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A dynamic model of wetland extent and peat accumulation: results for the Holocene

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Abstract

Substantial deposits of peat have accumulated since the last glacial. Since peat accumulation rates are rather low, this process was previously neglected in carbon cycle models. For assessments of the global carbon cycle on millennial or even longer timescales, though, the carbon storage in peat cannot be neglected any more. We have therefore developed a dynamic model of wetland extent and peat accumulation in order to assess the influence of peat accumulation on the global carbon cycle.

The model is based on the dynamic global vegetation model LPJ and consists of a wetland module and routines describing the accumulation and decay of peat. The wetland module, based on the TOPMODEL approach, determines wetland area and water table. Peatland area is given by the innundated area at the summer minimum water table position and changes dynamically, depending on climate. The peat module describes oxic and anoxic decomposition of organic matter in the acrotelm, i.e., the part of the peat column above the permanent water table, as well as anoxic decoposition in the catotelm, the peat below the summer minimum water table.

We apply the model to the period of the last 8000 yr, during which the model accumulates 210 ± 40 PgC as catotelm peat in the areas above 40° N.

1 Introduction

Estimates of the amount of carbon stored in boreal peatlands vary. Gajewski et al. (2001) estimate that peat deposits of about 450 PgC have accumulated since the last glacial, though other estimates are substantially lower, e.g. 273 PgC estimated by Turunen et al. (2002). Nonetheless it is clear that boreal peatlands store substantial amounts of carbon, which may be up to a fifth of the total global soil carbon estimated as 2344 PgC in the top three meters (Jobbagy and Jackson, 2000).

On interannual timescales, the changes in peat storage are rather small, and peat accumulation has therefore been neglected in carbon cycle models so far. On millennial timescales, this is a substantial factor in the carbon cycle, though, which is why we have developed a dynamical model of wetland extend and peat accumulation as described in this paper.

In order to represent the carbon cycle forcing by peatlands on millennial timescales, previous authors used scenarios of peat accumulation (Wang et al., 2009; Kleinen et al., 2010). While the use of such scenarios is possible for the Holocene, where some measurement data to derive the scenarios exists, such data do not exist for previous interglacials, since the glaciation occurred in just those places, where present-day peat deposits are located. The investigation of carbon cycle dynamics in previous interglacials therefore requires the use of a dynamical model.

Models of peat accumulation have previously mainly been developed for single sites. Clymo (1984) developed a one-dimensional model of peat accumulation and decay in a single peat column. Later Clymo et al. (1998) determined parameter values for this model by fitting to profiles from numerous peat bogs from Finland and Canada. Contrary to the Clymo (1984) model which focuses on the biochemical decomposition processes in the peat layers below the water table, Ingram (1982) has developed a model of peat from a hydrological perspective, describing the groundwater table in a domed peat deposit. Over time these models have included more and more processes, for example three-dimensional bog growth, i.e., the lateral expansion of peatlands from the site of first initiation (Korhola et al., 1996), more sophisticated parameterisations of water table depth, or more plant functional types (PFTs) (Frolking et al., 2010). This line of development has so far culminated in the Frolking et al. (2010) model describing the development of annual peat layers and thereby resolving the accumulation and decomposition processes in considerable sophistication.

The other approach is to model peat accumulation and decomposition in the framework of dynamic global vegetation models (DGVM) or the land surface components of climate models. Such a global approach necessarily neglects some of the detail included in dedicated site models, due to computational constraints, but also due to the fact that detailed parameterisations often are dependent on site specific parameters. Examples of models following this global approach are Wania et al. (2009a,b) who developed an extension of the LPJ DGVM, accounting for organic soils with the rationale to derive methane emissions from wetlands. Their model relies on prescribed wetland areas and does not determine the extent of wetlands dynamically. Ringeval et al. (2010) similarly modeled wetland CH_4 emissions using maps of wetland extent.

Parameterisations for the determination of wetland extent have been developed and implemented in a number of cases. Kaplan (2002) used a digital elevation model (DEM) to determine areas, where wetlands could develop. Using an approach based on TOPMODEL (Beven and Kirkby, 1979), wetland parameterisations have been developed for the NCAR GCM (Niu et al., 2005), the ISBA land surface model ORCHIDEE (Habets and Saulnier, 2001; Decharme and Douville, 2006), and the MetOffice land surface scheme MOSES (Gedney and Cox, 2003; Gedney et al., 2004). In all of these cases an explicit accounting for the long term accumulation and decay of peat is missing, since these studies focus on methane emissions, not peat accumulation.

The present study therefore aims to combine these approaches, implementing a model for peat accumulation and decay, as well as a parameterisation for wetland extent in order to asses the carbon accumulation in peat over the course of the Holocene.

2 Model description

2.1 CLIMBER2-LPJ

In the present study we are using the model CLIMBER2-LPJ as described in Kleinen et al. (2010). CLIMBER2-LPJ consists of the earth system model of intermediate complexity (EMIC) CLIMBER2, coupled to the DGVM LPJ. This combination of models allows experiments on timescales of an interglacial due to the low computational cost of CLIMBER2, while accounting for the heterogeneity of land surface processes on the much more highly resolved grid of LPJ.

CLIMBER2 (Petoukhov et al., 2000; Ganopolski et al., 2001) is an EMIC consisting of a 2.5-dimensional statistical-dynamical atmosphere with a latitudinal resolution of 10° and a longitudinal resolution of roughly 51°, an ocean model resolving three zonally averaged ocean basins with a latitudinal resolution of 2.5°, a sea ice model, and a dynamic terrestrial vegetation model (Brovkin et al., 2002). In the present model experiments, the latter model is used only for determining biogeophysical responses to climate change, while biogeochemical effects, i.e., the corresponding carbon fluxes, are determined by LPJ.

In addition, CLIMBER2 contains an oceanic biogeochemistry model, and a phosphate-limited model for marine biota (Ganopolski et al., 1998; Brovkin et al., 2002, 2007). The sediment model resolves the diffusive pore-water dynamics, assuming oxic only respiration and 4.5-order $CaCO_3$ dissolution kinetics (Archer, 1996; Brovkin et al., 2007). Weathering rates scale to runoff from the land surface grid cells, with separate carbonate and silicate lithological classes.

To this EMIC we have coupled the DGVM LPJ (Sitch et al., 2003; Gerten et al., 2004) in order to investigate land surface processes at a resolution significantly higher than the resolution of CLIMBER2. We also implemented carbon isotope fractionation according to Scholze et al. (2003).

LPJ is run on an 0.5° × 0.5° grid and is called at the end of every model year simulated by CLIMBER2. Monthly anomalies from the climatology of the temperature, precipitation and cloudiness fields are passed to LPJ, where they added to background climate patterns based on the Climatic Research Unit CRU-TS climate data set (New et al., 2000). In order to retain some temporal variability in these climate fields, the anomalies are not added to the climatology of the CRU data set, but rather to the climate data for one year randomly drawn from the range 1901–1930. LPJ simulates the changes in carbon pools and the carbon flux F_{AL} between atmosphere and land surface, which is passed back to CLIMBER2 and is employed to determine the atmospheric CO_2 concentration for the next model year. Biogeochemical feedbacks between atmosphere and land surface are thus determined by the combination

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of CLIMBER2 and LPJ, while biogeophysical effects are solely determined by the CLIMBER2 land surface model, which includes its own dynamical vegetation model.

2.2 Modelling peat accumulation

A natural peatland consists of three functionally distinct layers. At the top there is a live plant layer, where plants generate organic matter through photosynthesis. Below that is an upper layer of peat, which usually is less than 30 cm in height (Belyea and Baird, 2006), the so-called acrotelm, located above the permanent water table and therefore aerobic for at least part of the year. At the bottom is the peat located below the permanent water table, the so-called catotelm. This latter layer can be several meters in height, and significant amounts of carbon can therefore be stored in the peat column.

During peat formation, the process can be described as follows. The live plants at the surface generate organic matter through photosynthesis. Dead organic matter is added to a litter layer, from which it passes to the acrotelm very quickly. In the acrotelm the organic mater is decomposed, either aerobically or anaerobically, depending on the position of the water table, and once decomposition has proceeded far enough, the organic matter suffers a structural collapse, which significantly enhances density while shrinking pore volume. The water is squeezed out of the organic matter and the permanent water table rises slightly, adding some more organic material to the catotelm (Belyea and Baird, 2006).

In principle it is possible to model this process of peat formation explicitly, as Frolking et al. (2010) have shown. Modelling the change in density of the organic matter requires keeping track of annual cohorts of organic material in order to model how they pass through the peat column and to determine how density changes in each peat layer. This approach therefore requires substantial amounts of computer memory if implemented on a global or even hemispheric scale. We therefore decided against this approach but rather approximate the peat formation process by assuming a catotelm

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formation rate which is proportional to the amount of carbon in the acrotelm, an approach very similar to how soil organic matter decomposition is modelled in LPJ.

For rates of peat accumulation, as well as the distribution of peatlands and total peat storage, numerous estimates exist for boreal regions, while far fewer estimates exist for tropical peatlands. Since these estimates are essential for calibrating and validating the model, we currently limit our investigation to the regions north of 40° N.

2.3 Peat dynamics

If we compare the soil carbon dynamics in wetland and non-wetland soils, the main difference is that part of the soil column in wetland soils is below the water table, which leads to anaerobic decomposition of soil organic matter. LPJ contains various carbon pools, as shown in Fig. 1, on the left. There are live biomass pools for carbon (C) in leaves, wood and roots, here shown as a single pool C_B for simplicity. Then there are pools for carbon stored in litter, here summarised as C_L, and finally there are pools C_S for carbon stored in soil.

For the peat version of LPJ, this pool structure needs to be extended by an additional belowground C pool containing the carbon in the catotelm that is decomposing under anoxic conditions all year round, shown in Fig. 1, right hand side, as C_C.

In line with this structure of C pools, a vertical column of peat would have to be seen as shown in Fig. 1, on the right. At the surface there is a litter layer. Below that is the acrotelm, a soil layer where carbon is decomposed partly under oxic and partly under anoxic conditions, depending on the position of the water table. Finally, below the minimum water table position, there is a soil layer where anoxic decomposition occurs all year round, the catotelm.

These C pools can then be modelled as follows:

$$\frac{dC_L}{dt} = F_{BL} - F_{LA} - R_L \qquad (1)$$

$$\frac{dC_A}{dt} = F_{LA} - F_{AC} - R_{A,o} - R_{A,a} \qquad (2)$$

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$$\frac{dC_C}{dt} = F_{AC} - R_C \tag{3}$$

Here, the F_{XY} are the carbon fluxes between the C pools, and the R_k are the respiration fluxes to the atmosphere. XY can have the meanings BL = biomass-litter, LA = litter-acrotelm, and AC = acrotelm-catotelm, while k can be L = litter, A = acrotelm (with "o" for oxic and "a" for anoxic conditions), and C = catotelm. Leaching of dissolved organic carbon (DOC) is not considered explicitly, but rather assumed to be implicitly contained in the respiration fluxes.

In order to keep the peat version of LPJ as close to the original as possible, we keep flux formulations for existing carbon pools as they are in the original version, but the fluxes F_{AC} and R_C have to be added. The fluxes F_{XY} are dependent on the carbon mass (or rather area density) C_X in the originating C pool X, as well as a temperature (and moisture) dependent decay function, and the same holds for the respiration fluxes R_X. In concrete terms this is:

$$F_{LA} = \alpha k_L C_L \tag{4}$$

$$F_{AC} = k_P C_A \tag{5}$$

$$R_L = (1-\alpha) k_L C_L \tag{6}$$

$$R_{A,o} = \beta k_A C_A \tag{7}$$

$$R_{A,a} = (1-\beta) \upsilon k_A C_A \tag{8}$$

$$R_C = k_C C_C \tag{9}$$

with the rate constants k_i modified multiplicatively by a response function $\gamma(T_{soil}, w_{Soil})$ depending on soil temperature T_{soil}, as well as soil moisture w_{Soil}. Since we are considering wetland processes, we assume that moisture is not a limiting factor to decomposition and therefore do not consider a dependence on w_{Soil}. The temperature dependence of decomposition is quite often modelled as an exponential function $\exp\left(\ln(Q_{10})(T - T_{Ref})/T_{Ref}\right)$, and measured Q_{10} factors vary widely. Scanlon and Moore (2000), for example, measured Q_{10} values ranging from 1.0 to 7.7 for peat

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decomposition. For ecosystem respiration, on the other hand, recent results indicate that Q_{10} is 1.4, despite the huge range of measured Q_{10} determined in laboratory studies (Mahecha et al., 2010). LPJ generally uses the formulation by Lloyd and Taylor (1994) for temperature dependence, which gives a temperature dependence roughly similar to a Q_{10} of 2. We also apply it in wetland systems. Finally, α determines the fraction of decomposed litter that is added to the soil, while the remainder is respired. This is not changed from the non-wetland version of LPJ, either.

The case of the acrotelm respiration rate R_A is slightly more complicated, since acrotelm peat above the water table decomposes aerobically, while it decomposes anaerobically below. In Eq. (7) for the acrotelm respiration under oxic conditions, β is the fraction of the acrotelm above the water table, which decomposes at the rate k_A, while the rest (Eq. 8) decomposes anaerobically at the rate υk_A, with υ the ratio of anaerobic to aerobic CO_2 production. We follow Wania et al. (2009b) in setting this to 0.35. Finally, catotelm formation is modelled similar to decomposition, with the formation flux F_{AC} depending on the peat formation rate constant k_P, while decomposition depends on a decomposition constant k_C. Clymo et al. (1998) determined this rate constant by fitting peat core data to a similar decomposition model, and their value translates to $k_C = 3.35 \times 10^{-5}\,a^{-1}$ if corrected for mean annual temperature. The peat formation rate constant k_P, as well as the acrotelm respiration rate k_A, we determined by comparing model results to measured acrotelm mass (Malmer and Wallen, 1993) and peat formation rates (Yu et al., 2010).

The fraction β of the acrotelm above the water table is determined by comparing acrotelm height, calculated from acrotelm density ρ_A, acrotelm carbon fraction $c_{f,A}$ and acrotelm mass density C_A, and water table position w_w. The latter is assumed to be relative to the acrotelm-catotelm interface, which is located at the summer minimum water table position. The acrotelm density ρ_A was taken from Granberg et al. (1999), who give a density for the surface peat and for the peat at the bottom of the acrotelm. We therefore assume the mean of these values to be the acrotelm density. All parameter values used are listed in Table 1.

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With regard to biomass input into the peat column, we decided not to implement special wetland PFTs in the interest of keeping the model as close to the original as possible. Initial tests showed that the productivity of (modelled) mosses is very similar to grasses, making an additional PFT unnecessary. We did follow Wania et al. (2009b), though, in introducing their parameterisation for inundation stress in trees since tree growth is strongly inhibited in wetlands.

2.4 Dynamic wetland model

While LPJ's 0.5° resolution already is rather high in comparison to typical resolutions of climate model land surface schemes, this still translates to a grid cell size of 50 km × 30 km at 60° N. Since most wetlands are of smaller extent than this, an approach is required that determines the grid cell fraction covered by wetlands. Since it is our aim to apply the model to previous interglacials, as well as times earlier in the Holocene, using a simple wetland map to determine grid cell wetland area based on present day observations is insufficient. Instead, a scheme to determine wetland extent dynamically is required.

For this purpose we have implemented an approach based on the TOPMODEL hydrological framework (Beven and Kirkby, 1979). TOPMODEL is a conceptual rainfall-runoff model that is designed to work at the scale of large watersheds using the statistics of topography, instead of requiring detailed topographic information. It is based on the compound topographic index (CTI) $\chi_i = \ln\left(\alpha_i / \tan(\beta_i)\right)$ with α_i a dimensionless index for the area draining through point i and $\tan(\beta_i)$ the local slope at that point. The CTI can be derived from digital elevation models and near global datasets are readily available, for example the HYDRO1k data set in a resolution of 1 km (USGS, 1996). Following Sivapalan et al. (1987), we are actually not using the CTI values themselves, but we rather approximate the distribution of CTI values with in a grid cell by fitting a gamma distribution to them. The parameters of this distribution can be derived from grid cell CTI mean, standard deviation and skewness. While this approach may be less precise in a few grid cells, it greatly reduces the required input data.

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The central equation of TOPMODEL determines the local water table z_i in point i in relation to the grid cell mean water table $\bar{z}$:

$$z_i = \bar{z} + \frac{1}{f}(\chi_i - \bar{\chi}) \tag{10}$$

with χ_i the local CTI index in point i, $\bar{\chi}$ the grid cell mean CTI index, and f a parameter describing the exponential decline of transmissivity with depth. Using this equation we can determine the grid cell fraction that is innundated, i.e., with a water table at or above the surface, as well as the mean water table height in the innundated fraction. The innundated area consists of all points within the grid cell that have a local water table depth $z_i \geq 0$, but since a local water table that is well above the surface implies either running water, i.e., a river, or a lake, we also set a maximum CTI value χ_{max} which is constant in space and time, similar to Gedney and Cox (2003). We therefore assume the grid cell area with $0 \leq z_i \leq z_{\text{max}}$, $z_{\text{max}} = z(\chi_{\text{max}})$ to be the grid cell wetland area. Finally, the wetland water table position w_{w} is determined from Eq. (10), using the mean CTI index of the grid cell wetland fraction.

In order to determine the grid cell mean water table, we slightly modified the Stieglitz et al. (1997) approach to a formulation appropriate for LPJ:

$$\bar{z} = z_{\text{b}} - \left(\frac{\overline{w} - w_{\text{thres}}}{1 - w_{\text{thres}}} \right) \Delta z \tag{11}$$

with z_{b} the bottom of the soil column, Δz the height of the soil column, $\overline{w}$ the soil column average soil moisture, and w_{thres} the minimum soil moisture for a water table to form. This modification of the original approach became necessary, since soil moisture in LPJ is a variable determining the plant available water as a fraction of field capacity, i.e., $w = 0$ at the wilting point and $w = 1$ at field capacity. Similarly, since the LPJ soil column is very shallow having only 2 layers and a total soil column height of 1.5 m, we are simply using the soil column average soil moisture $\overline{w}$ instead of the layer soil moistures.

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Sensitivity experiments showed that a minimum soil moisture $w_{thres} = 0.1$ yielded good results, and that we could obtain the best match to observed innundated area by using $f = 2.3\,m^{-1}$, though results are not very sensitive to small variations in f. In addition, an initial comparison with Lehner and Döll (2004) wetland area showed that wetland area in grid cells with a mean CTI index $\bar{\chi} \leq 5.5$ is negligible. Wetlands are therefore only determined for grid cells with $\bar{\chi} > 5.5$.

The scheme described above determines the monthly grid cell fraction that is innundated. For peat to develop, areas that are just innundated for a few days during the course of the year are not relevant, but rather the areas that are innundated permanently, or at least during the growing season. Modelled soil water dynamics during the winter season, on the other hand, cannot be trusted in high latitude areas since LPJ does not consider permafrost, leading to too low water table positions during periods when the ground is frozen. We therefore use the minimum innundated area and water table position during the summer (JJAS) season to determine peatland extent.

The scheme to determine wetland extent described in this section is a dynamic scheme, i.e., wetland extent can change as the climatic conditions and therefore the soil moisture change. In order to limit the interannual fluctuations in peatland extent, we are using a 50 yr running mean peatland extent in all calculations. As peatland extent changes, carbon pools have to be updated. For these transfers of carbon between the wetland and the non-wetland part of the grid cell, we are making two assumptions. First we assume the peat deposit to have a parabolic overall shape, i.e., carbon storage from the edge to the centre of the peat deposit follows a parabola, and second we assume that a wetland that shrunk previously and then expands again expands into the same area it covered previously.

In case wetland extent shrinks, some of the carbon stored in the catotelm pool C_C will have to be passed into the soil pool of the non-wetland part of the grid cell. The fraction of C_C to be passed is determined from the proportional change in wetland size assuming that the peat deposit has an overall parabolic shape with peat removed from the outer edge. The carbon is then transferred into the soil carbon pool of the

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non-wetland part of the grid cell, but a record is kept of previous wetland extent and amount of carbon transferred from the catotelm pool in case the wetland should grow again.

In case of a growth in wetland extent, carbon is proportionally transferred from the soil carbon pool to the oxic wetland carbon pool. In case a wetland expands into areas previously covered by a wetland, carbon is also added to the anoxic pool C_C.

2.5 Model experiments

We have performed two sets of model experiments. One is a model experiment under constant preindustrial boundary conditions, while the other experiment is a transient fully coupled model run with interactive CO_2 for the last 8000 yr, from 8 ka BP to preindustrial. Both experiments were initialised from a 2000 yr spinup under the corresponding boundary conditions. For each experiment, we performed three different model runs at slightly changed parameter settings in order to determine a plausible range of model result. Since we estimate the wetland extent to be the most important uncertain parameter in the model, we vary the maximum CTI value χ_{max}. In these experiments, core$^-$ with $\chi_{max} = 9.7$ gives a wetland extent very similar to observed, parameter set core with $\chi_{max} = 10.0$ gives a wetland extent about 10 % larger than observed for present day, which we estimate to be similar to preindustrial extent before anthropogenic drainage of wetlands, and parameter set core$^+$ with $\chi_{max} = 10.4$ obtains a wetland area 20 % larger than observed.

We only consider peat accumulation in the high northern latitudes, i.e., north of 40° N. Tropical peatlands are not considered at the moment.

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3 Results

3.1 Wetland extent

Data sets of global wetland areas are rare and uncertainties in these tend to be large. For validation of the wetland distribution the model produces, we are relying on two data sets. We used the global lakes and wetlands database (GLWD) (Lehner and Döll, 2004), which shows the annual maximum extent of wetlands, based on a combination of maps and remote sensing data. In addition, we applied the data set of remotely sensed surface water extent by Prigent et al. (2007). The latter shows the monthly surface water extent from January 1993 to December 2000. For wetlands in the high latitudes one has to keep in mind, though, that snow cover will mask wetlands during the snow season, making surface water extent measurements impossible during this time. For the annual maximum extent, the two data sets agree reasonably well (Papa et al., 2010). When it comes to the total annual maximum innundated area north of 40° N, modelled extent in experiment core$^-$ is 3 % larger than wetland extent in Prigent et al. (2007), and 15 % larger than GLWD. The total maximum innundated area therefore is captured quite reasonably. Two additional sensitivity experiments have wetland extent increased by 10 (core) and 20 % (core$^+$) relative to the base model. The rationale for these sensitivity experiments is that the wetland extent is, in our opinion, the most uncertain parameter in our model. In addition it is well known that some wetland areas have been drained since the beginning of the industrial era. Comparing the maximum wetland extent in the Prigent et al. (2007) data set to areas known to have been drained, we estimate that wetland area is underestimated by at least 10 % in these data, but less than 20 %. A wetland extent 10 % larger than in Prigent et al. (2007), as in experiment core, therefore appears to be a best estimate for preindustrial wetland area.

If we consider latitudinal variations in wetland area, shown as zonal sums in Fig. 2a, the model overestimates wetland area between 40° N and 53° N, while it underestimates wetland extent around 55° N. Investigating this more closely, it turns out that the

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model underestimates wetland extent in large wetland areas, like the Hudson's Bay lowlands in Canada and the western Siberian wetlands in Russia, while it overestimates wetlands elsewhere. While the latter could in principle be changed by adjusting χ_{max}, the underestimation of wetland extent in regions with extensive areas of relatively flat terrain is a shortcoming of the TOPMODEL approach we are using to estimate wetland area. It follows from Eq. (10) that the TOPMODEL approach redistributes the available water within a grid cell. Therefore the maximum grid cell fraction that can become a wetland is limited. In the underlying LPJ hydrology, grid cell water content is limited to field capacity. The grid cell mean water table therefore never is above the surface, and even if the grid cell water table is near the surface, some fraction of the grid cell will always have a water table below the surface.

So far we have discussed the maximum wetland extent, since estimates exist from two independent data source. For peat formation, on the other hand, the summer minimum water table and wetland extent is more important, since this determines peatland area. The modelled total wetland extent at the summer (JJAS) minimum water table position once again is similar to the remote sensing estimate (4 % smaller than Prigent et al., 2007), while the latitudinal distribution, shown in Fig. 2b, is somewhat different, the area is underestimated between 45° N and 65° N, while it is overestimated north of 65° N. For the summer minimum wetland extent, the difference in total area is larger than for the annual maximum extent, though: experiment core has a total area 15 % larger than Prigent et al. (2007), and core^{+} is 37 % larger.

Over the course of the Holocene, climate changed significantly. Due to the decrease in high latitude insolation through orbital changes, temperatures and therefore evapotranspiration decreased, while precipitation mostly increased in the mid to high latitudes. This would lead to wetter soils, a higher grid cell mean water table, and also a larger wetland extent. Model results for the last 8 ka, shown in Fig. 3, reflect the expected changes. Wetland area was smaller in most high latitude regions, especially pronounced in Eastern Canada and western Siberia, but also in Scandinavia. Contrary to this, wetlands just north of 40° N decreased.

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3.2 Peat accumulation: the acrotelm

In the wetland areas the model accumulates carbon as peat, i.e., organic matter that decomposes very slowly since decomposition takes place under anaerobic conditions. Contrary to data on catotelm carbon, very few studies exist on the acrotelm. We are aware of only a single study comparing sites at different locations with a common methodology. In this study Malmer and Wallen (1993) investigate the acrotelm at 12 sites in Canada, Scotland, and Scandinavia. Malmer and Wallen measured carbon to nitrogen ratios in the peat cores. The ratio first decreases quickly as one goes deeper in the peat, due to the fact that carbon decomposes, while nitrogen is conserved. Further down in the core, it decreases more slowly, and Malmer and Wallen identified the intersect of the two C/N trends as the "decay decrease level", i.e. the transition from acrotelm to catotelm. In their study, they determine decomposition rates for the acrotelm and report both the height of the acrotelm, as well as the amount of organic matter contained therein.

The density varies quite widely between the sites investigated, which is in line with other studies reporting that peat bulk density is highly dependent on the source of organic material going into the peat, as summarised by Kobak et al. (1998). Since our very limited number of PFTs cannot be expected to reproduce plant composition at the sites, we therefore focus on acrotelm organic matter as a measure to compare our model results to.

The acrotelm organic matter accumulated in LPJ compares reasonably well to the site data reported by Malmer and Wallen (1993) (Fig. 4). The spread of values is larger for the measured data than for the model results, but model results generally scatter around the equal mass line. Model results therefore are far from perfect, but they show reasonable agreement with measured values.

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3.3 Peat accumulation: the catotelm

For the peat accumulated in the catotelm, numerous studies exist. They range in scope from studies of single sites, as in Yu (2006), for example, over regional summaries for Siberia (Kremenetski et al., 2003; Beilman et al., 2009) and North America (Gorham et al., 2003) to a recently published global summary of peat accumulation rates by Yu et al. (2010). While the single core publications usually are rather difficult to compare due to different conventions used, as well as different measures reported, the regional and global summaries employ standard methodologies and can therefore be compared rather well. While Yu et al. (2010) report catotelm accumulation rates, Gorham et al. (2003); Kremenetski et al. (2003) and Beilman et al. (2009) report the basal date and basal depth of the peat accumulated. From this we determined the long term apparent rate of carbon accumulation (LORCA) by dividing the basal depth by the basal date in years BP and converting the height accumulation rates into a carbon accumulation rate by using the C fraction and density of catotelm peat. If multiple measurements existed for a single LPJ grid cell, we compared to the average.

This comparison, shown in Fig. 5 for experiment core, shows a good agreement between measurements and modelled values overall, though measured accumulation rates again scatter more widely around the equal accumulation line. This is to expected, though, since measured accumulation rates depend on local conditions which cannot always be captured in a $0.5° \times 0.5°$ model grid cell.

Related to the the grid cell size, as opposed to the wetland fraction, modelled increases in carbon density over the course of the experiment, i.e., between 8 ka BP and present day, are quite variable (Fig. 6). Large amounts of catotelm peat, up to 30 $\mathrm{kgC\,m^{-2}}$, have accumulated in Scandinavia. Western Siberia and eastern Canada show slightly lower values of about 25 $\mathrm{kgC\,m^{-2}}$, while most other areas show substantially lower accumulations. In terms of peat height (not shown), up to 6m of peat have accumulated in eastern Europe, as well as in North America south of the Hudson's Bay and on the British Isles. With peat heights between 3 and 4.5 m, peat accumulation in

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Western Siberia and Scandinavia is slightly less, which is similar along the North Sea coast. Peat accumulation in the Canadian Arctic, as well as Eastern Siberia, is rather small, though not quite negligible, but peat accumulation in the Asian wetland areas at the southern boundary of the study domain, in Kazakhstan and the surrounding areas, is very small.

3.4 Carbon uptake

From the point of view of Holocene carbon cycle dynamics, the final important question is how much carbon is actually taken up and stored in peatlands. Within the acrotelm, i.e., the part of the peat column that is under oxic conditions for at least part of the year, shown in Fig. 7 a, there is a total of between 7.1 and 9.6 PgC stored at 8 ka BP, depending on sensitivity experiment, with core^{-} showing the lowest and core^{+} the highest amount of acrotelm C. This total increases by about 1.3 PgC over the last 8000 yr. Carbon storage within the catotelm, i.e., the permanently anoxic part of the peat column, the peat storage, increases by about 174 PgC in core^{-}, 209 PgC in core, and 252 PgC in core^{+}, as shown in Fig. 7b. Carbon accumulation in peatlands therefore is roughly 210 ± 40 PgC for the peatland areas north of 40° N.

4 Discussion

A dynamic model of wetland extent and peat accumulation, such as the one we have constructed, is extremely difficult, if not impossible to validate. Data on wetland extent and location is available from some national agencies, but these national inventories quite often are compiled using different measures and methodology and therefore are not comparable. Existing syntheses like the Lehner and Döll (2004) data, used for evaluation in the present study, have attempted to bridge these different methodologies in order to determine global estimates of wetland area, but uncertainties remain large. In addition this data set gives the maximum wetland extent, not the permanent wetland

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area. Since the latter is relevant for peat formation, evaluation of the modelled peatland area is complicated further. We have attempted to remedy the situation by using a data set of innundated area determined by remote sensing (Prigent et al., 2007), which compares favourably with Lehner and Döll (2004) for the maximum extent (Papa et al., 2010). Using the summer minimum extent gives an approximation of peatland area, but remote sensing data is uncertain as well. Any kind of ground cover, be it trees or snow, makes remote sensing of wetland extent impossible. A further complication is that these data sets describe the present day situation, and many wetlands in Europe as well as the more densely populated parts of North America have been drained in order to convert them to agricultural use. Our model, on the other hand, aims at determining the natural extent of wetlands, and anthropogenic disturbances are not taken into account.

The approach we have chosen to determine wetland area is relatively simple and very much dependent on the hydrology of the underlying model. The latter has been evaluated positively with regard to streamflow (Gerten et al., 2004), but the rather shallow two-layer soil column in LPJ in combination with the limitation of soil moisture to field capacity and the lack of permafrost dynamics give reason for doubt with regard to soil water table dynamics. This is impossible to evaluate, though, since measurements of soil moisture exist only in very few places. Finally, the TOPMODEL approach itself is limiting as well, since a redistribution of the water within the grid cell, in combination with a grid cell mean water table that is always below the surface, implicitly limits the maximum wetland extent per grid cell.

We are not aware of previously published attempts at determining the extent of permanent wetlands in a dynamic way. Previous studies either focused on peat accumulation at single sites (Clymo, 1984; Frolking et al., 2001, 2010; Borren and Bleuten, 2006; Frolking and Roulet, 2007; St-Hilaire et al., 2010), which does not require an estimate of wetland area, or on methane emissions (Kaplan, 2002; Gedney et al., 2004; van Huissteden et al., 2006; Bohn et al., 2007; Wania et al., 2009a,b, 2010; Ringeval et al., 2010) from wetlands, which requires an estimate of innundated area, but not of the

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permanent wetland area. These approaches therefore either are map-based, or rely on a TOPMODEL approach similar to ours to determine the innundated area but do not specify permanently innundated areas, which are relevant for peat formation. Due to all these factors, the extent of permanent wetlands our model determines appears to be the largest uncertainty, though this is impossible to quantify. We tried to take this uncertainty into account by performing three sensitivity experiments, of which we assume experiment core, to best capture the preindustrial peatland extent and experiments core^{-} and core^{+} spanning the uncertainty range.

Despite the simplicity of the approach we have chosen to model peat dynamics, which neglects potentially important factors like soil pH, exact species composition of aboveground vegetation, litter composition, etc., comparison to the little measurement data that exists appears quite favourable. Acrotelm mass in measurements (Malmer and Wallen, 1993) varies widely, and our model seems to fit these data reasonably well. Similarly, catotelm accumulation rates appear similar to peat core measurements (Kremenetski et al., 2003; Beilman et al., 2009; Yu et al., 2010), which once again vary quite widely.

The model estimate of total carbon storage in peatlands (210 ± 40 PgC accumulated since 8 ka BP) is at the low end of global estimates (Gajewski et al., 2001; Turunen et al., 2002; Yu et al., 2010). These latter estimates are based on point measurements that are upscaled by an estimate of peatland area, though, and therefore are uncertain as well. Therefore our result is still quite reasonable.

The modelled carbon storage very likely is a slight underestimate since a number of factors are neglected in our experiments. The peat accumulation shown in Fig. 7 is the peat accumulation over the last 8000 yr, a timescale we chose since most of the continental ice sheets were melted at the time the model was initialised for. A significant number of peatlands started growing earlier than this, though, especially in areas that were not covered by ice sheets during the last glacial. In some other places like coastal areas of Scandinavia and the Hudson's Bay lowlands in Canada, though, peatland initiation took place later, usually because those areas were depressed below

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sea level by the overlying glacial ice sheet, and postglacial rebound took some time to elevate these areas above sea level. These processes are not considered in our model, and therefore we cannot estimate a total peat storage, just peat accumulation during the last 8000 yr.

5 Summary and conclusions

We have extended the coupled climate carbon cycle model CLIMBER2-LPJ by a module determining permanent wetland extent and peat accumulation north of 40° N. Wetland area, acrotelm magnitude and catotelm peat accumulation seem to agree reasonably well with the little measurement data that exists, giving us some confidence in model results. We initialised the model for conditions at 8000 yr before present, and determined the evolution of climate, wetland extent and peat accumulation until the present, assuming preindustrial conditions at present day.

Over the course of the 8000 yr, wetland areas increased by up to 5 % relative to the grid cell in Siberia and Eastern Canada and 1 % in Europe. Wetland areas at the southern end of the study domain either shrunk or did not change. The change in permanent wetland extent therefore reflects the changes in climate over the the course of the Holocene, a decrease in summer temperature and an increase in precipitation.

During this time our model accumulates about 210 ± 40 PgC as catotelm peat in the areas above 40° N, with the uncertainty range stemming from sensitivity experiments with peatland area varied by ±10 % to capture the largest uncertainty. This estimate is somewhat larger than the scenario we used previously when investigating Holocene carbon cycle dynamics (Kleinen et al., 2010), but still compatible with our previous conclusion that the Holocene rise in CO_2 can be explained by invoking natural causes only. The model we developed will allow estimates of carbon uptake by peatlands in climate states different from the present interglacial, for example in glacial climate or in previous interglacials.

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References

Archer, D. E.: A data-driven model of the global calcite lysocline, Global Biogeochem. Cy., 10, 511–526, 1996. 4809

Beilman, D. W., MacDonald, G. M., Smith, L. C., and Reimer, P. J.: Carbon accumulation in peatlands of West Siberia over the last 2000 years, Global Biogeochem. Cy., 23, GB1012, doi:10.1029/2007GB003112, 2009. 4821, 4824, 4837

Belyea, L. R. and Baird, A. J.: Beyond "The limits to peat bog growth": cross-scale feedback in peatland development, Ecol. Monogr., 76, 299–322, 2006. 4810

Beven, K. J. and Kirkby, M. J.: A physically based, variable contributing area model of basin hydrology, Hydrological Sciences Bulletin, 24, 43–69, 1979. 4808, 4814

Bohn, T. J., Lettenmaier, D. P., Sathulur, K., Bowling, L. C., Podest, E., McDonald, K. C., and Friborg, T.: Methane emissions from western Siberian wetlands: heterogeneity and sensitivity to climate change, Environ. Res. Lett., 2, 045015, doi:10.1088/1748-9326/2/4/045015, 2007. 4823

Borren, W. and Bleuten, W.: Simulating Holocene carbon accumulation in a western Siberian watershed mire using a three-dimensional dynamic modeling approach, Water Resour. Res., 42, W12413, doi:10.1029/2006WR004885, 2006. 4823

Brovkin, V., Bendtsen, J., Claussen, M., Ganopolski, A., Kubatzki, C., Petoukhov, V., and Andreev, A.: Carbon cycle, vegetation, and climate dynamics in the Holocene: Experiments with the CLIMBER-2 model, Global Biogeochem. Cy., 16, 1139, doi:10.1029/2001GB001662, 2002. 4809

Brovkin, V., Ganopolski, A., Archer, D., and Rahmstorf, S.: Lowering of glacial atmospheric CO_2 in response to changes in oceanic circulation and marine biogeochemistry, Paleoceanography, 22, PA4202, doi:10.1029/2006PA001380, 2007. 4809

Clymo, R. S.: The limits to peat bog growth, Philos. T. R. Soc. Lon. B, 303, 605–654, 1984. 4807, 4823

Clymo, R. S., Turunen, J., and Tolonen, K.: Carbon accumulation in peatland, Oikos, 81, 368–388, 1998. 4807, 4813, 4832

Decharme, B. and Douville, H.: Introduction of a sub-grid hydrology in the ISBA land surface model, Clim. Dynam., 26, 65–78, doi:10.1007/s00382-005-0059-7, 2006. 4808

Frolking, S. and Roulet, N. T.: Holocene radiative forcing impact of northern peatland carbon accumulation and methane emissions, Glob. Change Biol., 13, 1079–1088, doi:10.1111/j.1365-2486.2007.01339.x, 2007. 4823

Frolking, S., Roulet, N. T., Moore, T. R., Richard, P. J. H., Lavoie, M., and Muller, S. D.: Modeling Northern Peatland Decomposition and Peat Accumulation, Ecosystems, 4, 479–498, doi:10.1007/s10021-001-0105-1, 2001. 4823

Frolking, S., Roulet, N. T., Tuittila, E., Bubier, J. L., Quillet, A., Talbot, J., and Richard, P. J. H.: A new model of Holocene peatland net primary production, decomposition, water balance, and peat accumulation, Earth Syst. Dynam., 1, 1–21, doi:10.5194/esd-1-1-2010, 2010. 4807, 4810, 4823

Gajewski, K., Viau, A., Sawada, M., Atkinson, D., and Wilson, S.: Sphagnum peatland distribution in North America and Eurasia during the past 21,000 years, Global Biogeochem. Cy., 15, 297–310, 2001. 4806, 4824

Ganopolski, A., Rahmstorf, S., Petoukhov, V., and Claussen, M.: Simulation of modern and glacial climates with a coupled global climate model, Nature, 391, 351–356, 1998. 4809

Ganopolski, A., Petoukhov, V., Rahmstorf, S., Brovkin, V., Claussen, M., Eliseev, A., and Kubatzki, C.: CLIMBER-2: A climate system model of intermediate complexity. Part II: Model sensitivity, Clim. Dynam., 17, 735–751, 2001. 4809

Gedney, N. and Cox, P. M.: The Sensitivity of Global Climate Model Simulations to the Representation of Soil Moisture Heterogeneity, J. Hydrometeorol., 4, 1265–1275, doi:10.1175/1525-7541(2003)004<1265:TSOGCM>2.0.CO;2, 2003. 4808, 4815

Gedney, N., Cox, P. M., and Huntingford, C.: Climate feedback from wetland methane emissions, Geophys. Res. Lett., 31, L20503, doi:10.1029/2004GL020919, 2004. 4808, 4823

Gerten, D., Schaphoff, S., Haberlandt, U., Lucht, W., and Sitch, S.: Terrestrial vegetation and water balance – hydrological evaluation of a dynamic global vegetation model, J. Hydrol., 286, 249–270, doi:10.1016/j.jhydrol.2003.09.029, 2004. 4809, 4823

Gorham, E., Janssens, J. A., and Glaser, P. H.: Rates of peat accumulation during the post-glacial period in 32 sites from Alaska to Newfoundland, with special emphasis on northern Minnesota, Can. J. Botany, 81, 429–438, doi:10.1139/b03-036, 2003. 4821, 4837

Granberg, G., Grip, H., Löfvenius, M. O., Sundh, I., Svensson, B. H., and Nilsson, M.: A simple model for simulation of water content, soil frost, and soil temperatures in boreal mixed mires, Water Resour. Res., 35, 3771–3782, doi:10.1029/1999WR900216, 1999. 4813, 4832

Habets, F. and Saulnier, G. M.: Subgrid runoff parameterization, Phys. Chem. Earth. Pt. B, 26, 455–459, doi:10.1016/S1464-1909(01)00034-X, 2001. 4808

Ingram, H. A. P.: Size and shape in raised mire ecosystems: a geophysical model, Nature, 297, 300–303, doi:10.1038/297300a0, 1982. 4807

Jobbagy, E. G. and Jackson, R. B.: The vertical distribution of soil organic carbon and its relation to climate and vegetation, Ecol. Appl., 10, 423–436, doi:10.1890/1051-0761(2000)010[0423:TVDOSO]2.0.CO;2, 2000. 4806

Kaplan, J. O.: Wetlands at the Last Glacial Maximum: Distribution and methane emissions, Geophys. Res. Lett., 29, 1079, doi:10.1029/2001GL013366, 2002. 4808, 4823

Kleinen, T., Brovkin, V., von Bloh, W., Archer, D., and Munhoven, G.: Holocene carbon cycle dynamics, Geophys. Res. Lett., 37, L02705, doi:10.1029/2009GL041391, 2010. 4807, 4808, 4825

Kobak, K. I., Kondrasheva, N. Y., and Turchinovich, I. E.: Changes in carbon pools of peatland and forests in northwestern Russia during the Holocene, Global Planet. Change, 16–17, 75–84, doi:10.1016/S0921-8181(98)00011-3, 1998. 4820

Korhola, A., Alm, J., Tolonen, K., Turunen, J., and Jungner, H.: Three-dimensional reconstruction of carbon accumulation and CH_4 emission during nine millennia in raised mire, J. Quaternary Sci., 11, 161–165, 1996. 4807

Kremenetski, K. V., Velichko, A. A., Borisova, O. K., MacDonald, G. M., Smith, L. C., Frey, K. E., and Orlova, L. A.: Peatlands of the Western Siberian lowlands: current knowledge on zonation, carbon content and Late Quaternary history, Quaternary Sci. Rev., 22, 703–723, doi:10.1016/S0277-3791(02)00196-8, 2003. 4821, 4824, 4837

Lehner, B. and Döll, P.: Development and validation of a global database of lakes, reservoirs and wetlands, J. Hydrol., 296, 1–22, 2004. 4816, 4818, 4822, 4823

Lloyd, J. and Taylor, J. A.: On the Temperature Dependence of Soil Respiration, Funct. Ecol.,

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8, 315–323, 1994. 4813
Mahecha, M. D., Reichstein, M., Carvalhais, N., Lasslop, G., Lange, H., Seneviratne, S. I., Vargas, R., Ammann, C., Arain, M. A., Cescatti, A., Janssens, I. A., Migliavacca, M., Montagnani, L., and Richardson, A. D.: Global Convergence in the Temperature Sensitivity of Respiration at Ecosystem Level, Science, 329, 838–840, doi:10.1126/science.1189587, 2010. 4813
Malmer, N. and Wallen, B.: Accumulation and release of organic matter in ombrotrophic bog hummocks – processes and regional variation., Ecography, 16, 193–211, 1993. 4813, 4820, 4824, 4836
Malmer, N. and Wallen, B.: Input rates, decay losses and accumulation rates of carbon in bogs during the last millennium: internal processes and environmental changes, The Holocene, 14, 111–117, doi:10.1191/0959683604hl693rp, 2004. 4832
New, M., Hulme, M., and Jones, P.: Representing Twentieth-Century Space-Time Climate Variability. Part II: Development of 1901–96 Monthly Grids of Terrestrial Surface Climate, J. Climate, 13, 2217–2238, 2000. 4809
Niu, G.-Y., Yang, Z.-L., Dickinson, R. E., and Gulden, L. E.: A simple TOPMODEL-based runoff parameterization (SIMTOP) for use in global climate models, J. Geophys. Res., 110, D21106, doi:10.1029/2005JD006111, 2005. 4808
Papa, F., Prigent, C., Aires, F., Jimenez, C., Rossow, W. B., and Matthews, E.: Interannual variability of surface water extent at the global scale, 1993–2004, J. Geophys. Res., 115, D12111, doi:10.1029/2009JD012674, 2010. 4818, 4823
Petoukhov, V., Ganopolski, A., Brovkin, V., Claussen, M., Eliseev, A., Kubatzki, C., and Rahmstorf, S.: CLIMBER-2: A climate system model of intermediate complexity. Part I: Model description and performance for present climate, Clim. Dynam., 16, 1–17, 2000. 4809
Prigent, C., Papa, F., Aires, F., Rossow, W. B., and Matthews, E.: Global inundation dynamics inferred from multiple satellite observations, 1993–2000, J. Geophys. Res., 112, D12107, doi:10.1029/2006JD007847, 2007. 4818, 4819, 4823
Ringeval, B., de Noblet-Ducoudré, N., Ciais, P., Bousquet, P., Prigent, C., Papa, F., and Rossow, W. B.: An attempt to quantify the impact of changes in wetland extent on methane emissions on the seasonal and interannual time scales, Global Biogeochem. Cy., 24, GB2003, doi:10.1029/2008GB003354, 2010. 4808, 4823
Scanlon, D. and Moore, T.: Carbon Dioxide Production From Peatland Soil Profiles: the Influence of Temperature, Oxic/Anoxic Conditions and Substrate, Soil Sci., 165, 153–160, 2000. 4812

Scholze, M., Kaplan, J. O., Knorr, W., and Heimann, M.: Climate and interannual variability of the atmosphere-biosphere $^{13}CO_2$ flux, Geophys. Res. Lett., 30, 1097, doi:10.1029/2002GL015631, 2003. 4809

Sitch, S., Smith, B., Prentice, I. C., Arneth, A., Bondeau, A., Cramer, W., Kaplan, J. O., Levis, S., Lucht, W., Sykes, M. T., Thonicke, K., and Venevsky, S.: Evaluation of ecosystem dynamics, plant geography and terrestrial carbon cycling in the LPJ dynamic global vegetation model, Glob. Change Biol., 9, 161–185, 2003. 4809

Sivapalan, M., Beven, K., and Wood, E. F.: On Hydrologic Similarity. 2. A Scaled Model of Storm Runoff Production, Water Resour. Res., 23, 2266–2278, doi:10.1029/WR023i012p02266, 1987. 4814

St-Hilaire, F., Wu, J., Roulet, N. T., Frolking, S., Lafleur, P. M., Humphreys, E. R., and Arora, V.: McGill wetland model: evaluation of a peatland carbon simulator developed for global assessments, Biogeosciences, 7, 3517–3530, doi:10.5194/bg-7-3517-2010, 2010. 4823

Stieglitz, M., Rind, D., Famiglietti, J., and Rosenzweig, C.: An Efficient Approach to Modeling the Topographic Control of Surface Hydrology for Regional and Global Climate Modeling, J. Climate, 10, 118–137, 1997. 4815

Turunen, J., Tomppo, E., Tolonen, K., and Reinikainen, A.: Estimating carbon accumulation rates of undrained mires in Finland – application to boreal and subarctic regions, The Holocene, 12, 69–80, doi:10.1191/0959683602hl522rp, 2002. 4806, 4824, 4832

USGS: HYDRO1k Elevation Derivative Database, uS Geological Survey Earth Resources Observation and Science (EROS) Data Center (EDC), Sioux Falls, South Dakota, USA, 1996. 4814

van Huissteden, J., van den Bos, R., and Alvarez, I. M.: Modelling the effect of water-table management on CO_2 and CH_4 fluxes from peat soils, Neth. J. Geosci., 85, 3–18, 2006. 4823

Wang, Y., Roulet, N. T., Frolking, S., and Mysak, L. A.: The importance of Northern Peatlands in global carbon systems during the Holocene, Clim. Past, 5, 683–693, doi:10.5194/cp-5-683-2009, 2009. 4807

Wania, R., Ross, I., and Prentice, I. C.: Integrating peatlands and permafrost into a dynamic global vegetation model: 1. Evaluation and sensitivity of physical land surface processes, Global Biogeochem. Cy., 23, GB3014, doi:10.1029/2008GB003412, 2009a. 4808, 4823

Wania, R., Ross, I., and Prentice, I. C.: Integrating peatlands and permafrost into a dynamic global vegetation model: 2. Evaluation and sensitivity of vegetation and carbon cycle processes, Global Biogeochem. Cy., 23, GB3015, doi:10.1029/2008GB003413, 2009b. 4808, 4813, 4814, 4823, 4832

Wania, R., Ross, I., and Prentice, I. C.: Implementation and evaluation of a new methane model within a dynamic global vegetation model: LPJ-WHyMe v1.3.1, Geosci. Model Dev., 3, 565–584, doi:10.5194/gmd-3-565-2010, 2010. 4823

Yu, Z.: Holocene Carbon Accumulation of Fen Peatlands in Boreal Western Canada: A Complex Ecosystem Response to Climate Variation and Disturbance, Ecosystems, 9, 1278–1288, doi:10.1007/s10021-006-0174-2, 2006. 4821

Yu, Z., Loisel, J., Brosseau, D. P., Beilman, D. W., and Hunt, S. J.: Global peatland dynamics since the Last Glacial Maximum, Geophys. Res. Lett., 37, L13402, doi:10.1029/2010GL043584, 2010. 4813, 4821, 4824, 4837

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Table 1. Parameter values in peat module.

Parameter	description	value	Reference
k_A	acrotelm decomposition rate	$0.067\,a^{-1}$	this study
k_C	catotelm decomposition rate	$3.35 \times 10^{-5}\,a^{-1}$	Clymo et al. (1998)
k_P	catotelm formation rate	$1.91 \times 10^{-2}\,a^{-1}$	this study
υ	ratio anaerobic-aerobic CO_2	0.35	Wania et al. (2009b)
ρ_A	acrotelm density	$3.5 \times 10^4\,g\,m^{-3}$	Granberg et al. (1999)
ρ_C	catotelm density	$9.1 \times 10^4\,g\,m^{-3}$	Turunen et al. (2002)
$c_{f,A}$	carbon fraction in acrotelm peat	0.50	Malmer and Wallen (2004)
$c_{f,C}$	carbon fraction in catotelm peat	0.52	Malmer and Wallen (2004)

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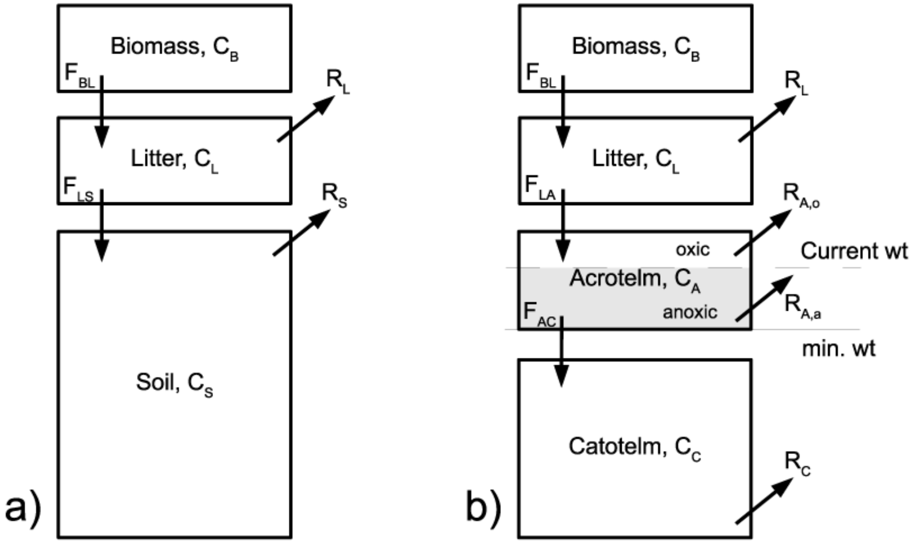

Fig. 1. The LPJ soil carbon pools C_X and fluxes F_{XY} and R_k. **(a)** non-wetland soil, **(b)** wetland soil. Suffixes X, Y, k designate the carbon pools with B (biomass), L (Litter), S (Soil), A (Acrotelm) and C (Catotelm).

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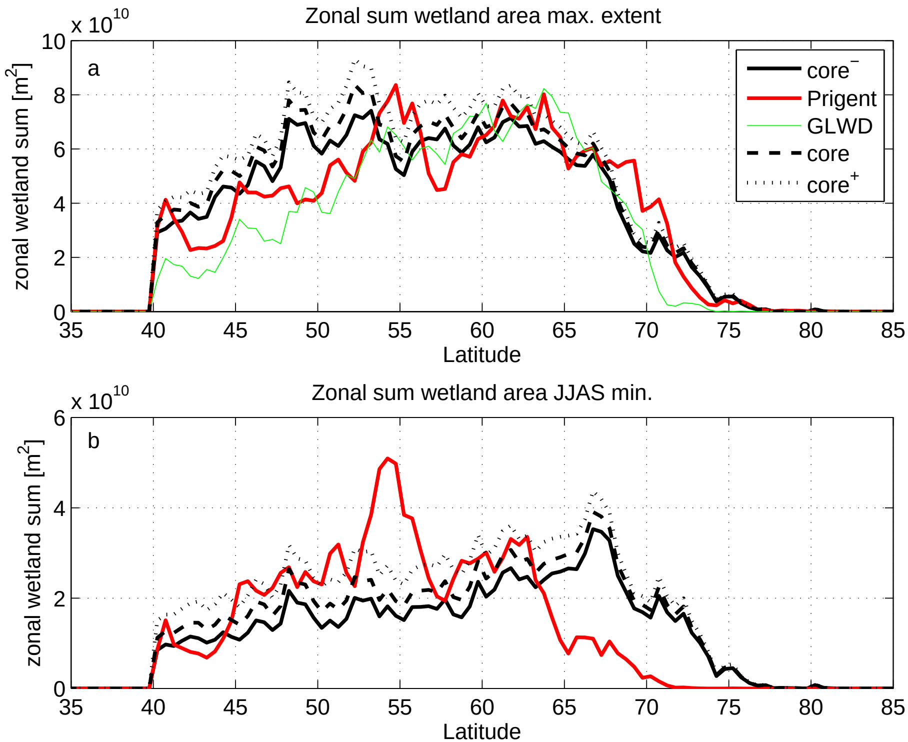

Fig. 2. Zonal sums of wetland extent for model configuration core^{-} (black), Prigent et al. (2007) (red), and GLWD (green), as well as model sensitivity experiments core (black dashed) and core^{+} (black dotted). Upper panel **(a)**: maximum extent, lower panel **(b)**: summer (JJAS) minimum extent.

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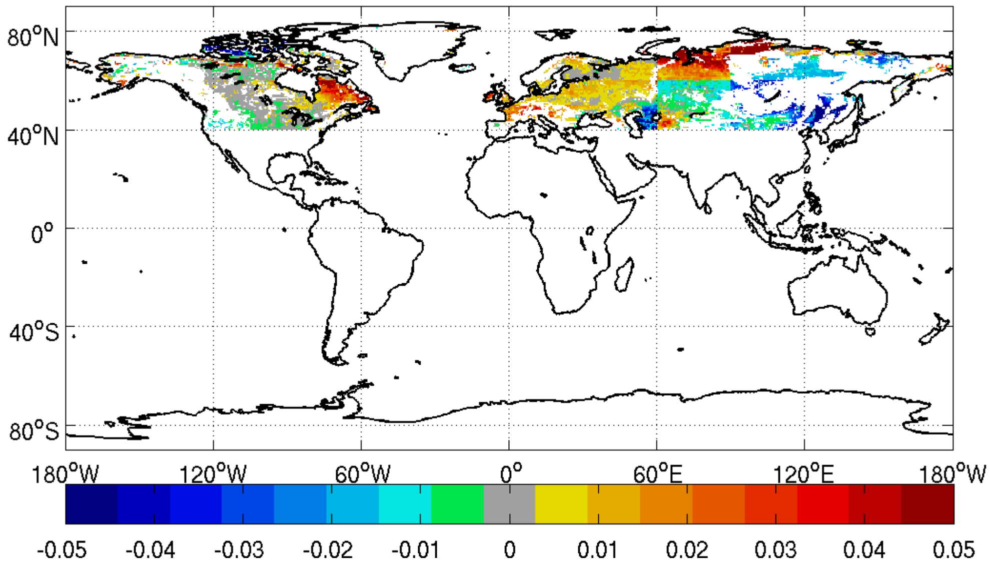

Fig. 3. Change in grid cell wetland fraction over the last 8 ka in experiment core.

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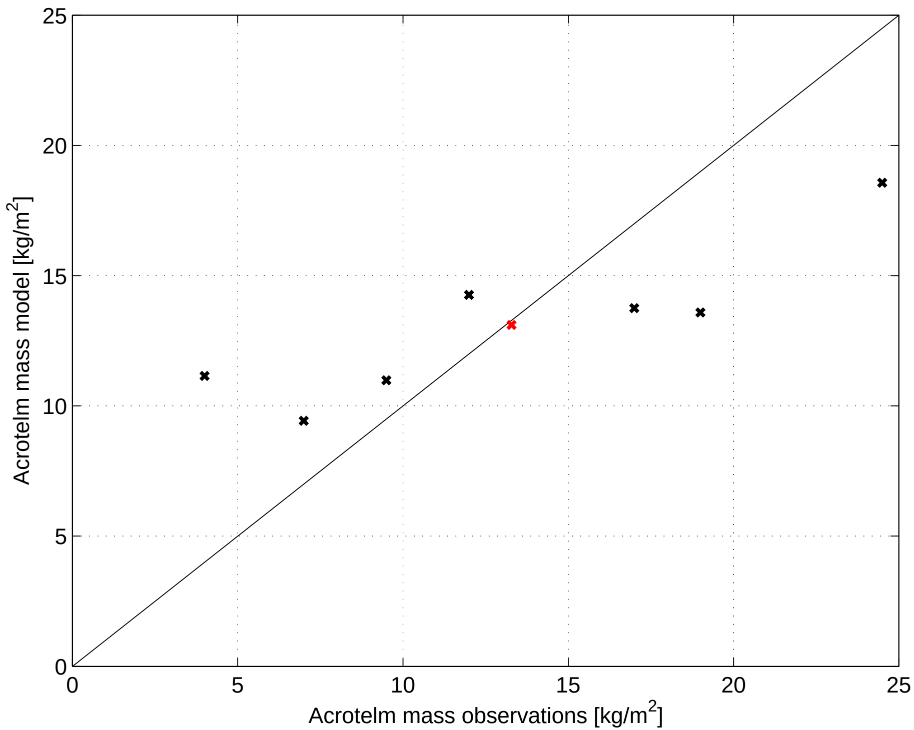

Fig. 4. Acrotelm organic matter accumulated in LPJ as a mean of the grid points surrounding the sites reported by Malmer and Wallen (1993) against the organic matter measured at the sites. Mean values are shown in red. The plot shows values from experiment core, with the other experiments showing very similar results.

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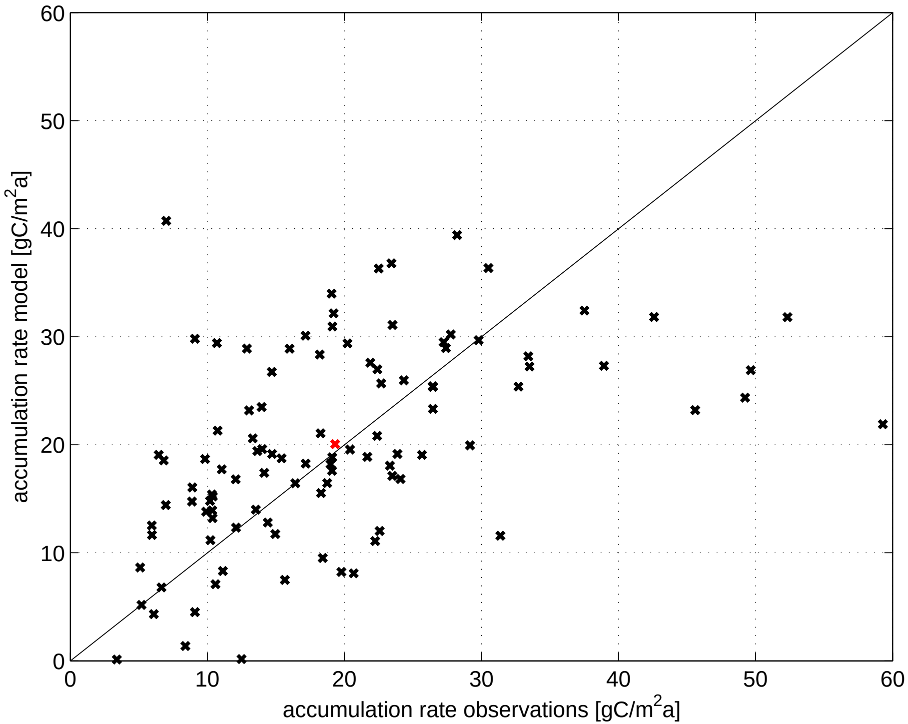

Fig. 5. Model catotelm peat accumulation rates in experiment core compared to measured values. Measurements compiled from Yu et al. (2010), reporting accumulation rates, and Gorham et al. (2003); Kremenetski et al. (2003) and Beilman et al. (2009), where we converted basal date and peat height into accumulation rate. Averages are shown for grid cells containing multiple measurements. Mean values are shown in red.

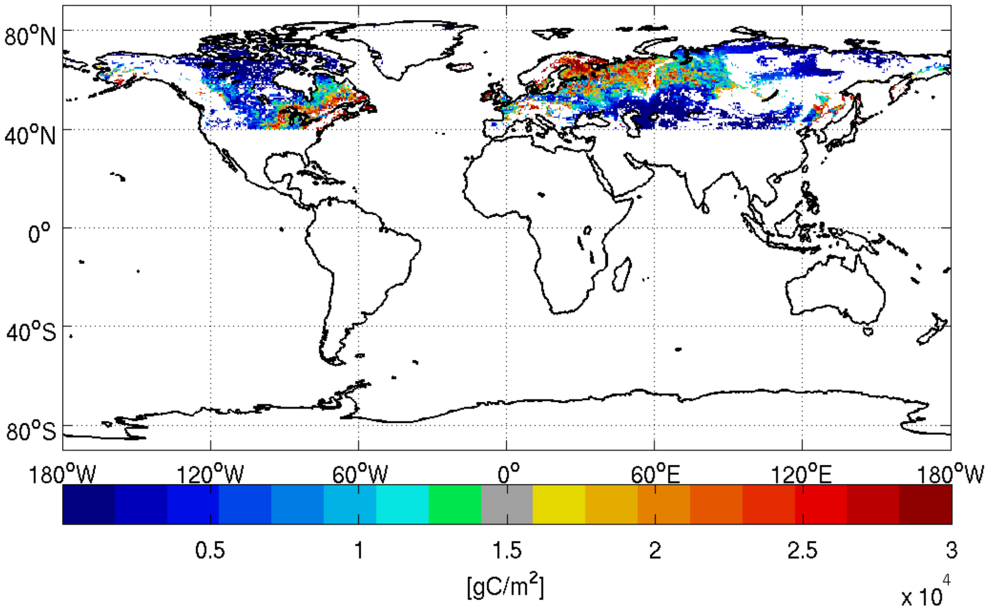

Fig. 6. Peat carbon density, relative to grid cell (*not* wetland fraction), accumulated over the last 8 ka. Results are from experiment core.

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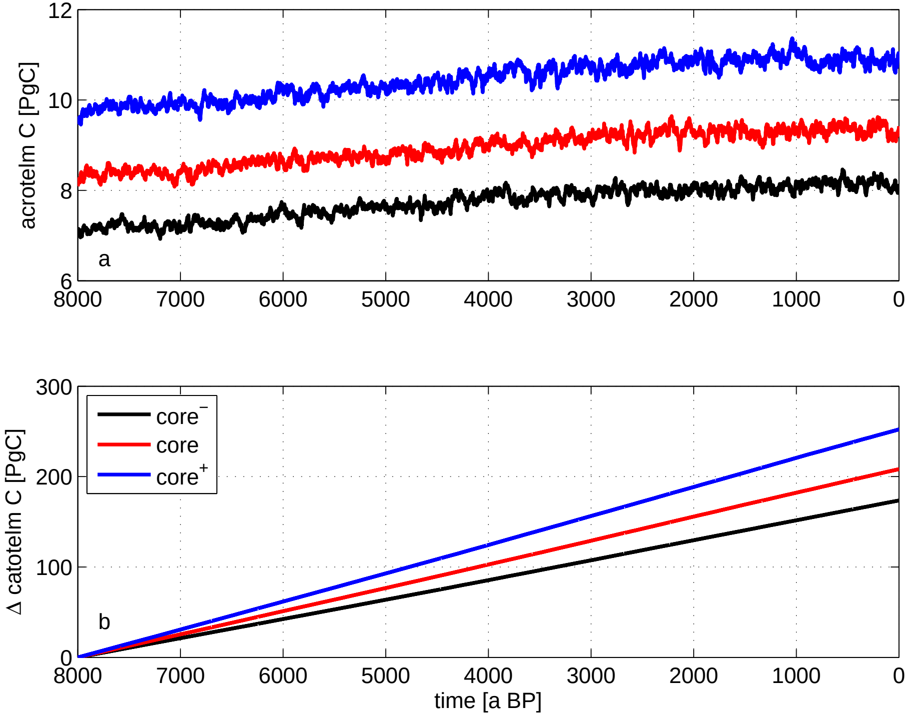

Fig. 7. Total carbon accumulation in wetlands. Upper panel **(a)**: total sum of acrotelm carbon. Lower panel **(b)**: *change* in total sum of catotelm carbon, relative to 8 ka BP.

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