Modelling the effect of water-table management on CO$_2$ and CH$_4$ fluxes from peat soils

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Abstract

Drainage of peatlands for agriculture causes an increase of CO$_2$ flux from peat decomposition, contributing to national CO$_2$ emission. The reverse process, i.e. for re-creation of wetlands, reduces the CO$_2$ flux, but increases the CH$_4$ flux. We developed a process model (PEATLAND) to simulate these fluxes from peat soils subject to different water-table management scenarios. The model combines primary production, aerobic decomposition of soil organic matter (including the soil-parent material, peat), CH$_4$ formation, oxidation, and transport. Model input requires specification of water table and air temperature data sets, vegetation parameters such as primary production and parameters related to gas transport, and basic soil physical data.

Validation using closed flux-chamber measurements of CO$_2$ and CH$_4$ from five different sites in the western Netherlands shows that seasonal changes in fluxes of CO$_2$ and CH$_4$ are correctly modelled. However, the CO$_2$ submodel underestimates peat decomposition when peat decomposition rates obtained from laboratory incubation experiments are used as input. Field decomposition rates are considerably higher. This is attributed to enhancement of decomposition by the addition of easily decomposable material from root exudation ('priming effect'). Model experiments indicate that 1) drainage increases the CO$_2$ production from peat decomposition strongly; 2) restoring a high water table may decrease the total greenhouse gas flux by a small amount although the CH$_4$ flux increases strongly; 3) a warmer climate may cause higher greenhouse gas fluxes from peat soils resulting in a positive feedback to climate warming, and 4) high vegetation productivity in fen meadows may stimulate peat decomposition by the priming effect.

*Keywords:* peat decomposition, drainage, methane, carbon cycle

Introduction

Peatland water table management has a strong effect on the carbon balance of these areas. Drainage of peatlands turns peat soils from a carbon sink into a carbon source since it enhances peat decomposition by increased soil aeration. On the other hand, wetlands are a source of CH$_4$, of which the global warming potential (GWP) is 62 times that of CO$_2$ on a time horizon of 20 years. Wetland contribution to the annual CH$_4$ flux to the atmosphere is approximately 20% (Houghton et al., 1995, 2001). Accurate knowledge of the carbon budget of peatlands has become imperative in policy decisions with respect to greenhouse gas emissions and global climate change.

In the coastal provinces in the Netherlands (Fig. 1) extensive peat swamps have been drained since medieval times for agriculture. Estimates of present soil subsidence in drained peatlands in the western Netherlands range from 0.2 to 5.1 cm year^{-1}, with common values between 0.5 - 1.5 cm year^{-1} (summary by Baas, 2001). Peat oxidation contributes for 1 - 3% of the annual greenhouse gas (GHG) emissions of the Netherlands (Van den Bos, 2003). However, plans have been developed to convert parts of the drained Dutch peatlands back into wetland nature areas by raising water levels (LNV, 2000).

Modelling is an instrument to assess and predict the effects of water management and climate change on the carbon balance of peatlands. The scope of most existing models of soil

Fig. 1.
Coastal peatlands in the western part of the Netherlands with location of the study sites, Amsterdam, and the De Bilt weather station.
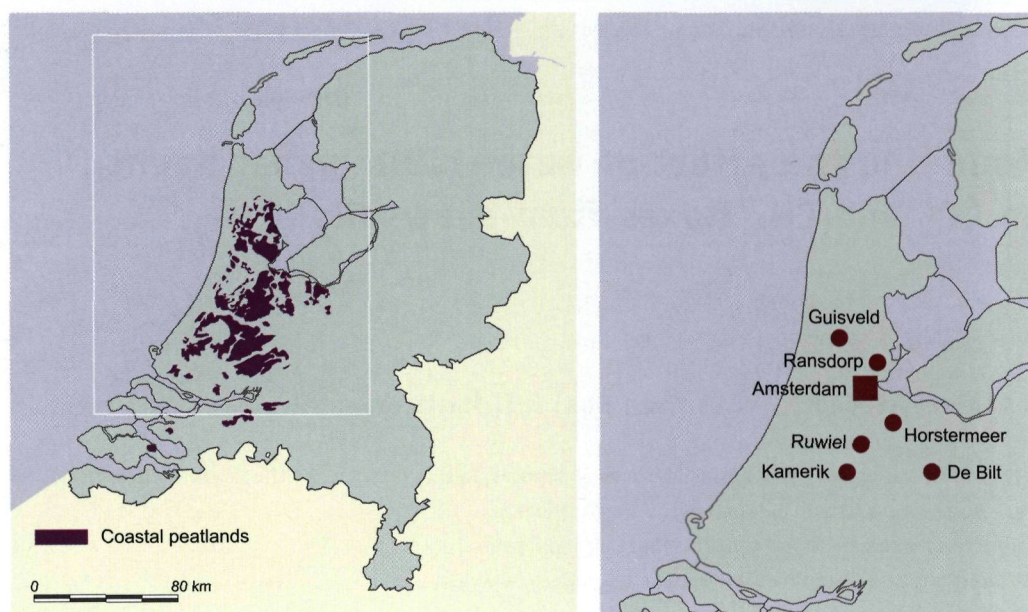
organic matter decomposition is primarily agricultural and they do not include peat substrate decomposition (Paustian et al., 1996, Smith et al., 1997). A model that has been applied successfully in the Netherlands to model aerobic organic matter decomposition is ANIMO (Groenendijk and Kroes, 1997). However, this model does not include CH_4. Models capable of simulating production of CH_4 and its transport processes also have been published recently (Arah & Stephen, 1998; Walter et al., 1996; Walter, 2000; Granberg et al., 2001, Segers & Leffelaar, 2001, Zhang et al., 2002).

A suitable model to evaluate CO_2 and CH_4 fluxes from peatlands is capable of treating the soil matrix (peat) as a reservoir of organic matter, integrates CO_2 and CH_4 fluxes within one model, and has modest parameter demands for upscaling purposes. The model presented here (PEATLAND) consists of a CO_2 submodel for aerobic organic matter decomposition (including the peat substrate of the soil) based on the approach described by Jenkinson and Rayner (1977) and Groenendijk and Kroes (1997). The CH_4 submodel is an adapted version of the CH_4-flux model developed by Walter (2000). Both submodels are linked to a submodel that simulates primary production of soil organic matter as a function of soil temperature. The model has been validated using field measurements, obtained with flux chambers, from five study areas in the western Netherlands (Fig. 1). Recently, it has been applied in a study of methane fluxes from northern wetlands during the Last Glacial (Van Huissteden, 2004).

Study areas and methods

The five study areas (Guisveld, Ransdorp, Kamerik, Horstermeer and Ruwiel) were selected to include a range of soil properties (different types of peat soils), land use (ranging from intensive cattle grazing to nature reserve) and vegetation (*Lolium perenne*

grassland to species-rich *Carex*-dominated vegetation). For the first three sites, an extensive description is given in Van den Bos and Van de Plassche (2003a, b). Flux measurements in the first three sites were made from November 1998 until August 2000, in the Horstermeer site from May 2003, and in the Ruwiel site from January 2004 until present. From the latter two sites shorter, but more detailed flux time series are available, next to hourly observations of water table and soil temperature.

Guisveld (52°28' N, 4°47' E) is a nature-reserve, managed by the National Forest Service (Staatsbosbeheer). Vegetation is mainly reed and grassland, with sedges and mosses. The grasslands are not or sparsely fertilized, and grazed or mown once a year. The approximately 4 m thick peat layer consists in its upper part of oligotrophic bog peat, nowadays in contact with eutrophic and slightly brackish groundwater. The area contains five measurement stations. Two measurement sites are located some 20 cm higher in elevation than the other three sites, resulting in different water tables. For model validation, the high and low water-table sites have been considered separately, called 'Guisveld high WT' and 'Guisveld low WT' respectively.

Ransdorp (52°24' N, 5°00' E) is located in the Waterland region and is subject to regular agricultural use (cattle grazing / hay production). As in Guisveld, the peat consists of approximately 4 m of eutrophied oligotrophic bog peat. It is covered by a surface layer (0 - 30 cm) of crumbly sandy clay with a high organic matter content. The grassland is dominated by Ryegrass (*Lolium perenne*). The water table shows a gradient from the first to the last measurement station with water-table depths from about 30 cm below surface to depths of ca. 70 cm.

Kamerik (52°09' N, 4°52' E) is situated near Utrecht. This area is underlain by thick (max. 6 m) layers of wood and sedge

peat, overlain by fluvial clay. The studied field is a hay pasture with different grass species, grazed during a small part of the year. The top soil (0 - 30 cm) is of anthropogenic origin and consists of a sandy clay layer mixed with organic material and some debris (bricks). The field is mown once or twice a year, not or moderately fertilized, and has a water table between 3 and 87 cm below surface.

Horstermeer (52°14'30"N, 5°5'E) is located SE of Amsterdam, in a drained natural lake. The water level in the ditches is at approximately 3.5 m below sea level, and up to 2 m below that of surrounding polder areas, and is subject to strong seepage. The soil consists of 2 m of clayey gyttja erosively overlying eutrophic fen peat on Pleistocene sand. Until 1997 the area was a grazed pasture, thereafter the water level has been raised to 0.2 - 0.4 m below the surface, to create a nature reserve. The present vegetation, a degraded wet pasture, is not mown or managed otherwise. Dominant species are *Holcus lanatus* mixed with *Equisetum palustre*, and patches dominated by *Urtica dioica*, *Cirsium arvense* or patches with *Glyceria fluitans*. Six measurement stations have been installed, divided over all vegetation types. Four stations with similar soil conditions have been selected for validation.

Armenland Ruwiel (52°10'30"N, 4°56'30"E) is a small nature reserve with a high water table. It is a species-rich, nutrient-poor hay pasture, dominated by sedges (*Carex sp.*), *Eriophurum angustifolium* and some *Sphagnum sp.*, which has never been fertilized. It is mown only once a year. The water table is kept artificially higher than that of the surrounding agricultural land. The soil is a clayey fen peat. Four measurement stations have been installed in the winter of 2003 - 2004 in the reserve. For model validation, the data of three similar (high water table) stations have been used.

The flux measurements were carried out using closed chambers (non-transparant PVC, 45 × 45 × 12 cm, Fig. 2). The measurement procedure has been described in detail by Van den Bos & Van de Plassche (2003a). At the first three sites, the vegetation within each frame was clipped prior to measurement, to reduce plant respiration. At the Horstermeer and Ruwiel sites vegetation was not clipped since it may disturb CH_4 fluxes, that largely consist of plant transported gas.

At least five gas concentration measurements were done at regular time intervals per chamber per flux measurement. CO_2 was measured with an infra-red gas analyzer (Geotech GA 94). For CH_4 analysis, syringe samples were taken from the chambers and analyzed on a gas chromatograph with flame ionization detector. After May 2004 (Horstermeer and Ruwiel sites) CH_4 analysis was performed in the field using an Innova 1312 photo-acoustic gas analyzer, fitted with a CO_2 (sodalime) and H_2O (silicagel) filter to prevent interference of these gases with the CH_4 analysis. At the Horstermeer and Ruwiel sites
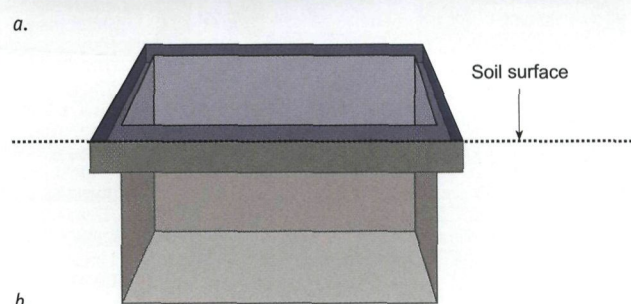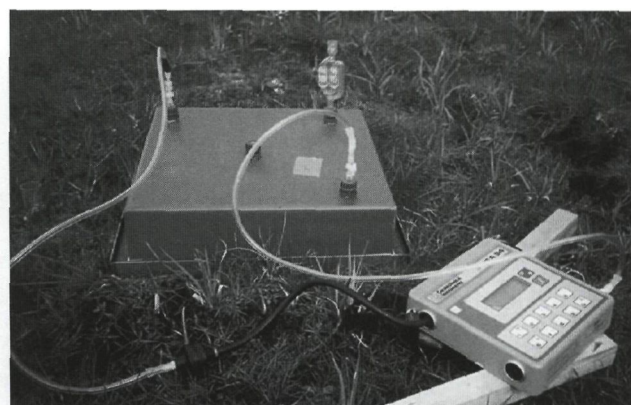

a.

b.

Fig. 2. Setup of the closed flux-chamber measurements. a. Chamber with Geotech GA 94 infrared gas analyzer attached; b. Permanent chamber frame inserted into the soil.

also CH_4 concentration in the soil pore water and air was measured using filters installed at depths of 0.1 - 1.3 m, and soil profiles have been analyzed for dry bulk density, organic matter content and water retention curves.

The PEATLAND model

Introduction to the model.

PEATLAND consists of four submodels, a soil physics submodel to calculate temperature and water saturation of the soil layers, a CO_2 production submodel, a CH_4 submodel and an organic production submodel (Fig. 3). The model represents a column of soil with unit surface, subdivided into fifteen layers of equal thickness (0.1 m). Organic matter content, dry bulk density and water retention curves describe the soil physical conditions of each layer. The model incorporates multiple soil organic matter (SOM) pools, similar to the approach of Jenkinson & Rayner (1977). These SOM pools (Fig. 3) are: Manure added to the soil, with separate pools for liquid and solid parts; roots and litter; root exudates or rhizodeposition (rapidly decomposing excretion products and other organic waste of plant roots); microbial biomass, and resistant humic material. These pools are the same as those implemented in the ANIMO model (Groenendijk and Kroes, 1997). However, the soil parent material (peat) is explicitly included in PEATLAND as a separate pool.
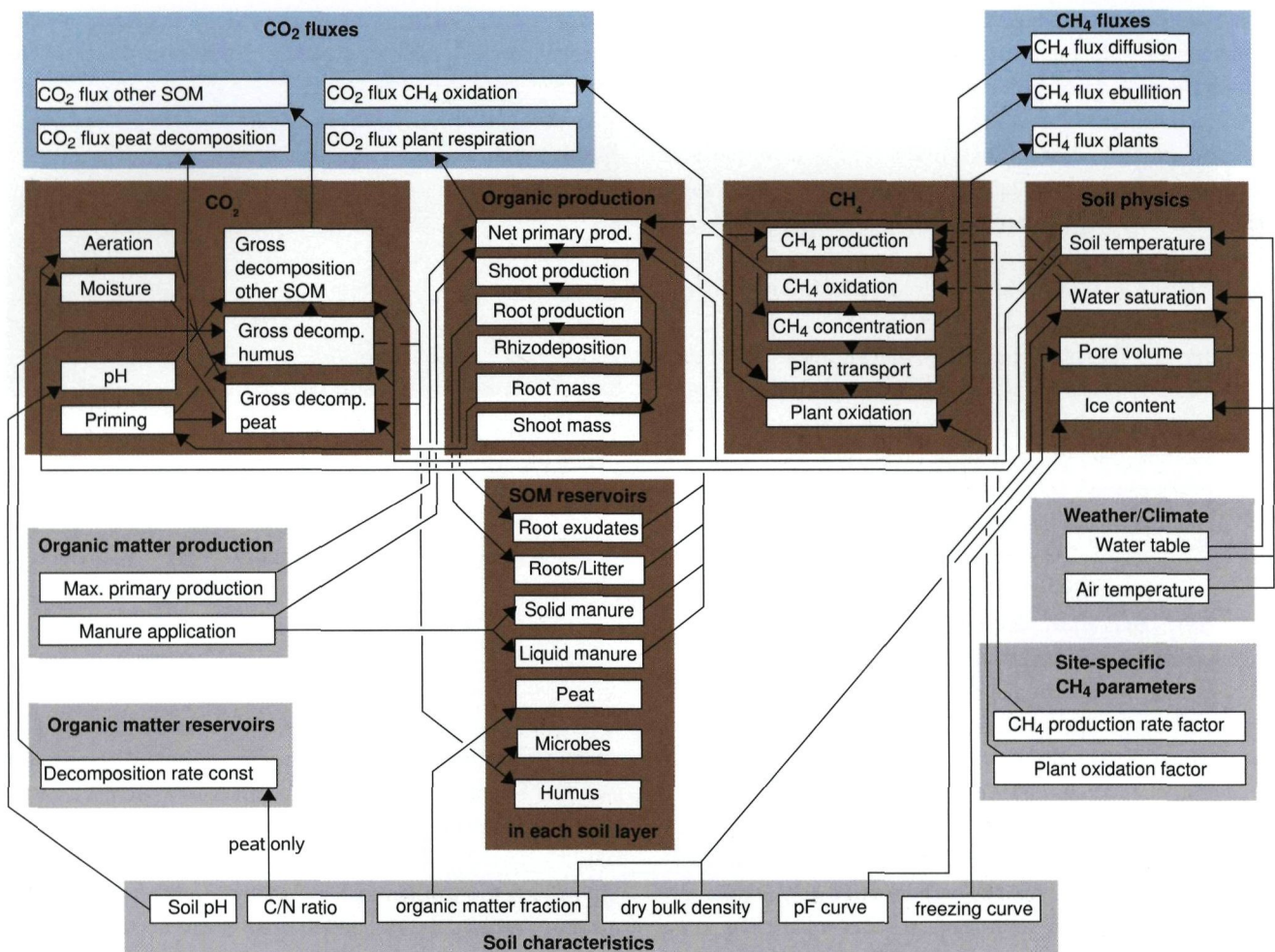
Fig. 3. Model scheme of PEATLAND, showing input, output and state variables and their relations. Grey boxes contain input variables, blue boxes contain output variables.

SOM decomposition is usually modelled using first order rate kinetics (see below). Decomposition rate constants from peat, as determined from soil columns under laboratory conditions are only slightly higher than those of the resistant humic matter pool (Vermeulen & Hendriks, 1996 and references therein; Van den Bos & Van de Plassche, 2003b). However, field measurements (Van den Bos & Van de Plassche, 2003a) suggest 5 to 10 times higher peat decomposition rates, in particular in the densely rooted topsoil. Possible explanations for this discrepancy between field and laboratory peat decomposition rate constants are 1) improved aeration of the topsoil by roots that remove water and transport O_2 into the soil (Drew, 1990), and 2) a priming effect that is caused by the presence of highly decomposable root exudates. The priming effect consists of enhanced decomposition of resistant SOM upon addition of easily decomposable material (Kuzyakov et al., 1999, 2000, 2001); a likely mechanism is co-metabolism of more resistant SOM upon stimulation of microbial growth by easily decomposable compounds. Kuzyakov et al. (2000) indicate an up to 9-fold increase of the decomposition rate due to the priming effect, which indicates that this may strongly enhance

peat decomposition in the highly productive fen meadows in the western Netherlands.

The CH_4 flux from wetland soils depends on CH_4 production in the anaerobic soil zone, its consumption by methanotrophic bacteria in the aerated zone above the water table, and the different transport pathways of CH_4 to the atmosphere (Walter, 2000). Most methane originates from fresh, labile organic matter, which is provided to the subsoil as root exudates. Photosynthesized C is emitted from wetland soils as CH_4-C within a few days (e.g. King & Reeburg, 2002). Compared to fresh organic matter sources, the peat substrate provides only a minor contribution, since it consists of more resistant organic matter which has already undergone a decomposition cycle during its formation. Mineral soils may show CH_4 emission similar to, or higher than peat soils since most of the CH_4 is derived from freshly produced organic matter (e.g. Fiedler & Sommer, 2000; Van der Nat & Middelburg, 2000). In field studies, the CH_4-production capacity in the deeper peat substrate is an order of magnitude lower than that in the root zone (Van den Pol - Van Dasselaar et al., 1999a; Van den Pol - Van Dasselaar & Oenema, 1999). In the model, the

substrate for CH_4 production is the sum of the reservoirs of fresh organic matter described above, a smaller contribution from the peat substrate is optional.

Soil physics

Soil temperature and water content determine microbial reaction rates. The model includes calculation of soil temperature, soil freezing and a simplified approach to soil water content for each soil layer. Alternatively, observation data or data generated by other soil physical models can be imported. The soil surface temperature at each time step is either derived from the mean annual soil surface temperature and its amplitude, assuming a sinusoidal cycle, or given as an input time series. The temperature gradient in the soil is calculated using the heat-flow equation

$$\frac{\partial T}{\partial t} = D_T \frac{\partial^2 T}{\partial z^2} \qquad (1)$$

in which t is time, and z depth (m), D_T (m^{2}d^{-1}) is the thermal diffusivity. In particular in peat soils, D_T varies strongly with water content. The model incorporates a numerical solution of equation (1) with D_T varying with moisture content of each layer, estimated according to Hillel (1998) from the volumetric heat capacity C and thermal conductivity K of the major soil constituents (mineral and organic fraction, water, air).

We apply a simplified approach to soil moisture, assuming that soil moisture is in equilibrium with gravity. This simplification has the disadvantage that changes in soil moisture due to evapotranspiration and precipitation cannot be simulated properly with deeper water tables. It is justified by the generally high water table of the peat soils. Moreover, sophisticated soil-water submodels in SOM decomposition models do not necessarily improve model performance due to parameterization problems as shown by a comparative study by Smith et al., 1997.

Soil freezing has been included, to allow modelling of methane fluxes in permafrost environments (Van Huissteden, 2004). Freezing and thawing influences the thermal diffusivity in the heat equation by introducing an apparent heat capacity (Williams & Smith, 1991). At below-zero temperatures, not all soil water freezes immediately. The relation between soil temperature $T(t,z)$ at time T and depth z and unfrozen water content $W(t,z)$ (kg water / kg dry soil) is modelled by assuming a hyperbolic relation (denoted in Fig. 3 as 'freezing curve'):

$$W(t,z) = W_{min} + \frac{1}{(s-T(t,z))^p} \qquad (2)$$

W_{min} is the unfrozen water content at infinitely low temperatures, assumed to be equal to the water content at wilting point (pF = 4.2). Parameter s is a scaling parameter that depends on the maximum water content W_0 (saturated soil at 0° C) and W_{min}:

$$s = \exp(-1/p \, \log(W_0 - W_{min})) \qquad (3)$$

The parameter p is a constant depending on soil composition. Lower values of p (~1.5) result in a less steep decrease of the unfrozen water content, as occurs in clays and peat, higher values (~2) result in a steeper decrease as occurring in sands.

CO_2 submodel

Similar to the model of Jenkinson & Rayner (1977), it is assumed that the decomposition reaction for each SOM pool listed above results in reaction products that are partitioned between CO_2, microbial biomass and the humus pool (Fig. 3). For decomposition first order rate kinetics is assumed:

$$\frac{dQ}{dt} = kQ \qquad (4)$$

where Q is the mass of organic C in a specific SOM pool per unit volume of soil (kg m^{-3}) and k the decomposition rate constant (d^{-1}). Each SOM pool is assigned its own decomposition rate constant k.

In the Jenkinson-Rayner approach, equation (4) describes of decomposition of SOM in general; dQ/dt includes both organic matter transferred into other SOM pools and into CO_2. The amount of carbon partitioned to the microbial biomass pool is calculated from the total decomposed amount dQ/dt using a transfer coefficient a_{micr}, which is the microbial assimilation rate. Similarly a transfer coefficient a_{humus} is used to calculate the transfer to the resistant humus pool. Thus, from the total amount of decomposed organic matter dQ/dt, a fraction a_{micr} is transferred to the microbial biomass, a fraction a_{humus} to the humus pool, and a fraction $1 - a_{micr} - a_{humus}$ is transferred into CO_2. The microbial biomass itself is subject to decomposition after death also. The decomposition rate of the microbial biomass is a function of the microbial death rate and decomposition of its dead organic matter.

The decomposition rate of each SOM reservoir is adapted by environmental correction factors that influence microbial activity, after Groenendijk & Kroes (1997). Starting from a k value specified for optimal conditions, the actual k_{env} is calculated by reducing or increasing k by a factor f for a specific environmental parameter: f_T for temperature, f_m for soil moisture, f_{pH} for soil pH, f_{ae} for soil aeration and f_{prim} for priming effects according to:

$$k_{env} = f_{ae} \cdot f_m \cdot f_T \cdot f_{pH} \cdot f_{prim} \cdot k \qquad (5)$$

With exception of f_T and f_{prim} (see below) all correction factors range from 0 to 1. The factors f_m and f_{ae} are complementary. A near-saturated soil moisture condition leads to a reduction of decomposition due to anaerobic conditions, but a low moisture supply also retards decomposition. Optimum conditions are generally found at pF 2.2 - 2.7 (near field capacity). The decomposition rate declines steeply at high water content and more gradually at low water content (Paul & Clark, 1996).

For the effect of soil dryness, we assumed a linear decrease of f_m from 1.0 to 0.2 between pF 2.7 and pF 4.2 (wilting point). In practice, the latter is never reached in the peat soils studied here. For the effect of soil moisture on aeration we apply a linear relation of the aeration factor on soil water content, which lets f_{ae} decline from 1 to 0 between a pore-water saturation from 80% to 100%. Since roots decrease the water content of the soil layer and also contribute to aeration by O_2 transport (Drew, 1990), a root mass dependent correction of this relation to a steeper slope is optional in the model.

Microbial activity depends on soil temperature according to the Arrhenius equation (e.g. Paul & Clark, 1996):

$$f_T = \exp\left[\frac{E_a}{R}\left(\frac{1}{T_{ref}} - \frac{1}{T}\right)\right] \tag{6}$$

where T is the soil temperature (°K), T_{ref} a reference temperature (°K) at which $f_T = 1$, E_a the molecular activation energy of the reaction (J mol^{-1}), and R the gas constant (J mol^{-1} K^{-1}).

Soil pH values also influence decomposition; different groups of organisms are active at different pH's. Most of the known bacterial species grow within a pH range of 4 to 9, and fungi at pH 4 - 6. Decomposition tends to be slower at low pH (Paul and Clark, 1996). The peat soils studied here generally have moderate pH values (5 - 7) at the topsoil, decreasing with depth (Vermeulen & Hendriks, 1996). Groenendijk & Kroes (1997) adopted an empirical relation in their ANIMO model, which is also used in our model:

$$f_{pH} = \frac{1}{1+e^{-2.5(pH-5)}} \tag{7}$$

The priming effect discussed above is incorporated into the model by a correction factor f_{prim} (see organic submodel, eq. 17) that acts on resistant SOM reservoirs only (peat, humus), and depends on the production of root exudates simulated by the organic production submodel.

Nutrient status, nitrogen content in particular, is another important factor influencing the decomposition rate. This is not incorporated as an environmental correction factor on k, but expressed by the different values of k of the various SOM pools. For peat, an empirical relation between k and C/N ratio is found by Vermeulen & Hendriks (1996):

$$k_{peat} = 0.016 - 0.00021\,(C/N) \tag{8}$$

We use this relation to estimate k for each peat layer whenever C/N ratios are available, otherwise k_{peat} is set at 0.02 d^{-1} as has been determined from soil columns under laboratory conditions (Vermeulen & Hendriks, 1996).

CH$_4$ submodel

The CH$_4$ submodel is based on Walter et al. (1996), Walter (2000) and Bogner et al. (2000). A condensed description of the model of Walter (2000) and our implementation is given below. The model of Walter includes 1) CH$_4$ production depending on substrate availability; 2) CH$_4$ oxidation within the aerated topsoil and in plant roots and stems; 3) CH$_4$ transport by diffusion above and below the water table; 4) transport by bubble formation (ebullition) below the water table; and 5) transport through plants:

$$\frac{\partial}{\partial t}D_{CH_4} = -\frac{\partial}{\partial z}F_{diff}(t,z)+Q_{eb}(t,z)+Q_{pl}(t,z)+R_{pr}(t,z)+R_{ox}(t,z) \tag{9}$$

where $C_{CH_4}(t,z)$ is the CH$_4$ concentration at time t and depth z, F_{diff} is the diffusive flux, Q_{eb} and Q_{pl} represent ebullition and plant transport, R_{pr} and R_{ox} are the CH$_4$ sources and sinks due to CH$_4$ production and oxidation. Boundary conditions for solving (9) are the atmospheric CH$_4$ concentration at the top of the soil column, at the base $\partial C/\partial z = 0$.

CH$_4$ production is linearly related to substrate availability, which in turn depends on organic production and root distribution (Walter, 2000). In our model, the substrate is the sum of the same labile SOM reservoirs used in the CO$_2$ submodel (plant roots, root exudates, manure). These reservoirs are the principal link between the CO$_2$ and CH$_4$ submodels. The CH$_4$ production is temperature dependent based on a Q$_{10}$ value:

$$R_{pr}(t,z) = R_0 \cdot C_{fresh} \cdot Q_{10}^{\left(\frac{T(t,z)-T_{ref}}{10}\right)} \tag{10}$$

in which R_0 is a constant rate factor (µMh^{-1}), C_{fresh} is the C concentration in the fresh SOM reservoirs, T the soil temperature at depth z (m) and time t (hr), and T_{ref} a reference temperature, approximately the yearly mean soil temperature below the water table. R_0 is a site-dependent tuning parameter that incorporates site-specific factors such as organic matter quality and environmental factors (Walter, 2000) (set at 0.5 in our simulations, Table 1). An environmental factor which strongly influences CH$_4$ production is soil pH (Dunfield et al., 1993; Bergman et al., 1998; Segers, 1998). Based on the data of Dunfield et al. (1993) R_0 is decreased by 0.1 linearly at each pH unit lower than 7.

CH$_4$ is oxidized by methanotrophic bacteria in the aerated topsoil and in plant roots and stems. Oxidation within the soil pores is modelled using the Michaelis-Menten equation:

$$R_{ox} = -\frac{V_{max}\,C_{CH_4}(t,z)}{K_m+C_{CH_4}(t,z)}\,Q_{10,ox}^{\left(\frac{T(t,z)-T_{ref}}{10}\right)} \tag{11}$$

where K_m (µM) and V_{max} (µM hr^{-1}) are the Michaelis-Menten constants. $Q_{10,ox}$ determines the temperature sensitivity of the process.

Plant transport of CH$_4$ is modelled by the transport rate Q_{plant}:

$$Q_{pl}(t,z) = -c_p \cdot V \cdot f_{root}(z) \cdot f_{grow}(t) \cdot C_{CH_4}(t,z) \tag{12}$$

and surface flux F_{pl}

$$F_{pl}(t,z) = \int_{nroots}^{0} \left[Q_{pl}(t,z)(1-P_{ox}) \right] dz \qquad (13)$$

Q_{pl} depends on rate constant c_p, a vegetation factor V, the root distribution f_{root} and a function describing the growth rate of the vegetation, f_{grow}, proportional to the primary production. Since grasses are good gas transporters, we used the maximum value for V (Table 1). The transport rate is integrated over the depth of the root zone to obtain the surface flux. P_{ox} is the fraction of CH_4 that is oxidized during plant transport.

Ebullition occurs when CH_4 concentration of the soil water rises above a threshold concentration; its rate Q_{eb} is described by:

$$Q_{eb} = -c_e \cdot f(CH_4) \cdot (C_{CH_4}(t,z) - C_{thresh}) \qquad (14)$$

in which c_e is a rate constant, C_{thresh} the threshold concentration, and $f(CH_4)$ a step function that is 1 if the threshold concentration is exceeded and 0 otherwise. The flux rate is integrated over all layers to obtain the flux. If the water table is below the surface (which is always the case in our model simulations), the ebullition flux is added to the CH_4 concentration in the unsaturated zone.

Primary production and soil organic matter production submodel

The SOM production submodel simulates the addition of fresh organic material to the root zone and the addition of manure. We adopted a simple model in which the above and below-ground primary productivity depends on the temperature of

Table 1. Model parameters. References in third column: 1: Jenkinson & Rayner (1977); 2: Groenendijk & Kroes (1997); 3: Kuzyakov et al. (2000); 4: Hillel (1998) ; 5: Van den Bos & Van de Plassche (2003b); 6: Walter et al. (1996), Walter (2000) ; 7: Van den Bos & Van de Plassche (2003a); 8: Kuikman (1996); 9: Kuzyakov et al. (1999, 2001), Whipps (1990); 10: Williams & Smith (1991); 11: this paper.

Symbol	Dimension	Description	Ref.	Value and range
Soil physics				
C	$J \cdot m^{-3} \cdot {}^\circ C$	volumetric heat capacity	4, 10	see text
K	$W \cdot m^{-1} \cdot {}^\circ C$	thermal conductivity	4, 10	see text
p		Soil texture dependent power constant unfrozen water content relation	10, 11	1.5 (clay, peat) - 2 (sand)
CO_2 submodel				
a_{humus}		transfer coefficient to the humus (resistant SOM) pool	1, 2	0.1
a_{micr}		transfer coefficient to the microbial biomass	1, 2	0.27
k	d^{-1}	rate constant organic matter decomposition		see text
c_{prim}		priming correction constant	3	10 (0 - 10)
T_{ref}	$^\circ K$	reference temperature for temperature dependent decomposition	11	284
E_a	$J \cdot mol^{-1}$	molecular activation energy decomposition reaction	5	96000 - 126000
CH_4 submodel				
R_0	$\mu M h^{-1}$	rate factor methane production	6, 11	0.5 (0.3 - 0.6)
$Q_{10,p}$		Q_{10} factor methane production	6, 7	7.5 (1.7 - 16)
$Q_{10,ox}$		Q_{10} factor for methane oxidation	6	1.4 (1.4 - 2.1)
V_{max}	$\mu M h^{-1}$	maximum reaction rate Michaelis-Menten equation	6	50 (5 - 50)
K_m	μM	half saturation constant Michaelis-Menten equation	6	1.0 (1 - 5)
V		vegetation-dependent rate factor plant transport methane	6	15 (0 - 15)
c_p	h^{-1}	rate constant plant transport methane	6	0.01
P_{ox}		fraction of methane oxidized during plant transport	6, 11	0.5 - 0.9 (0.4 - 0.9)
c_e	h^{-1}	rate constant ebullition transport methane	6	1
C_{tresh}	μM	treshold concentration ebullition	6, 11	500 - 5000
SOM production submodel				
P_0	$kg\ C\ m^{-2}\ d^{-1}$	maximum primary production	8	0.0057
f_{man}		manure correction factor production	5, 8	0.6 - 1
f_{ox}		oxygen availability correction factor production	10	0.0 (0 - 1)
f_P		part of primary production allocated to shoots	8, 9	0.6
f_{dep}		part of root production allocated to rhizodeposition	8, 9	0.4 (0.1 - 0.4)
f_{spring}		correction factor for higher rhizodeposition in spring	11	0.5 (0 - 1)
$f_{sen,r}$		root senescence factor	11	0.0025
$f_{harvest}$		fraction harvested from biomass	5, 11	0.5
$f_{sen,s}$		shoot senescence factor	8, 11	0.01

the upper 10 cm model soil layer. If the top layer is nearly saturated with water a reduction factor is applied for oxygen availability. A second reduction factor is applied for application of manure:

$$P = K(T_{surf}) \cdot f_{ox} \cdot f_{man} P_0 \qquad (15)$$

in which P_0 is the maximum possible above- and below ground primary production (kg C m^{-2} d^{-1}), P is the actual primary production, $K_T(T_{surf})$, f_{ox} and f_{man} are the correction factors for temperature, oxygen availability and manure application. The function $K_T(T_{surf})$ ranges between 0 and 1. The temperature dependence of the production is linear between a minimum temperature of 5° C and an optimum of 15° C; below the minimum temperature the production equals the minimum production, which is set to zero in all model runs. The relation between T_{surf} and K_T is sigmoid, based on a sinusoidal function. Values for P_0 have been derived from ^{14}C pulse labeling experiments on carbon fluxes in Dutch grasslands (Kuikman, 1996). The value of f_{man} is based on the same study, and amounts 0.6 if no manure is applied and 1.0 at optimal supply. The oxygen availability factor f_{ox} ranges between 1 and 0. It depends linearly on the water saturation of the top layer and is applied when a saturation threshold value of 0.9 is exceeded.

The primary production P is partitioned using partitioning factors f_P and fdep between newly produced shoots (P_s) and subsoil production (P_r) and rhizodeposition (P_{dep}):

$$Ps = f_p \cdot P$$
$$Pr = P - P_s \qquad (16)$$
$$P_{dep} = f_{dep} \cdot f_{spring} \cdot P_r$$

Values of f_P and f_{dep} for grasslands have been based on Kuikman (1996), Kuzyakov et al. (1999, 2001), and on comparable values derived for wheat summarized by Whipps (1990) (Table 1). Kuzyakov et al. (2001) found that during the entire growth period of the grasses up to 50% of the carbon transferred to the soil may consist of easily decomposable substances. Exudation rate is higher for younger roots than for older roots (Whipps, 1990), and therefore exudation tends to be higher in spring. This effect is optionally incorporated in the model by a time dependent spring factor f_{spring} (0 < f_{spring} < 1) which enhances rhizodeposition during the early part of the growing season and depresses it in the later part.

The roots are partitioned over the soil layers using an exponential root distribution function $f_{root}(z)$ and added to the root mass $M_{roots}(t,z)$. During each time step, part of the root mass in each layer is subject to senescence based on a senescence factor $f_{sen,r}$ and added to the roots and litter reservoir. Shoot production is added to the above-ground biomass M_{shoots}. A fraction $f_{harvest}$ of the shoots is harvested on specified days during the simulation period, and another fraction $f_{sen,s}$ dies off each day and contributes to the litter reservoir of the

upper layer. $f_{sen,s}$ and $f_{harvest}$ can be specified for each day of the year to allow for grazing and enhanced shoot senescence after harvesting. Root growth can be restricted in the model to layers above the groundwater table in case of vegetations (e.g. agricultural crops) that lack telmatophytic species. Addition of manure can be specified for particular days of the year. The manure contributes to the manure reservoir in the upper layer. Transport of manure to layers below the first layer is not included.

The root mass dependent priming correction (see previous) is calculated as a function of the root mass $M_{roots}(t,z)$:

$$f_{prim}(t,z) = 1 + c_{prim} \cdot f_{spring} \cdot f_{grow} \frac{M_{roots}(t,z)}{\sum_{i=1}^{n} M_{roots}(t,i)} \qquad (17)$$

in which c_{prim} is a user-defined constant (value 0-10), f_{spring} and f_{grow} are the spring correction and growth factor defined above in the organic production and CH_4 submodels, and M_{roots} is the root mass.

An estimate of total plant respiration (including root respiration) is included in the model, using the relation

$$R = r_1 P + r_2 B \qquad (18)$$

In this relation respiration is split into two components $r_1 P$, growth respiration depending on biomass production P, and $r_2 B$, the maintenance respiration, depending on biomass B (Thornley and Johnson, 1990).

Model parameterization and data requirements

Most of the input data for the model can be obtained from generic data, e.g. soil profile descriptions and weather station data. Different levels of detail can be selected in these data, e.g. soil profiles consisting of several horizons or only one. However, part of the data is difficult to obtain, and may need to be based model optimization. Examples are the sizes of organic matter reservoirs.

Basic soil composition and physical data, such as organic matter content, dry bulk density and moisture retention curves of each soil horizon can be derived from on-site measurements for a particular soil profile, or from standardized soil types in the Netherlands (Wösten et al., 1994). For instance in the validation below, the moisture retention curves for the Guisveld, Kamerik and Ransdorp sites have been derived from Wösten et al. (1994), while for the Horstermeer and Ruwiel sites, the water retention curves have been determined from field samples. Meteorological input is represented by time series of water table and air temperature, or their average yearly minima and maxima, depending on the amount of temporal detail that is required in the model simulation.

The initial C content of several organic matter reservoirs is difficult to quantify, in particular the amount of dead roots,

litter, and rhizodeposition (Whipps, 1990). A check on the initial conditions can be performed by assuming steady state conditions. A large increase or decrease in the order of 25% or more of the amount of C in a reservoir should not occur. Addition of manure has been parametrized according to field observation on application dates and average values derived from agricultural statistics (typically in the order of 0.1 kg $C \cdot m^{-2}$).

Aerobic decomposition constants (k in equation 1) for peat, roots, litter, and exudates have been derived from the literature. Vermeulen & Hendriks (1996) report decomposition constants for different types of peat based on incubation experiments of soil columns from areas comparable to our study sites. Values of k for roots (between 4 and 5 d^{-1}) have been listed by Jenkinson & Rainer (1977), Martin (1989) and Kuikman (1996). In particular for rhizodeposition the decomposition constants vary widely (Drew, 1990). In line with Kuzyakov et al. (1999) we assume that aerobic decomposition of these substances is extremely rapid (with k values of 0.07 hr^{-1}); which means that during one model time step (1 - 10 days) the rhizodeposition input is consumed almost entirely.

In the CH_4 submodel, the CH_4 production rate R_0 (eq. 8) is a tuning parameter (Walter, 2000). The validation model runs below indicate that this value has to be set at the low end of the range indicated by Walter (2000) for our sites (0.5 μMh^{-1}). The Q_{10} value for CH_4 production has been inferred from Van den Bos & Van de Plassche (2003a, b). It was assumed that conditions for CH_4 oxidation in the generally well aerated top soils of the grasslands are optimal; the parameters of equation (11) have been set to reflect this condition. Since grasses and sedges are efficient gas transporters (Busch & Lösch, 1999; Walter, 2000), the plant transport rate factor V also has been set at a maximum value.

Model validation

The model is validated with measured CO_2 and CH_4 fluxes of the sites described above. Locally measured time series of water table and air temperature have been supplemented with air temperature data from the De Bilt weather station (Fig. 1). The temperatures for De Bilt are slightly lower than the temperatures measured at the Guisveld and Ransdorp field sites; the De Bilt temperatures have been adapted by adding 1° C. Each model run is preceded by an artificial one year cycle to reduce the influence of initial conditions of the vertical temperature profile, CH_4 concentration profile, and organic matter reservoirs. For the Guisveld, Ransdorp and Kamerik sites a model time step of 10 days is used, the data availability for the Horstermeer and Ruwiel sites allowed a 1 day time step.

The closed chamber-flux measurements for CO_2 cannot be compared directly with the model output since the measurements include CO_2 from plant respiration. The procedure adopted to solve this problem differs among the sites. For the Guisveld, Kamerik and Ransdorp sites, a correction has been applied, based on field experiments on low water table sites by Van den Bos & Van de Plassche (2003a). These indicate that the CO_2 flux from the soil is on average 80% of the total measured flux when vegetation respiration is excluded. The contribution by peat decomposition is approximately 60% of the flux. The largest part of the peat decomposition flux is derived from the shallow root zone (Van den Bos & Van de Plassche, 2003a). We therefore reduce the measured CO_2 flux data by 20% before comparison with the model output. To assess the model performance for peat degradation, we also compare the CO_2 flux from the peat reservoir with 60% of the data, which provides a second validation criterion for the CO_2 flux. For the Horstermeer and Ruwiel sites, where shoot respiration is included in the measurements, an estimate of the respiration has been made using equation 18, and the total of soil and shoot respiration has been compared with the measured data.

The magnitude and pattern of the total CO_2 fluxes generally agree with the data (Fig. 4). Since the frequency of the (approximately monthly) measurements is low compared to the model time step of one to ten days, the exact pattern of flux peaks and lows of data and model cannot be compared, but the general yearly cycle is produced correctly. The simulated summer CO_2 flux at the Guisveld high WT stations is slightly too low for the first summer. For the Guisveld low WT, Kamerik and Ransdorp sites the model shows higher summer peaks than the data. Possible sources of error are 1) overestimation of the primary production by the rather simple SOM production submodel; 2) uncertainties in the addition of manure, or 3) uncertainty in the temperature dependency of the decomposition rates. Also the winter flux tends to be underestimated by the model for the same reason. For the Horstermeer and Ruwiel sites, the simulated flux (including simulated respiration) closely follows the measured pattern in the data. For the high water table sites (Guisveld high WT, Horstermeer, Ruwiel) the modelled peat decomposition is considerably less than the 60% of the measured flux derived for the other sites. However, this 60% is a coarse estimate based on low water table sites (Van den Bos & Van de Plassche, 2003a).

Although the CH_4 submodel computes the CH_4 fluxes from diffusion, ebullition and plant transport separately, we used only the summed total CH_4 flux which was measured with the flux chambers. For all sites the magnitude and general seasonal pattern of the modelled values compare well with the data (Fig. 5). A few deviations occur, for which the probable causes have been discussed below. In the first summer a flux peak is present at Guisveld, Kamerik and Ransdorp, which is not reproduced by the model. This flux peak follows on a heavy rain event after a prolonged dry period. For the Horstermeer site, the modelled summer flux of 2003 is somewhat higher than the observed flux. This is attributed to soil crack formation during the extreme dry summer, promoting a better soil aeration. The start of the summer of 2004 shows lower fluxes than observed, but at that time the variability between

CO$_2$ mg.m^{-2}.hr^{-1}
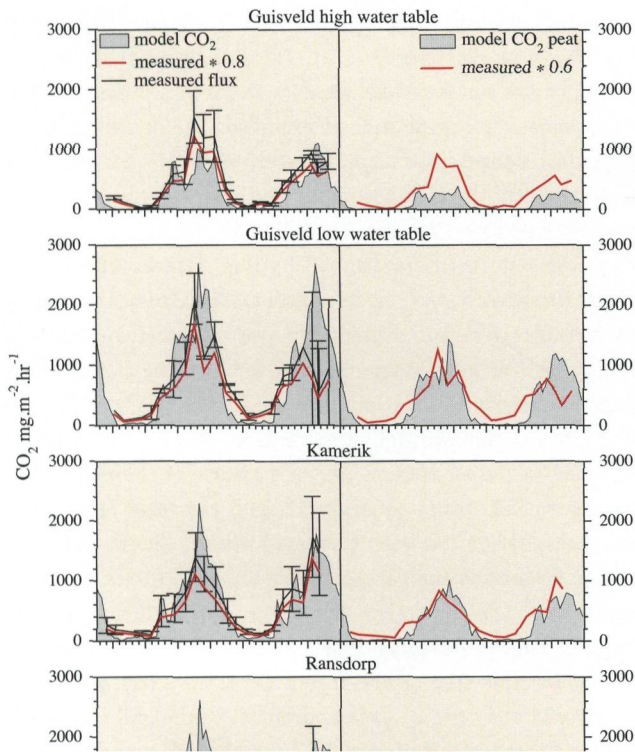
the stations was very large; the data of two of the four stations agree closely with the modelled flux. Small negative fluxes may occur above low water table soils when CH_4 is taken up from the atmosphere by methanotrophic bacteria (Van den Pol - Van Dasselaar et al., 1997; Van Huissteden et al., 2005). These were not measured at the sites, due to insufficient precision of the measurements based on gas sampling. However, the model generated near-zero and slightly negative fluxes for the Ransdorp site.

The parameters for soil CH_4 production and oxidation (eq. 10, 11) were kept constant for all simulations. Plant transport is in all cases the dominant mode of transport, therefore the model is quite sensitive to changes the parameters of plant transport (eq. 12, 13). Slight adjustments have been made for each site to the fraction of CH_4 oxidized during plant transport, P_{ox}. For the high water table Ruwiel and Horstermeer sites lower values for P_{ox} were used (0.5 vs 0.8 - 0.9). An alternative may have been adjustment of the CH_4 production rate factor R_0, as done by Walter (2000). However, it is more likely that considerable variation in plant transport properties exists between the sites, based on differences in plant communities. In particular sedges and arenchymous plants (
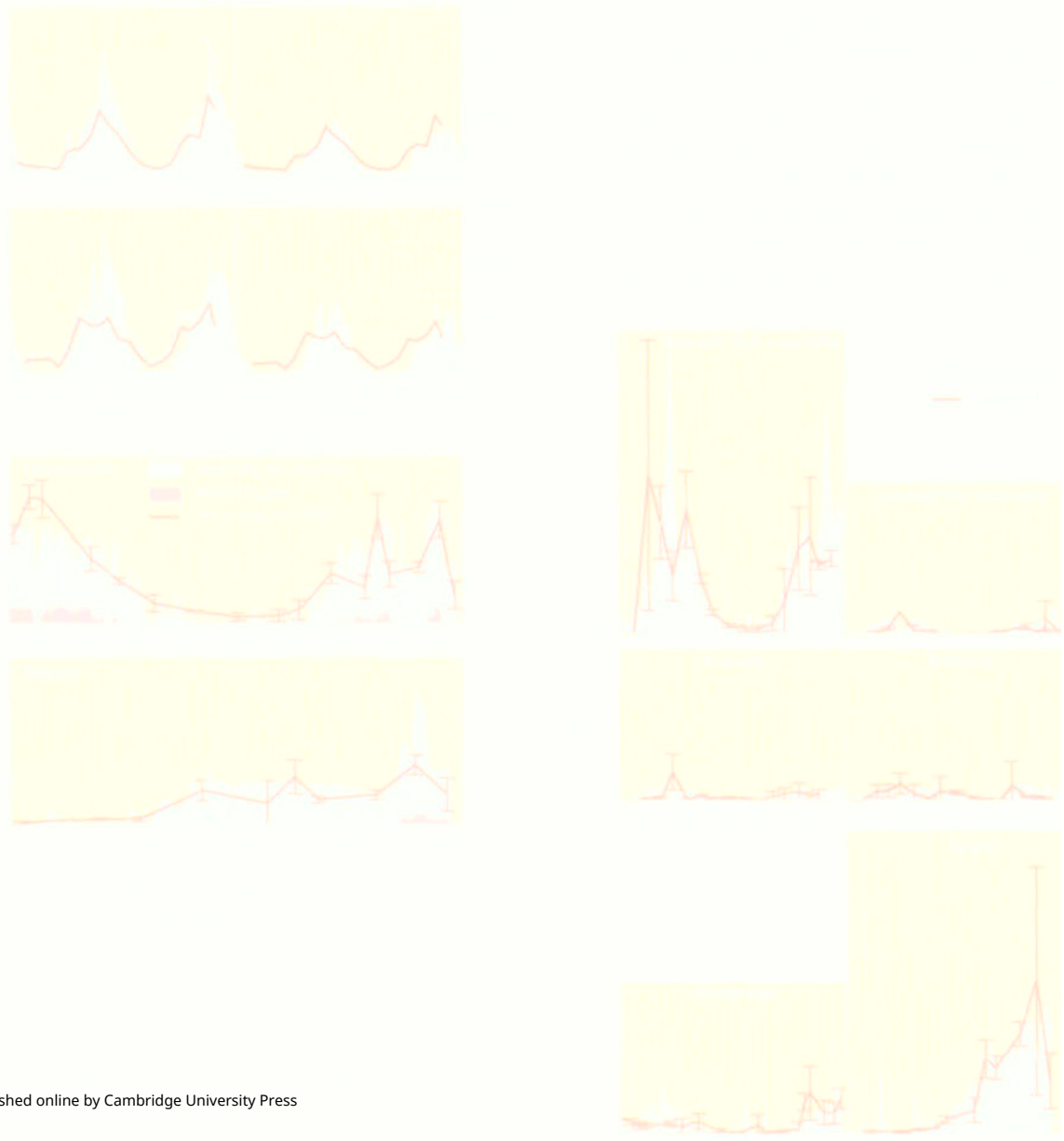
Equisetum, Glyceria) at the Ruwiel and Horstermeer sites may show very high plant transport with low oxidation (Verville et al., 1998; Busch and Lösch, 1999).

Model sensitivity and examples

A number of sensitivity tests have been performed to test the model under a range of operating conditions, and to assess the influence of input parameters which are difficult to quantify. In particular the water table and temperature sensitivity tests can also be used as examples of model model behaviour. To assess the total greenhouse gas effect (henceforth total GHG), the CH_4 fluxes have been multiplied by a factor of 62, which is the GWP in CO_2 equivalents for CH_4 at a time scale of 20 years (Houghton et al., 1995). This time horizon agrees with the approximate time scale for planning of water management. Departure point for all tests is the simulation for the Guisveld high WT site, the effects of the different tests have been calculated for one yearly cycle.

Water table. A drop of the water table with respectively 0.1 and 0.2 m increases the CO_2 flux and decreases the CH_4 flux. Also slight changes occur in the yearly pattern of the fluxes (Fig. 6, Table 2). On the other hand, the CH_4 flux is reduced by a factor of 0.35. The total GHG effect is an increase of the flux by a factor of 2.13. The model also produces large changes in the fluxes when the water table is kept at a constant high level (equal to the yearly average at the Guisveld high WT site) throughout the year. The CH_4 flux increases by a factor of 1.47 resulting from a combination of a high water table with high summer temperatures. However, the net effect is a decrease of the total GHG flux by a factor of 0.81, since the CO_2 flux is considerably lower.
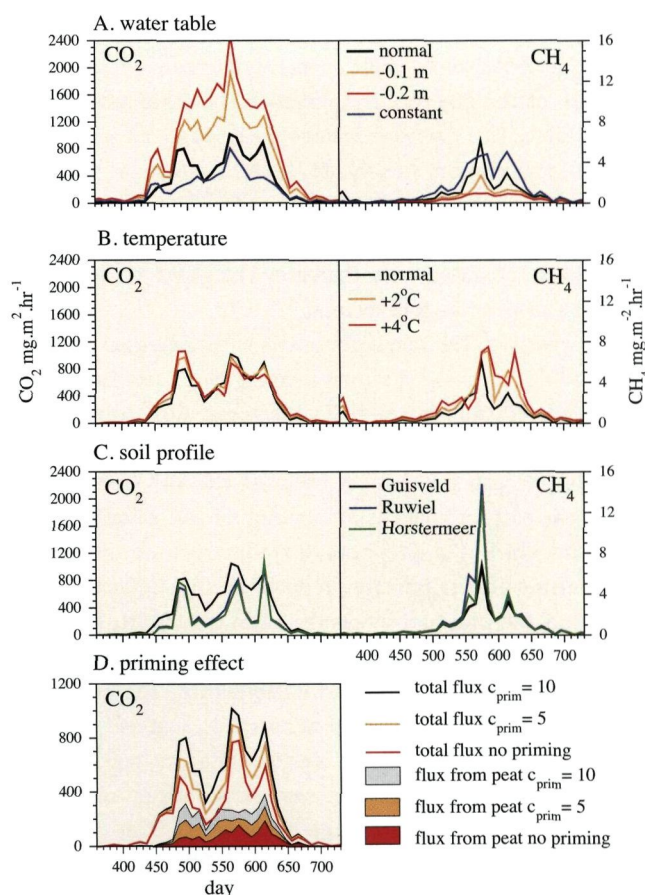

Fig. 6. *Sensitivity test results. Modelled flux time series over one year are shown. The sensitivity tests are based on the temperature and water table time series and soil profile at the Guisveld site; the model run indicated as 'normal' represents the validation model run for Guisveld high water table in Fig. 4. Note different scale for the latter graph.*

Table 2. *Results of sensitivity tests (see text). In the third to sixth column, all test results are compared with the Guisveld high water table validation model run depicted in Fig. 4. The total GHG computation is based on a global warming potential of 62 x CO2 for CH4, at a time horizon of 20 years.*

Sensitivity test	Parameter change	Effect with respect to unchanged parameters (multiplication factor)			
		CO_2 total	CO_2 from peat	CH_4	Total GHG
Water table	water table –0.1 m	1.86	2.43	0.54	1.65
	water table –0.2 m	2.46	3.50	0.35	2.13
	no fluctuation, at –0.08 m	0.68	0.57	1.48	0.81
Temperature	temperature +2° C	1.07	1.10	1.55	1.15
	temperature +4° C	1.09	1.17	1.85	1.21
Primary production	decrease by factor 0.5	0.92	1.00	0.72	0.89
	increase by factor 1.5	1.08	1.00	1.30	1.12
Decomposition roots/litter k	decrease by factor 0.5	1.00	1.00	1.03	1.01
	increase by factor 2.0	1.00	1.00	0.96	1.00
Methane oxidation in plants	decrease P_{ox} by 0.2	0.99	1.00	2.79	1.27
	decrease P_{ox} by 0.24	0.97	1.00	4.57	1.54
Priming effect	$c_{prim} = 5$	0.83	0.70	1.00	0.86
	$c_{prim} = 0$ (no priming)	0.65	0.38	0.99	0.71
Soil profile data	Ruwiel profile	0.58	0.26	1.45	0.71
	Horstermeer profile	0.55	0.24	1.28	0.66

Water table decrease experiments in steps of 0.1 down to 0.6 m serve as an example model application (Fig. 7). The increase of the CO_2 flux is approximately linear with deeper water table. This decrease is largely caused by an increase of peat decomposition (nearly 10 times increase at a 0.6 m deeper water table). The CH_4 flux is reduced to a tiny fraction (0.037) of the original flux. This decrease is most rapid at the first 0.2 m of water table decrease. The effect on total GHG flux is a nearly five-fold increase.

Temperature. The temperature has been increased by +2° C and +4° C. CH_4 shows a strong increase by a factor of 1.85 at +4° C, the CO_2 flux increases by a factor of 1.09. In particular the summer CH_4 flux increases (Fig. 6, Table 2), while also considerable pattern changes occur at the +4° C run. These nonlinear pattern changes are caused by the many different ways in which the CH_4 flux is influenced in the model: temperature affects primary production, CH_4 production, CH_4 oxidation and transport by plants. Since Q_{10} for CH_4 oxidation is lower than that of CH_4 production, the net effect is a strong increase of the CH_4 flux with temperature. The total GHG effect is also an increase by a factor of 1.21 at +4° C. These temperature experiments also demonstrate the sensitivity of SOM decomposition to climate change. A further application of the model in climate change experiments is described by Van Huissteden (2004).

Primary production. The primary production has not been measured at the field sites; values have been estimated from vegetation characteristics. We tested the model's sensitivity by multiplying primary productivity P with a factor of 0.5 and of 1.5. Both CH_4 and CO_2 increase with increasing primary productivity (Table 2). The effect on CH_4 is stronger (~30% increase/decrease) than that on CO_2 (8%). The increase affects mainly the summer fluxes.

Decomposition constants of fresh organic matter. The decompositon constants of fresh organic matter (k) are difficult to measure and have been derived from literature. For rhizodeposition very rapid decomposition is assumed ($k = 100$ d^{-1}), decomposition constants for the root and litter reservoirs are set at values ($k = 5$ d^{-1}). We experimented for the Guisveld high WT site by multiplying k for rhizodeposition and roots and litter with factors of 0.5 and 2.0. Changing the decomposition constant for rhizodeposition resulted in negligible change of the fluxes, the effect of the roots and litter decomposition constant is very minor; less than 0.5% for CO_2 and up to 3% for CH_4 (Table 2). The decrease of the CH_4 flux is caused by a more rapid depletion of labile organic matter reservoirs at low water table.

Methane oxidation during plant transport. A certain amount of methane is oxidized during plant transport, expressed by the parameter P_{ox} in eq. 13. This factor is very difficult to determine. We experimented with different values of P_{ox}. (0.5, 0.7 and 0.9). Since most of the CH_4 flux consists of plant transport the resulting flux is very sensitive to this parameter,

resulting in a more than four-fold increase at a decrease of P_{ox} down to 0.5 (Table 2) and a large effect on the total GHG flux.

Priming effect. The correction for the priming effect enhancement of peat decomposition (see above) has been tested by applying different values for c_{prim} (eq. 17), with values of 10, 5 and 0, the first being the value applied in the model validation runs for Guisveld, while the latter value results in no priming correction. The effect is considerable, in particular on the peat decomposition. Peat decomposition is reduced by a factor of 0.38 at absence of the priming effect, and a reduction of the total GHG flux of 0.71 (Fig. 6, Table 2).

Soil profile. The effect of differences in the soil profile data (organic matter content, bulk density, pF curves) has been tested by replacing the soil profile data file of Guisveld with that of the Horstermeer and Ruwiel sites, keeping all other input the same. For both profiles, the CO_2 flux is decreased and the CH_4 flux increased. The difference between the Horstermeer and Ruwiel profiles is small. Although these profiles are geogenetically different, their physical properties are very similar. The Guisveld profile deviates more strongly by a much larger organic matter content, which explains its higher CO_2 flux. The higher CH_4 flux for the Horstermeer and Ruwiel profiles is explained by differences in the moisture retention (pF) curves. These cause a higher saturation with water in the topmost horizon of the Horstermeer and Ruwiel profiles, and consequently a lower aerobic decomposition of the labile SOM reservoirs. This results in more labile SOM that is available for methanogenesis. The water retention curve for Guisveld has been estimated from Wösten et al. (1994); more precise site-specific data might have reduced the difference in the CH_4 fluxes.

Discussion

The validation shows that PEATLAND is capable of simulating of the seasonal variation in the fluxes of CO_2 and CH_4 in five sites with different land management. The magnitude of the simulated fluxes agrees well with the measured fluxes. For the CO_2 model, this is achieved without site-specific adjustment of parameters. For the CH_4 model, some site specific adjustment in the plant transport oxidation parameter has been made, which can be justified on the basis of differences in vegetation at the sites. The seasonal change of the fluxes and reaction to related temperature and water table change is modelled correctly. However, the effects of shorter timescale features related to synoptic weather events may result in differences between model and data, insofar these could be tested against the field measurements with their low temporal resolution. These differences can be explained by incompletely known model input data, in particular organic production, or processes that are not incorporated into the model. For instance, the high summer flux peak in the Guisveld, Ransdorp and Kamerik sites after a heavy rainfall event in the first summer may be

the result of downward transport of fresh organic matter from the surface by water.

Validation has been restricted to the measured surface fluxes. Further aspects of the model that are amenable to validation are results of the soil physical submodel, such as soil temperature and soil moisture, and the pore water methane concentration. Unfortunately relevant time series were not available from the measurement sites; measurement of temperature and methane concentration profiles is in progress at the Ruwiel and Horstermeer sites. Validation of the size of the SOM pools is difficult since these are difficult to measure.

The explicit modelling of the decomposition of the peat substrate allows an estimate of the effects of continued peat degradation due to water-level lowering for agricultural purposes. Discrepancies between data and model for the CH_4 submodel may have been caused by soil cracking at dry spells and downward transport of fresh organic matter by rainwater. The model therefore could improve by including a more adequate soil moisture model, simulation of soil shrinkage and transport of dissolved and particulate organic matter. In particular dissolved organic matter export may play a significant role in the carbon balance of peatlands (Freeman et al., 2004). Also we used a simplified approach to include the effect of soil aeration. This could be improved by modelling of O_2 diffusion. However, further refinement of the model can be achieved only at the cost of increasing the number of soil physical parameters, which has the drawback of introducing more parameter uncertainty.

The sensitivity analysis shows that PEATLAND is sensitive to uncertainty in the following input data: 1) water table; 2) air temperature; 3) net primary production (in particular the CH_4 submodel); 4) peat decomposition rate and the priming effect; 5) factors influencing methane transport by plants; 6) soil physical data, in particular organic matter content, dry bulk density and moisture retention. In practice, water table, air temperature, net primary production estimates and organic matter content of the soil will not provide large problems, since these can be obtained from field measurements. The influence of peat decomposition rate and soil physical data is not very large. Estimation of these parameters instead of using measurement data should not inhibit realistic flux estimates with the model.

On the other hand, the influence of the CH_4 plant transport rate parameters is substantial. We therefore regard the plant oxidation rate parameter P_{ox} as a tuning parameter, rather than the methane production factor R_0 (eq. 10) which has been used by Walter (2000) for model tuning. In general, wetland plants are efficient transporters of gases, and the proportion of wetland plants (reeds, rushes, sedges) vs. species adapted to drier sites should give some indication of P_{ox}. Most graminoid wetland plants are efficient transporters of CH_4 (Walter, 2000), resulting in a rather low value of P_{ox} when present in the vegetation. However, plant transport and methane oxidation

in plants is species-dependent, and moreover depends on assimilation rates and microclimatic factors (e.g. Van der Nat & Middelburg, 1998; Busch & Lösch, 1999). Physically more realistic formulations than that of Walter (2000) have been applied in models by Segers and Leffelaar, (2001); Segers et al., (2001) and Beckett et al. (2001). However, these require more extensive parametrization of the root system.

Peat degradation in the topsoil is underestimated when peat decomposition rates are based on the rates obtained from incubation experiments (Van den Bos & Van de Plassche, 2003a, b). We attribute a significant influence on peat degradation to the effect of priming as described by Kuzyakov et al. (1999, 2000). Kuzyakov et al. (2000) suggest several possible mechanisms for the priming effect, of which co-metabolism of the peat substrate with root exudates, and fertilization of the microbial population by exudates are the most likely. The peat substrate, which decomposes relatively easy compared with the stabilized humus studied by Kuzyakov et al. (1999), may be quite sensitive to priming. Freeman et al. (2004) have demonstrated the potential importance of the priming effect in Welsh peatlands. Our model could not reproduce the peat decomposition rates derived by Van den Bos and Van de Plassche (2003a, b) without including a considerable priming effect. However, the 60% contribution of peat decomposition to total soil CO_2 flux obtained by these authors might be an overestimate for high water table sites. The priming effect should be validated further using C labelling experiments. The priming effect may be of particular importance in the highly productive grasslands we studied.

Another source of uncertainty is the CH_4 production rate constant R_0, according to Walter (2000) a tuning parameter in the CH_4 submodel. Because of the uncertainty in the plant CH_4 oxidation parameter P_{ox}, it cannot be evaluated whether site-to-site variation in this parameter exists. It is not unlikely that R_0 differs among high and low WT sites, because of different bacterial populations adapted to site-specific environmental conditions. The presence of alternative electron acceptors (SO_4^{2-}, Fe oxides) may depress methanogenesis (e.g. Segers and Kengen, 1998; Segers and Leffelaar, 2001). Considerable variation in hydrochemistry exists between the sites, in particular the Horstermeer site deviates with its strong seepage. Variation in bacterial populations within and between the sites has been demonstrated (Fernández Raga, 2004). Besides between-site variation in methane production, also vertical variation of R_0 within the soil profile may exist. Van den Pol - Van Dasselaar et al. (1999b) demonstrate that highest CH_4 production capacity occurs slightly below the surface, rapidly decreasing with depth with a factor of 4 to 5. This indicates that a more active methanogenic microbial community exists in the topsoil where most of the labile SOM is produced. Likewise, parametrization for the CH_4 oxidation rates may be site or depth dependent since oxidation rates cited in the literature also show considerable variation (Segers,

1998). In our application the parameters have been set at high values, in agreement with generally high values found in Dutch peat soils (Van den Pol - Van Dasselaar, 1999a).

The model has not been validated against methane concentrations in the soil pore water, because current data availability is insufficient. However, available measurements of the CH_4 concentration in the soil at the Horstermeer and Ruwiel sites suggest high methane concentrations in the soil pores (up to 50000 $\mu M \cdot l^{-1}$), partly in the shape of bubbles. This suggests that CH_4 cannot readily escape by ebullition in these clayey soils. The formulation of the ebullition process cf. Granberg et al. (2001), which includes resistance to bubble movement, is therefore more appropriate than the formulation by Walter (2000), which assumes rapid bubble transport. For the Horstermeer and Ruwiel sites the methane threshold concentration C_{thresh} was set at a high value that excluded ebullition (Table 1).

The strong increase of the CO_2 flux from peat after drainage (Fig. 7) should be a transient phenomenon. Decomposition of the peat results in surface subsidence and a relative rise of the water table (Schothorst, 1977). Presently this effect is not incorporated into the model. Also the reverse effect – surface rise by peat growth – has not been incorporated, but the model simulates an increase of the roots and litter reservoirs at high water table simulations.
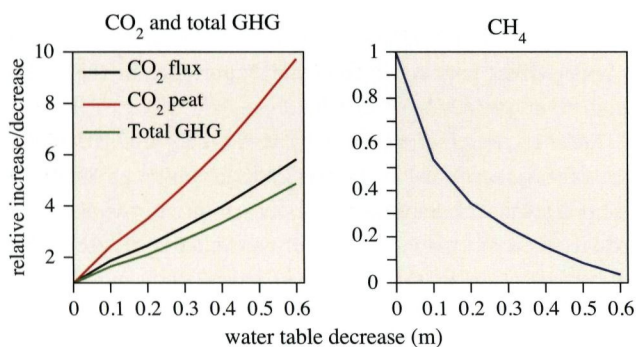

Fig. 7. Water table decrease model experiments. The water table is lowered in steps of 0.1 m down to 0.6 m in the Guisveld site. The effects on the fluxes are shown as decrease/increase relative to the validation model run for Guisveld high water table in Fig. 4 (= 1.0). Left: CO_2 and total GHG flux, right: CH_4 flux.

The model results suggest tentative implications for management of the studied peatlands. The experiments with water-table changes show that lowering of the water table increases the GHG fluxes from these peat soils by increasing CO_2 production from the peat substrate. Conversely, raising the water table reduces the CO_2 production from peat. The reduction of the CO_2 flux is partly compensated by an increase in the CH_4 flux. However, the water table sensitivity experiments show that even if this effect is included, the effect of a water table rise on the total GHG flux remains a net decrease, in line with Van den Bos (2003).

Conclusions

The integrated modelling of CO_2 and CH_4 fluxes in PEATLAND enables assessment of the effects of land management and climate change on greenhouse-gas emission from peat soils in the Netherlands. Parameter requirements for the model are relatively modest. Further refinement of the model can be achieved, but at the cost of increasing the number of parameters, making it more vulnerable to parameter uncertainty, and restricting its use. Combined with site flux measurements for calibrating the model, it should be capable of extending C flux estimates over larger spatial or time scales. More accurate modelling of fluxes at specific sites will require a measured set of soil physical and chemical data, and primary production data.

The water table sensitivity experiment indicate that continued drainage contributes to the GHG fluxes of Dutch peatlands by an enhancement of peat decomposition, and that raising the water level may reduce the GHG flux despite an increase of the CH_4 flux. Furthermore, higher temperatures increase the GHG fluxes from peat soils, indicating a positive feedback on climate warming. A particular interesting feature is the assumed priming effect (enhancement of peat decomposition by the addition of fresh organic matter to the soil by a high root exudate production). Without assuming a considerable priming effect, observed peat decomposition rates cannot be modelled adequately. It is likely that the priming effect plays an important role in peat degradation of Dutch peat soils.

Acknowledgements

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