of the apparatus. They also argued that the reconstructed δ$_{13}$C data from the ice core cannot exclude the possibility that there has been no measurement trend in atmospheric δ$_{13}$C. (2) Joos et
al. [2004] noted that the amplitude of their simulated δ\(^{13}\)C change (0.2 \(^{\circ}\)/oo) was small compared to the data range (0.4 \(^{\circ}\)/oo). Thus, they argued that the precision of δ\(^{13}\)C data needed to be ± 0.1 \(^{\circ}\)/oo or better for a reliable separation of terrestrial and ocean carbon processes.

In summary, based on our inverse method, we find that: (1) The first 10-ppmv CO\(_2\) rise from 8 to 6 kyr BP comes from the oceanic carbon sources, which include outgassing due to the SST increase, calcite compensation, and other oceanic processes. (2) The second 10-ppmv CO\(_2\) rise from 6 to 0 kyr BP comes from terrestrial carbon release (5-7 ppmv), and the oceanic carbon sources (3-5 ppmv), which include outgassing due to the SST increase, calcite compensation, and other oceanic processes. In order to gain insight into the details of the oceanic carbon processes during the pre-industrial Holocene, we next plan to develop an oceanic carbon cycle module and couple it to the green MPM, in the form used in this paper.

Acknowledgments.

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Yi Wang, Earth System Modelling Group, Department of Atmospheric and Oceanic Sciences, McGill University, Montreal, Quebec, Canada, H3A 2K6. (yi-wang@zephyr.meteo.mcgill.ca)
**Figure 1.** (A) Taylor Dome derived atmospheric $\text{CO}_2$ concentration and corresponding carbon storage in PgC ($10^{15}$ gC). To convert ppmv to PgC, multiply the former by 2.13 PgC/ppmv, the well-known conversion constant for the atmosphere.

**Table 1.** Summary of equilibrium and transient simulations carried out.

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Orbital forcing</th>
<th>$\text{CO}_2$ (ppmv) in climate model</th>
<th>$\text{CO}_2$ (ppmv) in VECODE</th>
<th>LIS$^1$</th>
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<tr>
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<td>AOV$_I$</td>
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<td>From Taylor Dome</td>
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<td>AOV$_I$*</td>
<td>Variable</td>
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<td>260</td>
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</table>

$^1$Laurentide Ice Sheet.
$^2$The $\text{CO}_2$ fertilization suppressed in the terrestrial carbon cycle module.

**Table 2.** Results of terrestrial carbon storage in pools B3 and B4 from the above four proposed scenarios (I, II, III and IV) as listed in the text.

<table>
<thead>
<tr>
<th>Acronym</th>
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<th>Slow SOM$^1$ (B4)</th>
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<td>I 512 II 512 III 512 IV 1026</td>
<td>I 1026 II 1026 III 1026 IV 1026</td>
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<td>EQ-8k-clm*</td>
<td>I 519 II 519 III 524 IV 1041</td>
<td>I 1043 II 1043 III 1043 IV 1044</td>
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</tbody>
</table>

$^1$Soil Organic Matter in PgC.
$^2$We do not have green and structural biomass whenever the land surface is covered by an ice sheet.

**Figure 2.** Schematic diagram of the terrestrial carbon cycle in the green MPM for the year of (A) 8 kyr BP, (B) 6 kyr BP and (C) 0 kyr BP. Units are PgC for carbon storages and PgC/yr for carbon fluxes. Turnover times are in years.
**Figure 3.** Vegetation cover in the green MPM of (A) tree fraction, (B) grass fraction and (C) desert fraction for the pre-industrial period.

**Table 3.** Results for the terrestrial carbon storage from five equilibrium simulations.

<table>
<thead>
<tr>
<th>Acronym</th>
<th>NPP</th>
<th>TTCS</th>
<th>TCBO</th>
<th>TCSA</th>
<th>TCSH</th>
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</table>

1Net Primary Productivity in PgC/yr.
2Total terrestrial carbon storage in PgC.
3Total carbon storage in northern boreal forest (60°N to 75°N) in PgC.
4Total carbon storage in Sahara region (15°N to 30°N and 15°W to 50°E) in PgC.
5Total carbon storage in Southern Hemisphere (south of 10°S) in PgC.
6Total carbon storage in equatorial area (10°S to 10°N) in PgC.

**Table 4.** Carbon storage changes for the atmosphere, land and the ocean (the latter change inferred assuming a global carbon balance). A positive (negative) value denotes an increase (decrease).

<table>
<thead>
<tr>
<th>Period</th>
<th>Atmospheric carbon</th>
<th>Land carbon</th>
<th>Ocean carbon</th>
</tr>
</thead>
<tbody>
<tr>
<td>8 to 6 kyr BP</td>
<td>20</td>
<td>26</td>
<td>-46</td>
</tr>
<tr>
<td>6 to 0 kyr BP</td>
<td>22</td>
<td>-68</td>
<td>46</td>
</tr>
</tbody>
</table>

1Carbon storage changes from late to early times in PgC, from Fig. 1.
2Carbon storage changes from late to early times in PgC, from the simulations.
3Inferred carbon storage changes using the inverse method from late to early times in PgC.

**Figure 4.** The simulated biomass carbon difference in the green MPM (A) between 6 and 8 kyr BP and (B) between 0 and 6 kyr BP. Units are $kgCm^{-2}$. Note that a positive value shows an increase of biomass from earlier to later times.

**Figure 5.** The simulated soil carbon difference in the green MPM (A) between 6 and 8 kyr BP and (B) between 0 and 6 kyr BP. Units are $kgCm^{-2}$. Note that a positive value shows an increase of soil carbon from earlier to later times; the color scale is different from Figs. 4.
Table 5. Changes in the terrestrial carbon storage, given by the difference between the sensitivity simulation and the EQ-0k simulation. Note that a positive number means an increase in the sensitivity simulation over the EQ-0k simulation.

<table>
<thead>
<tr>
<th>Acronym</th>
<th>NPP</th>
<th>Biomass</th>
<th>SOM</th>
<th>Total carbon</th>
</tr>
</thead>
<tbody>
<tr>
<td>EQ-0k-CO₂</td>
<td>-0.9</td>
<td>-10</td>
<td>-24</td>
<td>-34</td>
</tr>
<tr>
<td>EQ-8k-clm</td>
<td>3.1</td>
<td>9</td>
<td>63</td>
<td>71</td>
</tr>
<tr>
<td>EQ-Sk</td>
<td>2.2</td>
<td>2</td>
<td>41</td>
<td>42</td>
</tr>
</tbody>
</table>

1 Net Primary Productivity change in PgC/yr.
2 Total carbon storage change in green and structural biomass in PgC.
3 Total carbon storage change in soil organic matter in PgC.
4 Total carbon storage change in PgC.

Table 6. Holocene modelled or reconstructed terrestrial carbon changes.

<table>
<thead>
<tr>
<th>Times</th>
<th>Changes¹</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>8 to 0 kyr BP</td>
<td>-195</td>
<td>Indermühle et al. [1999]</td>
</tr>
<tr>
<td>8 to 0 kyr BP</td>
<td>-90</td>
<td>Brovkin et al. [2002]</td>
</tr>
<tr>
<td>8 to 6 kyr BP</td>
<td>11 or 83</td>
<td>Joos et al. [2004]</td>
</tr>
<tr>
<td>8 to 6 kyr BP</td>
<td>26</td>
<td>This paper</td>
</tr>
<tr>
<td>6 to 0 kyr BP</td>
<td>-40</td>
<td>Foley [1994]</td>
</tr>
<tr>
<td>6 to 0 kyr BP</td>
<td>-2 to 0</td>
<td>Peng et al. [1998]</td>
</tr>
<tr>
<td>6 to 0 kyr BP</td>
<td>-92 to 132</td>
<td>François et al. [1999]</td>
</tr>
<tr>
<td>6 to 0 kyr BP</td>
<td>28 or 75</td>
<td>Joos et al. [2004]</td>
</tr>
<tr>
<td>6 to 0 kyr BP</td>
<td>-68</td>
<td>This paper</td>
</tr>
</tbody>
</table>

¹ The difference in global terrestrial carbon between earlier and later times in PgC. Note that a positive number means an increase from earlier to later times.

Figure 6. Global annual mean SST (A), SAT (B), and the NH SAT (C), and the SH SAT (D) in the green MPM from three transient simulations.

Figure 7. Global NPP (A), terrestrial carbon (B), terrestrial biomass (C), and soil organic matter (D) in the green MPM from four transient simulations.

Figure 8. Total carbon storages for (A) northern Boreal forest (60°N to 75°N), (B) Sahara region (15°N to 30°N and 15°W to 50°E), (C) Equatorial area (10°S to 10°N), and (D) Southern Hemisphere (south of 10°S) in the green MPM from four transient simulations.
Figure 9. Southern Hemisphere (A) biomass carbon, and (B) SOM carbon in the green MPM from four transient simulations.

Figure 10. Southern Hemisphere (A) forest area, (B) grassland area, and (C) desert area in the green MPM from four transient simulations.

Figure 11. Carbon storage changes in the green MPM for the AOV1 simulation.
Taylor Dome ice core data

Atmospheric CO₂, ppmv

Derived atmospheric carbon storage, PgC

Time, years

-12000 -10000 -8000 -6000 -4000 -2000 0

Atmospheric CO₂, ppmv

255 260 265 270 275 280 285

Derived atmospheric carbon storage, PgC

550 560 570 580 590 600 610
Tree fraction in the green MPM

Grass fraction in the green MPM

Desert fraction in the green MPM
The simulated biomass carbon difference (in kgC/m²) in the green MPM

A. 6–8 kyr BP

B. 0–6 kyr BP
The simulated soil carbon difference (in kgC/m²) in the green MPM

A. 6–8 kyr BP

B. 0–6 kyr BP
Global annual mean SST from three transient runs

A.

Global annual mean SAT from three transient runs

B.

The NH annual mean SAT from three transient runs

C.

The SH annual mean SAT from three transient runs

D.
Global net primary production from four transient runs

A. 

Global total land carbon from four transient runs

B. 

Terrestrial biomass from four transient runs

C. 

Soil organic matter from four transient runs

D.
Carbon storage in northern Boreal forest (60°N to 75°N)

Carbon storage in Sahara region

Carbon storage in Equatorial area

Carbon storage in Southern Hemisphere
Southern Hemisphere terrestrial biomass from four transient runs

A. Carbon storage, PgC

Southern Hemisphere soil organic matter from four transient runs

B. Carbon storage, PgC

Time, years
Southern Hemisphere forest area from four transient runs

Southern Hemisphere grassland area from four transient runs

Southern Hemisphere desert area from four transient runs
Changes of carbon in land, atmosphere and ocean for the AOVI simulation

- **Land (simulated)**
- **Atmosphere (forcing)**
- **Ocean (inferred)**

**Carbon storage change in PgC**

**Time, years**

- **Ocean (inferred)**
- **Atmosphere (forcing)**
- **Land (simulated)**

**Legend**
- Red line: Ocean (inferred)
- Blue line: Atmosphere (forcing)
- Green line: Land (simulated)