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Climate-CH₄ feedback from wetlands and its interaction with the climate-CO₂ feedback

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Abstract. The existence of a feedback between climate and methane (CH₄) emissions from wetlands has previously been hypothesized, but both its sign and amplitude remain unknown. Moreover, this feedback could interact with the climate-CO₂ cycle feedback, which has not yet been accounted for at the global scale. These interactions relate to (i) the effect of atmospheric CO₂ on methanogenic substrates by virtue of its fertilizing effect on plant productivity and (ii) the fact that a climate perturbation due to CO₂ (respectively CH₄) radiative forcing has an effect on wetland CH₄ emissions (respectively CO₂ fluxes at the surface/atmosphere interface).

We present a theoretical analysis of these interactions, which makes it possible to express the magnitude of the feedback for CO₂ and CH₄ alone, the additional gain due to interactions between these two feedbacks and the effects of these feedbacks on the difference in atmospheric CH₄ and CO₂ between 2100 and pre-industrial time (respectively ΔCH_4 and ΔCO_2). These gains are expressed as functions of different sensitivity terms, which we estimate based on prior studies and from experiments performed with the global terrestrial vegetation model ORCHIDEE.

Despite high uncertainties on the sensitivity of wetland CH₄ emissions to climate, we found that the absolute value of the gain of the climate-CH₄ feedback from wetlands is relatively low (<30 % of climate-CO₂ feedback gain), with either negative or positive sign within the range of estimates.

Whereas the interactions between the two feedbacks have low influence on ΔCO_2 , the ΔCH_4 could increase by 475 to 1400 ppb based on the sign of the C-CH₄ feedback gain.

Our study suggests that it is necessary to better constrain the evolution of wetland area under future climate change as well as the local coupling through methanogenesis substrate of the carbon and CH₄ cycles – in particular the magnitude of the CO₂ fertilization effect on the wetland CH₄ emissions – as these are the dominant sources of uncertainty in our model.

1 Introduction

Increased atmospheric CO₂ due to anthropogenic emissions is expected to lead to significant climate change in the 21st century (IPCC, 2007). Such climate change may indirectly affect the atmospheric CO₂ concentration by modifying the exchange of carbon between the atmosphere and the land and ocean. Several models have evaluated this climate-carbon cycle interaction, generally finding a positive feedback between climate change and the global carbon cycle (Friedlingstein et al., 2006). Methane (CH₄) is a very efficient greenhouse gas, with a Global Warming Potential of 25 (for given time horizon to 100 years) (IPCC, 2007), and is currently the second anthropogenic greenhouse gas after CO₂. Very few studies have investigated the potential feedback between CH₄ emissions by wetlands and climate.

CH₄ emissions from wetlands, the largest natural source in the present-day global CH₄ budget, are directly controlled by climatic conditions (e.g. Christensen et al., 2003). CH₄



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emissions from wetlands depend on the global areal extent of wetlands (Ringeval et al., 2010; Bloom et al., 2010) and on the emission rate of these wetlands (e.g. Conrad, 1989; Fung et al., 1991). Both of these terms are controlled by, among other variables, soil temperature and hydrology. For instance, temperature controls the rate of methanogenesis, exerts a control on the quality and quantity of organic material substrate for CH₄ production and has an influence on the area of wetlands through control of surface evaporation and the soil water budget. There is a large uncertainty in current global wetland emissions (estimates range from 115; Fung et al., 1991 up to 237 TgCH₄ yr⁻¹; Hein et al., 1997). Because the sensitivity of wetland CH₄ emissions to the environmental control factors is poorly understood, the behavior of wetland CH₄ emissions under future climate change (e.g. Updegraff et al., 2001) and the amplitude of the resulting climate-CH₄ emission feedback is far from being well understood.

Several studies have estimated changes in CH₄ emissions from wetlands under future climate change: e.g. Shindell et al. (2004) (hereinafter SWF04), Gedney et al. (2004) (hereinafter GCH04), Eliseev et al. (2008) (EMA08) and Volodin (2008) (V08). They all found an increase in CH₄ emissions despite differences in processes accounted for. GCH04 and EMA08 also estimated the resulting climate-CH₄ feedback and both found it to be relatively small. The additional warming induced by this feedback is small (e.g. only 3.7–4.9 % of the total projected warming by 2100 under the IS92a scenario found by GCH04). SWF04 accounts for changes in wetland area, using thresholds for variables they define as influencing wetland CH₄ emissions while GCH04 use a more realistic approach using a subgrid topographical model. EMA08 and V08 do not account for change in wetland extent. In these approaches, base CH₄ emissions are calculated using an empirical approach: parameterization for GCH04 and EMA08 and correlations between climate anomalies and wetland CH₄ emissions derived under current conditions for SWF04. With the exception of V08, none of these studies account for increasing CO₂ and its effect on plant productivity and hence on soil carbon available for methanogenesis. Similarly, they do not account for the climate change (driven by CO₂ or CH₄) effect on soil carbon dynamics and hence on CH₄ emission rates. The strategy used by V08 is based on a more process-based approach which could allow accounting for these two effects but the contribution of each driver is not discussed.

In fact there is a tight coupling between the climate-CO₂ feedback and the climate-CH₄ feedback. As mentioned before, increasing atmospheric CO₂ has a direct concentration effect on wetland CH₄ emissions. Moreover, CO₂-induced climate change will affect CH₄ emissions, and hence CH₄ concentration and climate. CH₄-induced climate change will in turn affect the land and ocean CO₂ cycle and hence atmospheric CO₂ and climate. The combined effect of these two feedbacks (climate-CO₂ and climate-CH₄) needs to be

explicitly accounted for in order to estimate the overall response of the coupled CO₂ cycle – CH₄ cycle – climate system.

Friedlingstein et al. (2003) expressed mathematically the magnitude of the climate-carbon cycle feedback using a gain formalism following Hansen et al. (1984). Here, we revisit this theoretical framework, first applying it to the climate-CH₄ gain in the absence of CO₂ perturbation; then generalizing it to the climate, CO₂ and CH₄ interactions. These gains and the interaction between the feedbacks are expressed as functions of sensitivity terms that we estimate from the values reported in the literature and from simulations performed with the ORCHIDEE global terrestrial carbon cycle model. Once these terms are estimated, we quantify the different gains and the increase of atmospheric CH₄ and CO₂ due to the feedbacks and their interactions.

2 Theoretical analysis

In the following, the climate-CH₄ emissions by wetlands feedback will be referred hereafter as “C-CH₄ feedback” as well as “C-CO₂ feedback” for climate-carbon cycle feedback (both terrestrial and oceanic) in the sense of Friedlingstein et al. (2006).

2.1 C-CH₄ feedback analysis

Similarly to the C-CO₂ feedback analysis by Friedlingstein et al. (2003), we assume that the coupling between CH₄ emissions by wetlands and the climate system can be linearized by the following set of equations:

$$\Delta\text{CH}_4 = F_{\text{MF}} + F_{\text{NAT}}^{\text{add}} - F_{\text{MA}}^{\text{add}} \quad (1a)$$

$$\Delta T = \alpha_{\text{M}} \Delta\text{CH}_4 \quad (2a)$$

where ΔCH_4 (in GtC) is the difference of CH₄ concentration in the atmosphere between a given time, t_1 , and the initial state, t_0 , defined here as the preindustrial state estimated at 1860. ΔT (in K) is the change in global air temperature due to the change in CH₄ concentration. Equation (2a) can also be considered as a linearization of the more stringent, square root dependence between CH₄ radiative forcing and its concentration (IPCC, 2001). F_{MF} (GtC) represents the integral over the period since t_1 of the anthropogenic emissions of CH₄. $F_{\text{NAT}}^{\text{add}}$ (GtC) represents the integral of the change in natural CH₄ emissions relative to the preindustrial emissions baseline. As the focus of this study is on wetlands, we assume here that $F_{\text{NAT}}^{\text{add}}$ represents the change in CH₄ emissions by wetlands only. Even though other natural sources (such as biomass burning) are also climate dependent, and the general framework presented here applies to other CH₄ sources and sinks as well, we will focus only on the wetland component as assessment of climate-CH₄ feedbacks from all natural sources and sinks is beyond the scope of this paper.

The last term of Eq. (1a), F_{MA}^{add} (GtC), is the integral of the atmospheric sink of CH₄ through reaction with OH radicals (again relative to the preindustrial baseline) and closes the CH₄ budget. For the pre-industrial state, we assume here that CH₄ concentration was constant (apart from interannual to decadal variations), and hence natural sources were balanced by the atmospheric OH sink (and the minor soil sink neglected in Eq. 1a). Departure from that steady-state equilibrium can be represented by Eq. (1a), using a perturbation approach, accounting only for additional sources and sinks. The change in CH₄ emissions, integrated over time t_1-t_0 , can be driven by a change in climate and by a change in CH₄ concentration. As in Friedlingstein et al. (2003), we use a single global ΔT as a proxy for climate change. It is clear that a change in emissions could be also driven by changes in hydrology, and that regional variations in both the magnitude of ΔT and hydrology will also occur, but we assume here that these other climate variables change would scale with global temperature.

The integral of additional natural sources of CH₄ is then expressed by:

$$F_{NAT}^{add} = \beta_M \Delta CH_4 + \gamma_M \Delta T \quad (3a)$$

where β_M (unitless) and γ_M (in GtC K⁻¹) are the CH₄ flux sensitivities to the atmospheric CH₄ concentration and to climate, respectively. The β_M term results from the CH₄ atmospheric concentration affecting the CH₄ flux through its control on diffusion (via soil air or plants) from wetland soils to the atmosphere. The Eq. (3a) is constructed by analogy with that for CO₂ given by Friedlingstein et al. (2006, Eq. 7a). Even if the effect of increased atmospheric CH₄ concentration on concentration gradient between soil and atmosphere (and thus the value of β_M) is presumed small (atmospheric concentration in CH₄ ~ 1% of wetland soil concentration), we keep it to be consistent with CO₂. Although there is evidence that, at the site scale and on sub-annual timescales, an exponential dependence of CH₄ flux to temperature is observed (e.g. Christensen et al., 2003), Eq. (3a) here aims to represent the overall global response of wetlands to climate (not just temperature). To remain simple and comparable to the CO₂ framework, we thus assume that a linear relationship is appropriate. More investigations concerning (i) the relationship between global climate and global wetland CH₄ emissions and (ii) the range of temperature over which such a relationship may be valid are required. These investigations would be based, for instance, on long-term or interannual time scales.

Finally, the integral of the additional atmospheric sinks can be expressed by:

$$F_{MA}^{add} = \int_{t_0}^{t_1} \frac{\Delta CH_4(t)}{\tau} dt \quad (4a)$$

as the additional sink at each time step is assumed here to be equal to $\frac{\Delta CH_4(t)}{\tau}$, where τ is the atmospheric lifetime of

CH₄. We assume here that τ is constant in time. There is a slight dependency of τ on CH₄ concentration and on climate (IPCC, 1994) which is neglected here. In doing so, we de facto assume that there is neither year-to-year variability nor any trend in atmospheric OH concentration. Recent findings seem to indicate that global OH is quite stable (Montzka et al., 2011). In order to solve the set of Eqs. (1a) to (3a), one must linearize the sink term. Here by applying the mean value theorem, the integral of the changes of CH₄ sink over time (between the time t_1 and preindustrial period t_0) can be written as proportional to the change of CH₄ at time t_1 .

$$F_{MA}^{add} = \mu \frac{\Delta CH_4}{\tau} (t_1 - t_0) \quad (4b)$$

with μ considered here as a constant for a given scenario of CH₄ increase. For instance, μ would be equal to 0.5 if CH₄ concentration increases linearly with time. For a given scenario of atmospheric CH₄ increase, μ can be diagnosed as the ratio of the cumulative changes of CH₄ along the full length of the scenario to the change of CH₄ at the end of the scenario (equating the right members of Eqs. 4a and b).

Equation (1a) now reads:

$$\Delta CH_4 = F_{MF} + \beta_M \Delta CH_4 + \gamma_M \Delta T - \mu \frac{\Delta CH_4}{\tau} \Delta t \quad (1b)$$

We can now express the amplitude of the feedback using a “gain” as Friedlingstein et al. (2003) did for CO₂. Combining Eqs. (2a) and (1b) we have:

$$\Delta CH_4^{COU} = \frac{1}{1 - g_M} \Delta CH_4^{UNC} \quad (5)$$

with

$$\Delta CH_4^{UNC} = \frac{F_{MF}}{(1 + \frac{\mu}{\tau} \Delta t - \beta_M)} \quad (6)$$

and

$$g_M = \alpha_M \gamma_M / \left(1 + \frac{\mu}{\tau} \Delta t - \beta_M\right) \quad (7a)$$

ΔCH_4^{COU} is the change of atmospheric CH₄ concentration in the case of C-CH₄ feedback while ΔCH_4^{unc} is the change of atmospheric CH₄ concentration in the absence of C-CH₄ feedback (i.e. $\gamma_M=0$). g_M is the gain of this feedback and it is larger if: α_M and γ_M are positive and large and if β_M is positive and low. This is analogous to the C-CO₂ feedback gain defined in Friedlingstein et al. (2003) as:

$$g_c = -\alpha_c \gamma_c / (1 + \beta_c) \quad (7b)$$

2.2 Cross feedbacks

The previous feedback analysis was done for the case of a changing CH₄ concentration alone, together with climate. Here we extend the gain formalism in the more realistic case where both CO₂ and CH₄ vary at the same time.

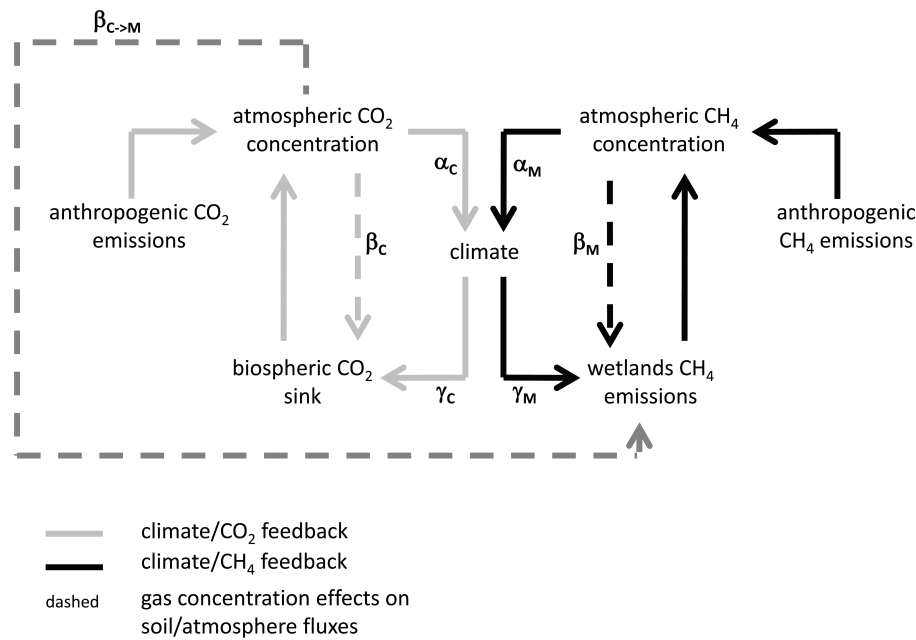


Fig. 1. Schematic representation of the C-CO₂ feedback (light grey), the C-CH₄ feedback (black) and of their interaction. Gas concentration effects on the soil/atmosphere fluxes are represented by dashed lines. The different sensitivity terms (α , β , γ) are listed on this schematic representation.

First, as mentioned before, both CO₂ and CH₄ will affect the climate through their radiative forcing. ΔT should be now expressed as:

$$\Delta T = \alpha_C \Delta \text{CO}_2 + \alpha_M \Delta \text{CH}_4. \quad (2b)$$

As a consequence, $F_{\text{NAT}}^{\text{add}}$ is affected by the change in climate (Eq. 3a) regardless of whether this climate change is induced by an increase in atmospheric CH₄ (as showed before) or by an increase in atmospheric CO₂. The same applies to a climate-induced change in land CO₂ sinks. Interactions resulting from the Eq. (2b) will be referred to as the “*climate interaction*” below.

The second interaction comes from the dependence of CH₄ emissions to atmospheric CO₂. Increasing atmospheric CO₂ is believed to enhance plant photosynthesis (fertilization effect) (e.g. Norby et al., 2005). As a result, rising CO₂ should increase the amount of available organic substrate for methanogenesis and hence CH₄ emissions from wetlands (see discussion). This effect is expressed by an additional term (β_C) in the original Eq. (3a):

$$F_{\text{NAT}}^{\text{add}} = \beta_M \Delta \text{CH}_4 + \gamma_M \Delta T + \beta_{C \rightarrow M} \Delta \text{CO}_2. \quad (3b)$$

This interaction will be called the “*fertilization interaction*” hereafter.

The different sensitivity terms (α , β , γ) are listed on a schematic representation of the feedbacks and of their interaction in Fig. 1.

Other minor interactions could be expressed between CO₂ and CH₄ (e.g. oxidation of CH₄ in atmosphere [EMA08] or

in the oxic part of wetland soils releases CO₂) but these are not accounted for in our modelling approach below and are not quantified here.

One can introduce Eq. (2b) into Eq. (3b) then combine the resulting expression with Eq. (4b) into Eq. (1a) to obtain the following Eq. (8). Then, doing the same work for CO₂ (see Appendix A), we can obtain a two equations system with 2 unknowns (ΔCO_2 and ΔCH_4), i.e.

$$\begin{cases} -(\beta_{C \rightarrow M} + \gamma_M \alpha_C) \Delta \text{CO}_2 + (1 + \frac{\alpha_C}{\tau} \Delta t - \beta_M - \gamma_M \alpha_M) \Delta \text{CH}_4 = F_{\text{MF}} & (8) \\ (1 + \beta_C + \gamma_C \alpha_C) \Delta \text{CO}_2 + \gamma_C \alpha_M \Delta \text{CH}_4 = F_{\text{CF}} & (9) \end{cases}$$

Using this system, we can express ΔCO_2 (and ΔCH_4) as a function of the integral of anthropogenic CO₂ and CH₄ emissions, i.e. respectively F_{CF} and F_{MF} (or $\Delta \text{CO}_2^{\text{unc}}$ and $\Delta \text{CH}_4^{\text{unc}}$; using Eq. (6) and its equivalent for CO₂). We show in the following the CH₄ and CO₂ gains for the idealized (and simpler) case where $\beta_{C \rightarrow M}$ is null (i.e. no fertilization interaction). This allows keeping symmetry between CO₂ and CH₄. The more realistic case, accounting for this $\beta_{C \rightarrow M}$ term and the introduced asymmetry is given in Appendix B. Although not shown until the Appendix, this term was taken into account in all the calculations of the next sections.

For the coupled climate-CO₂-CH₄ system neglecting fertilization interaction, we obtain now:

$$\begin{aligned} \Delta \text{CH}_4^{\text{COU}} &= \frac{1}{1 - \left[g_M + \frac{g_C g_M}{1 - g_C} \right]} \Delta \text{CH}_4^{\text{UNC}} & (10) \\ &+ \frac{1}{1 - \left[g_M + \frac{g_C g_M}{1 - g_C} \right]} \frac{\alpha_C}{\alpha_M} \frac{g_M}{1 - g_C} \Delta \text{CO}_2^{\text{UNC}} \end{aligned}$$

and

$$\Delta\text{CO}_2^{\text{COU}} = \frac{1}{1 - \left[g_C + \frac{g_C g_M}{1 - g_M} \right]} \Delta\text{CO}_2^{\text{UNC}} \quad (11)$$

$$+ \frac{1}{1 - \left[g_C + \frac{g_C g_M}{1 - g_M} \right]} \frac{\alpha_M}{\alpha_C} \frac{g_C}{1 - g_M} \Delta\text{CH}_4^{\text{UNC}}$$

Equation (10) shows that the interaction between CO₂ and CH₄ results in an additional gain in the first term of the right hand side of the equation. For CH₄, this additional gain is $g_C g_M / (1 - g_C)$ and is in addition to the initial gain considering CH₄ alone, g_M . It represents the overall contribution on the CH₄ concentration of the positive climate-CO₂ feedback initiated by the original emission-induced change in CH₄ concentration (climate interactions loop) in the case of no CO₂ anthropogenic emissions, i.e. when $\Delta\text{CO}_2^{\text{unc}} = 0$.

If we then account for anthropogenic CO₂ emissions, an additional contribution to ΔCH_4 appears: the second term in the right hand side of the Eq. (10). This originates from the anthropogenic emissions of CO₂, which induces an increase in the CO₂ concentration ($\Delta\text{CO}_2^{\text{UNC}}$). This CO₂ increase induces a climate change that will affect CH₄ emissions and hence CH₄ concentrations. In Eq. (10), $\Delta\text{CO}_2^{\text{UNC}}$ is multiplied by $\frac{\alpha_C}{\alpha_M} \frac{g_M}{1 - g_C}$ to obtain its equivalent in ΔCH_4 . Finally, it is multiplied by the same net feedback factor as one in the front of $\Delta\text{CH}_4^{\text{UNC}}$. Anthropogenic emissions of CO₂ are independent of CH₄ and thus cannot be expressed as a function of ΔCH_4 . This prevents us from fully expressing the total additional gain of each feedback in the case of coupling between CO₂, CH₄ and climate. Obviously, the same interpretation can be done to C-CO₂ feedback in presence of CH₄ with Eq. (11).

Changes in CO₂ and CH₄ can hence be computed from Eqs. (10) and (11), once the different sensitivity terms (α , β , γ) are estimated. This is the aim of the next section.

3 Estimates of the gain components

In this section, we will first use simulations performed with ORCHIDEE, a dynamic global vegetation model (DGVM), to estimate the wetland emission sensitivity terms (β_M , $\beta_{C \rightarrow M}$ and γ_M) as well as terms relative to C-CO₂ cycle. We will also make use of previous estimate of future changes in CH₄ emissions taken from the available literature. This will allow us to estimate the range of the climate-CH₄ gain and its effect on the projection of atmospheric CH₄, CO₂ and global temperature (Sect. 4).

3.1 Based on an ORCHIDEE modelling approach

3.1.1 Wetland CH₄ emissions modelling into ORCHIDEE

ORCHIDEE simulates the land energy, hydrology and the carbon cycle (Krinner et al., 2005). The version used here was further developed to incorporate CH₄ emissions from wetlands. The computation of wetlands CH₄ emissions is based on the modelling of wetland area dynamics as well as one of the CH₄ flux by surface unit. The resulting model will be named ORCHIDEE-WET hereafter.

In ORCHIDEE-WET, wetland area dynamics were computed using the TOPMODEL (Beven and Kirkby, 1979) approach of Decharme et al. (2006). For each gridcell, using both topographic heterogeneities and soil moisture computed by ORCHIDEE-WET, the TOPMODEL subroutine computes a sub-grid saturated fraction. Saturated areas as simulated by ORCHIDEE-WET do not correspond necessarily to water-logged soil and emitting wetland areas. Thus, we used a climatology (1993–2000) constructed from the Prigent et al. (2007) dataset as a baseline for our present day estimate. Future simulated wetland extent was then calculated from the ORCHIDEE-WET simulations, corrected to subtract the systematic biases between the present day simulated saturated area and observed wetland distributions. More details will be found in Appendix C.

At each time step, CH₄ flux densities (per unit area of emitting surface) were computed using a process-based model (Walter et al., 2001) for each sub-grid water-table class calculated as above. The model simulates CH₄ flux from natural wetlands based on the calculation of: (a) the methanogenesis in the saturated deeper soil horizons; (b) the methanotrophic oxidation in the aerated upper soil; and (c) the upward transport by diffusion, ebullition and/or plant-mediated transport (Walter and Heimann, 2000). When including the Walter et al. (2001) CH₄ emission model in ORCHIDEE, we made the same following modification, described in Ringeval et al. (2010). The substrate for methanogenesis is computed from the active soil organic carbon pool computed by ORCHIDEE rather than using linear regression against soil temperature and Net Primary Productivity (NPP) as is done in Walter et al. (2001) based on 6 sites. More information can be found in Appendix D. As in the initial Walter et al. (2001) model, methanogenesis sensitivity to temperature for each grid-cell is expressed by a function g as followed:

$$g = f(T(t, z)) \cdot Q_{10}^{T(t, z) - T_{\text{mean}}} \quad (12)$$

where $T(t, z)$ is the soil temperature at time t and depth z and $f(T)$ is a step-function equal to 0 if temperature is negative and 1 otherwise. A Q_{10} of 3, close to the mean value found in Ringeval et al. (2010), was used for all wetlands. As there is a high uncertainty about the value of Q_{10} (e.g. Valentine et al., 1994), a sensitivity test with a Q_{10} of 5.5 (in rough

agreement with higher value of range used by GCH04) is also performed. Relative to Walter and Heimann, “the temperature function describes the response to the seasonal variation of the soil temperature [...] relative to the annual mean temperature T_{mean} at the site”. As it is not clear if T_{mean} evolves in time or not, we have tested both configurations. Such a changing T_{mean} in time corresponds to the hypothesis that micro-organisms adapt relatively quickly to their environment (see Discussion).

The computation of a carbon stock whose active pool is used as a proxy for methanogenesis substrate is explained in a detailed way in Krinner et al. (2005). Briefly, in ORCHIDEE, the parameterizations of litter decomposition and soil carbon dynamics essentially follow Parton et al. (1988). Carbon dynamics are described through the exchanges of carbon between the atmosphere and the different carbon pools in plants and soils. Metabolic activity in the soil results in carbon fluxes within the three carbon pools (active, slow, and passive). Optimal residence times are prescribed for each pool, with temperature and moisture inhibition multipliers in order to parameterize the decrease of soil metabolic activity under cold or dry conditions. No modification is brought to ORCHIDEE-WET as regards soil wetlands conditions (see discussion below).

For each sub-grid water-table class given by TOPMODEL, ORCHIDEE-WET computes CH₄ fluxes with the corresponding mean water table depth value (respectively 0 and -5 cm). Other water table ranges could be calculated as well but would increase the time for calculation of the simulation. In the model, oxidation happens only in the second case, i.e. in the oxic soil layer between -5 cm and 0 for soils where the water table is 5 cm below the surface. The Q_{10} for methanotrophy is kept equal to the initial value (=2) of Walter et al. (2001). The model accounts also for oxidation when CH₄ entering the roots of plants has to pass through the small oxic zone around the root tips. A value of 50 % of the methane entering in the plant is considered as oxidized in the model (see Eq. 16 of Walter and Heimann, 2000). The CH₄ flux due to plant-mediated transport is a function of the Leaf Area Index (LAI) computed in ORCHIDEE-WET. As in the Walter et al. (2001) model, computation of a CH₄ flux which reaches the atmosphere by diffusive transport is based on the Crank-Nicolson scheme to resolve Fick’s first law. CH₄ atmospheric concentration serves as the upper boundary condition. Ebullition and transport by plant are not functions of the CH₄ atmospheric concentration in the model.

Under current climate forcing (the monthly NCEP climate forcing data corrected by CRU – N. Viovy, personal communication, 2009, <http://dods.extra.cea.fr/data/p529viov/cruncep/readme.htm>), ORCHIDEE-WET simulates a global mean wetland CH₄ emission flux of ~251 Tg yr⁻¹ over the 1990–2000 period. This is slightly above the upper end of IPCC range of estimates (100 to 231 TgC yr⁻¹) (IPCC, 2007). Both (i) the spatial extrapolation of parameter relative to CH₄ flux densities optimized on 3 sites and (ii) the

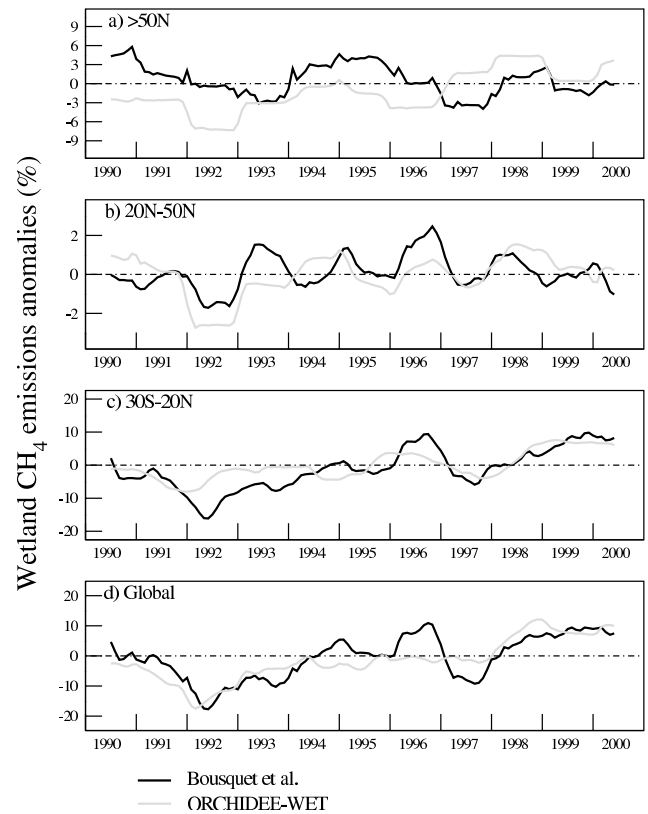


Fig. 2. Year-to-year variability of simulated CH₄ wetlands emissions (grey curve) and comparison with a top-down approach (Bousquet et al., 2006) (black curve) over 1990–2002 period. The anomalies obtained by 12 months-shift mean are divided by the global annual average of each estimation (Bousquet et al., 2006 or ORCHIDEE-WET).

mismatch between mean real annual wetland extent and mean annual Prigent et al. (2007) data can lead to high simulated wetland emissions. The distribution over latitude bands is 68, 53 and 125 Tg yr⁻¹ for boreal (>50° N), temperate (20° N–50° N) and tropical wetlands (30° S–20° N), respectively. High uncertainty remains for both total wetland emissions and their distribution. Wetland CH₄ emissions diagnosed from one atmospheric inversion Bousquet et al. (2006) give an estimation of 155 Tg at the global scale over the same period with a distribution of: 32, 21 and 95 Tg for the same latitude bands as above. Comparison of the year-to-year variability of wetland CH₄ emissions given by ORCHIDEE-WET and Bousquet et al. (2006) is shown on Fig. 2.

This modelling approach will allow us to estimate the wetlands CH₄ emissions sensitivities to climate and to atmospheric CO₂ and CH₄ concentrations for a transient run over the period 1860–2100. Some hydrological processes such as floodplain storage of water (Decharme et al., 2008) are not included in the model. Concerning the representation of permafrost, we account here for the freeze of the soil water content and decrease in soil carbon decomposition and

soil water holding capacity under these conditions but not for high carbon content in deep soil horizons which could be decomposed under warming, nor for the possible effects of thermokarst on lake and wetland expansion.

3.1.2 Experimental design

We forced the ORCHIDEE-WET model with climate fields taken from coupled ocean-atmosphere general circulation model (OAGCM) simulations and with the associated time-varying atmospheric CO₂ and CH₄ concentrations scenarios. This allows us to estimate the different sensitivity terms of Eqs. (10) and (11) (or equations in Appendix B in the most general feedback calculation framework). Four ORCHIDEE-WET simulations have been performed over the period 1860–2100. Each ORCHIDEE-WET simulation needs as forcing: climate, atmospheric CO₂ and CH₄ concentration values. For each forcing, we use either the pre-industrial state or a transient evolution from 1860 to 2100 following SRES-A2 scenario. The four ORCHIDEE-WET simulations are varied from one to the next by combining pre-industrial or transient forcing values as summarized in Table 1. This experimental design does not allow the testing of each term independently, nor their interaction effects, but is chosen to keep computational costs reasonable. Simulations 1 and 3 are also realized with a Q_{10} of 5.5 for methanogenesis to test the sensitivity of our results to this parameter. They will be called respectively Simulation 1- Q_{10} and Simulation 3- Q_{10} in the following. We performed all these simulations twice: first, considering a constant T_{mean} (see Eq. 12) and second, considering a T_{mean} varying in time.

The transient climate (1860–2100) is obtained from the IPSL-CM4 OAGCM simulations (Marti et al., 2010) with prescribed GHG-forcing for historical and future (SRES A2) scenarios. These climate fields were bias corrected by removing the difference between the climate model climatology (over the period 1961–1990) and the “observed” climatology (Sheffield et al., 2006 forcing data for air humidity and CRU – University of East Anglia’s Climate Research Unit, <http://www.cru.uea.ac.uk/> – for all other variables). For the pre-industrial climate forcing, we use a random succession of climate data taken from the first ten years (1860–1869) of the OAGCM simulation. The same climate data succession was used for all simulations with preindustrial climate forcing. Before the different simulations, ORCHIDEE-WET was first brought to equilibrium using preindustrial climate forcing. The simulation forced by pre-industrial climate, CO₂ and CH₄ concentration can be considered as the control simulation (CTRL hereafter).

By calculating the difference of the CH₄ emissions between the different simulations, we can isolate the CH₄ flux sensitivities to atmospheric CO₂, CH₄ and climate. The same is done with the net terrestrial CO₂ flux in order to get the carbon sensitivity terms. The difference between simulation 1 (change in atmospheric CO₂ only) and CTRL gives

Table 1. Set of ORCHIDEE-WET simulations. Performed ORCHIDEE-WET simulations are defined by climate, atmospheric CO₂ and CH₄ concentration values used as forcing. For each forcing, pre-industrial values (PI) or transient following SRES-A2 scenario (T) can be used.

	CO ₂	CH ₄	Climate
CTRL	PI	PI	PI
Simulation 1	T	PI	PI
Simulation 2	PI	T	PI
Simulation 3	T	PI	T

PI: Pre-industrial; T: Transient over 1860–2100

the sensitivity of CO₂ and wetland CH₄ emissions to atmospheric CO₂ (resp. β_C and $\beta_{C \rightarrow M}$). The difference between simulation 2 (change in atmospheric CH₄ only) and CTRL gives the wetland CH₄ flux sensitivity to atmospheric CH₄ (β_M); and the difference between simulation 3 (change in CO₂ and climate) and simulation 1 gives the sensitivities of CO₂ and wetland CH₄ flux to climate (γ_C and γ_M).

Furthermore, we also estimated the contribution of changes in wetland extent vs. changes in CH₄ emission rate. To do so, we remove a posteriori, the evolution of the wetland extent from the previous estimates. For each simulation and each year, the CH₄ flux densities calculated per unit wetland area are combined with the climatological pre-industrial (but seasonally varying) wetland area to compute the wetland CH₄ emissions in the absence of changes to the wetland extent. Comparison as described above of such emissions gives an estimate of $\beta_{C \rightarrow M}^f$ and γ_M^f , respectively the wetland CH₄ flux sensitivity to atmospheric CO₂ and to climate under constant wetland area. Regardless, using CH₄ flux densities and wetland area from two different simulations to compute wetland CH₄ emissions does not allow the possibility of removing the indirect effects of the variation of wetland extent on CH₄ fluxes: indeed, change in simulated wetland extent leads to change in the computed soil water content (through change in modelled runoff) that could have a small effect on the ORCHIDEE modelled carbon cycle.

Finally, the difference between Simulation 1- Q_{10} and 3- Q_{10} , after removing wetland area evolution, gives us the wetland CH₄ flux densities sensitivity to global climate with a higher Q_{10} , i.e. the sensitivity term $\gamma_{M-Q_{10}}^f$.

The different γ_M terms were computed for two cases: first, from simulations performed considering a constant T_{mean} and second, from simulations considering a T_{mean} that varies with climate. In the case where a constant T_{mean} is chosen, mean climatological pre-industrial surface temperature is used. If T_{mean} varies in time, it is computed in ORCHIDEE-WET by a slow relaxation method, as described by the Eq. (5) of Krinner et al. (2005) with a $\tau = 365$ days.

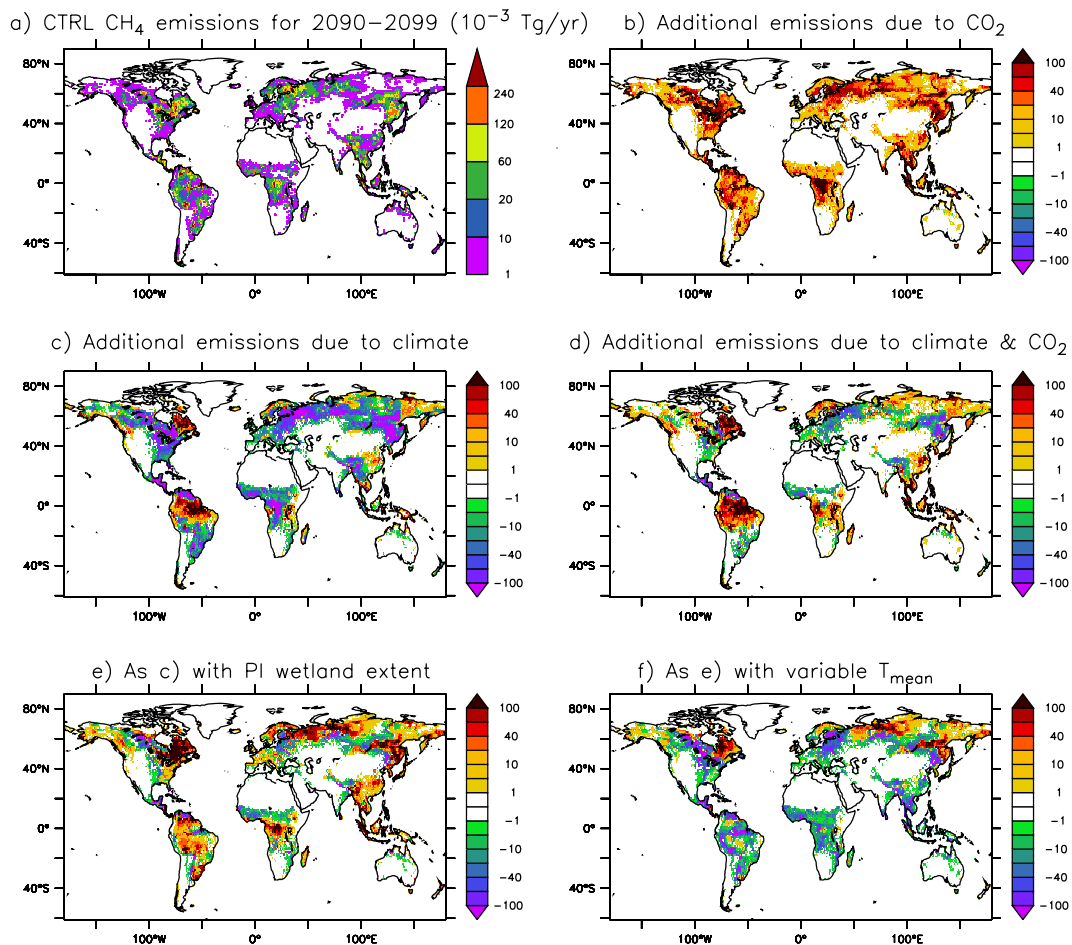


Fig. 3. Mean annual CH₄ emissions by wetlands over 2090–2099 period for CTRL simulation (a) and changes in emissions due to increase in atmospheric CO₂ (b), climate change (c) and both (d). The shown changes in emissions are obtained by ORCHIDEE-WET simulations with $Q_{10} = 3$, constant in time T_{mean} and accounting for wetland extents variation, which is the basic configuration. Climate effect on CH₄ flux densities alone (i.e. without accounting for wetland extent evolution, see above) is given in (e). (f) displays the change in CH₄ flux densities due to climate but obtained with a T_{mean} variable in time.

3.1.3 Response of ORCHIDEE wetlands CH₄ emissions to CO₂, CH₄ and climate

Figure 3a–d shows the mean annual CH₄ emissions by wetlands for the CTRL simulation over the period 2090–2099 (Fig. 3a) as well as the changes in emissions (2090–2099 average relative to the control) due to the change in atmospheric CO₂ (Fig. 3b), climate (Fig. 3c) and both atmospheric CO₂ and climate (Fig. 3d). Changes in CH₄ emissions due to an increase in atmospheric CH₄ are negligible (not shown). The changes in emissions shown in Fig. 3a–d are obtained by ORCHIDEE-WET simulations considering a methanogenesis Q_{10} of 3, a time-constant T_{mean} and accounting for variation in wetland extent, which is considered below as the basic configuration. The climate effect on CH₄ flux densities alone (i.e. without accounting for wetland extent evolution, as above) is given in Fig. 3e. Figure 3f displays the change in

CH₄ flux densities due to climate, as in Fig. 3e, but obtained with a time-varying T_{mean} .

The global averaged pre-industrial wetland CH₄ emission amounts to 253 TgC yr⁻¹ which is, as for present-day, slightly higher than previous estimates (e.g. Chappellaz et al., 1993). Changes in CH₄ emissions due to the various forcing show a large spatial variability. The overall effect of CO₂ and climate (Fig. 3d) is an increase in high latitudes, in the northern half of the Amazon basin, in South-east Asia and in some parts of central Africa. Elsewhere, the emissions decrease under future climate and CO₂. This pattern is a combination of a widespread increase due to CO₂ alone (Fig. 3b) and of a general decrease due to climate change alone (Fig. 3c).

The atmospheric CO₂ concentration and the climate affect CH₄ wetland emissions via two main pathways: one due to changes in wetland areas (resulting from changes in the soil water balance); and one due to changes in CH₄ flux

per unit of wetland area (resulting from changes in methanogenesis rate, in the contribution of each sort of transport, etc.). Production of CH₄ can be affected by changes in the temperature-dependant methanogenesis rate but also by changes in substrate quantity.

Removing the wetland extent evolution leads to reducing the increase of CH₄ emissions under elevated CO₂ (not shown). The mechanism underlying this is that elevated CO₂ reduces the transpiration of plants, and therefore leads to an increase in soil water content given by ORCHIDEE-WET and thus an increase in the wetlands CH₄ emissions via an increase of the wetland areas. Wetland CH₄ flux densities also increase with atmospheric CO₂ increases. As mentioned before, this response can be explained by the fertilization effect. Increased atmospheric CO₂ stimulates plant productivity, which leads to a rise in the active soil carbon pool and hence to more substrate available for methanogenesis. The effect of CO₂ fertilization increasing productivity in ORCHIDEE-WET is similar to one of the four other DGVM models analyzed by Sitch et al. (2008).

Except for the north of South-America, the north and the north-east of Siberia, the west of China as well as the west of Canada the effect of climate change is to reduce CH₄ emissions (Fig. 3c). Removing the wetland area's sensitivity to climate decreases largely the reduction in CH₄ emissions (Fig. 3e). The region of the Amazon river is an exception as climate change leads to an extension in wetlands area. In high latitudes, the emission decrease is primarily driven by a decrease of wetland extension. In spite of extension of the active season and thus of the inundated period in this regions, the climate change would lead to a decrease in the maximum of inundated area that coincides with the period of maximum CH₄ flux density. In some places, the increase in methanogenesis rate, through its temperature dependence seems counterbalanced by a decrease in methanogenesis substrate.

Considering a T_{mean} that changes with climate over time restricts to high latitudes the places where we find an increase of CH₄ flux density due to climate (Fig. 3f). A varying T_{mean} reduces the CH₄ flux density sensitivity to temperature represented by Q_{10} formulation (Eq. 12; last term of right member). Thus, reduction in methanogenesis substrate drives the decrease displayed below 40° N. In high latitudes regions, the increase in active soil depth (switch of $f(T)$ from 0 to 1 for some soil layers) counterbalance the evolution of carbon soil and explain the obtained increase in CH₄ flux densities (Eq. 12; first term of right member).

From these simulations, we can now calculate the CO₂ and CH₄ flux sensitivity terms of Eqs. (7a) and (7b) in order to estimate the climate-CH₄-CO₂ gains (Fig. 4). The Fig. 4a–e displays the integral of changes over 1860–2100 for CO₂ uptake and wetland CH₄ emissions as function of atmospheric CO₂ (Fig. 4a–b), atmospheric CH₄ (Fig. 4c) and climate (Fig. 4d–e) and the slopes of these different curves give the sensitivity terms' values. The results shown in Fig. 4 shows

a linear relation between additional CH₄ and CO₂ flux and atmospheric CH₄, atmospheric CO₂, or climate over the period modeled here is supported by the model. This supports the assumption of a first-order linear relation in the theoretical analysis of Sect. 2. The ORCHIDEE-WET computed sensitivity values in 2100 are summarized in the Table 2 in the same units for CO₂ and CH₄. We note that the global net terrestrial CO₂ flux sensitivity to rising atmospheric CO₂ (β_C) and to climate change (γ_C), depend also on the ocean carbon response. We used ocean sensitivity terms (β_O and γ_O) from Friedlingstein et al. (2006) for the IPSL coupled-climate-carbon model to account for the ocean CO₂ uptake feedbacks. Thus, β_C (respectively γ_C) in the Table 2 is the sum of the land flux sensitivity β_C (resp. γ_C) computed using Fig. 4a (resp. Fig. 4d) and of the ocean flux sensitivity β_O (resp. γ_O).

The individual sensitivity of the land CO₂ flux to atmospheric CO₂ (Fig. 4a) as well as its response to global warming (Fig. 4d) is not discussed here. A comprehensive analysis can be found in Friedlingstein et al. (2003).

Concerning CH₄ emissions from wetlands, as mentioned above, we obtain an increase of emissions when atmospheric CO₂ increases (Fig. 4b). $\beta_{C \rightarrow M}$ amounts to 0.0142 by 2100. ORCHIDEE-WET simulates a negative effect of atmospheric CH₄ on wetland CH₄ emissions, β_M (Fig. 4c). The increase in atmospheric CH₄ leads to a decrease in the concentration gradient between wetland soil and the atmosphere, which drives a decline in the diffusive flux of CH₄ from soil, and thus a larger proportion of the CH₄ created is consumed by methanotrophy within the soils. However, the negative value of the sensitivity of emissions to atmospheric CH₄ is very low ($\beta_M = -0.0040$) as assumed before and is explained by the fact that the CH₄ atmospheric concentrations always remains much lower than the CH₄ concentration in wetland soils.

As mentioned before (Fig. 3c), the simulated overall effect of climate changes is to reduce CH₄ emissions from wetlands (Fig. 4e). We find a sensitivity γ_M of -1.83 GtC K^{-1} . Assuming constant wetland area would change the sign of the sensitivity term, with a γ_M^f of $+1.27 \text{ GtC K}^{-1}$. Hence, the overall climate-driven decline in CH₄ emission from wetlands is mainly driven by a decrease in wetland area. The negative value of γ_M obtained in the case where we consider a varying T_{mean} and do not include the dynamics of wetland area changes ($\gamma_M^f = -0.84 \text{ GtC K}^{-1}$) shows that a reduction of the methanogenesis substrate reinforces the negative effect of climate on emissions driven by wetland extent. If a constant T_{mean} is used, taking a higher methanogenesis Q_{10} value ($Q_{10} = 5.5$) leads to few changes when accounting for wetland extent (+17%; from -1.83 to -1.51) but a large change when not accounting for (more than 3 times; from $+1.27$ to $+5.37$).

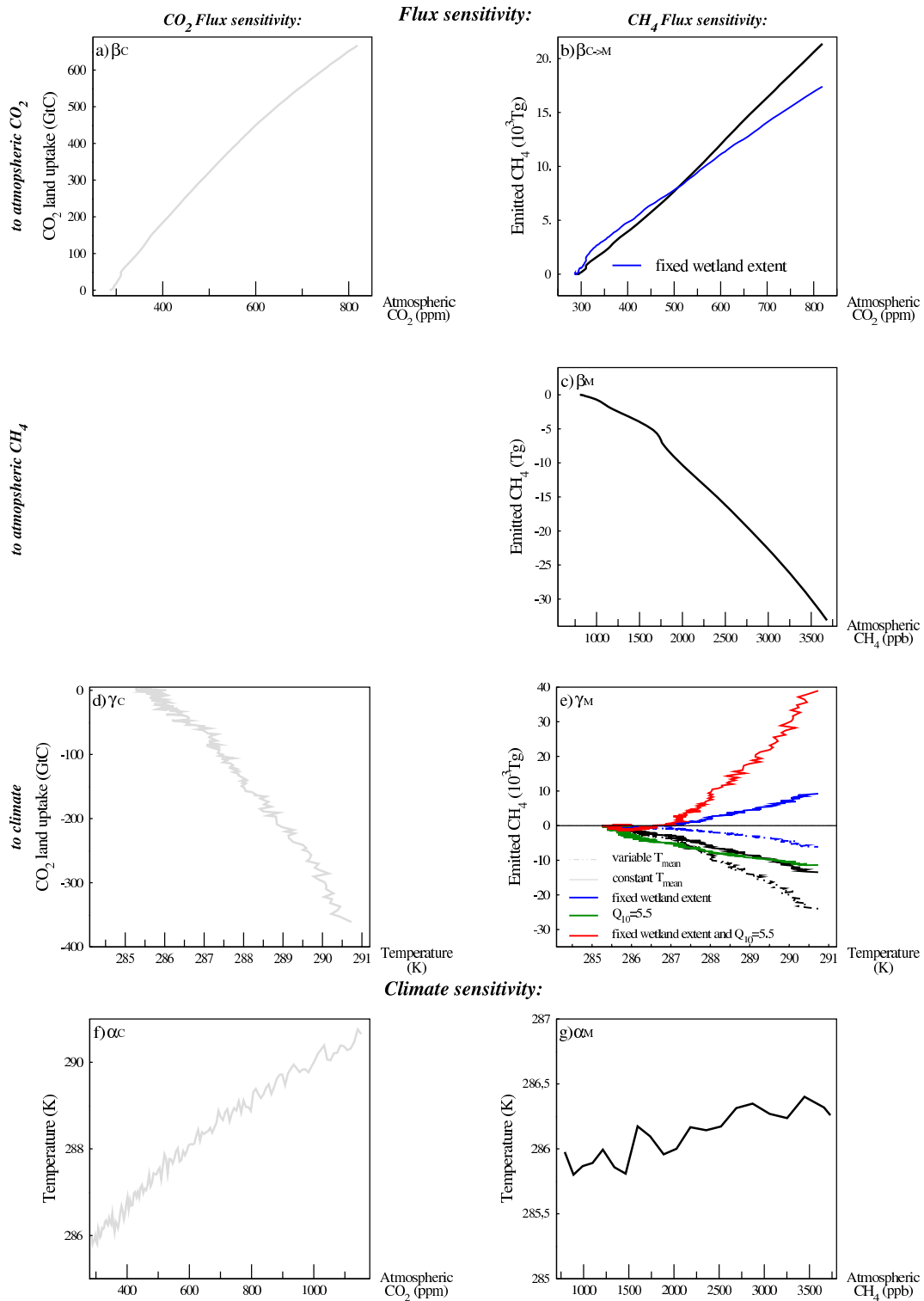


Fig. 4. (a)–(e): Evolution of the integral of change in CO₂ land uptake and CH₄ wetlands emissions as function to atmospheric CO₂ concentration – (a) and (b), atmospheric CH₄ concentration (c) and global air temperature (e). Be careful for the different y-axis unit for (c). Blue curves of (b) and (e) correspond to the evolution of the integral of change in CH₄ wetlands emissions after removing the wetland extent evolution (i.e. using for all the time step the pre-industrial wetland extent). Red curve of (e) is the same as blue one but with a higher Q_{10} for the methanogenesis. (f)–(g): Temperature sensitivity to atmospheric CO₂ and CH₄.

Table 2. Values of the CO₂ and CH₄ flux sensitivity in 2100 (top) as well as climate one (bottom). Given global net terrestrial CO₂ flux sensitivities are sum of ocean sensitivity terms from Friedlingstein et al. (2006) and the estimation of land flux sensitivity based on ORCHIDEE-WET simulations (cf. Fig. 4). Wetland CH₄ emissions sensitivity reported in this table are only based on ORCHIDEE-WET simulations and are also consistent to each other.

		Flux sensitivity in 2100			
		CO ₂ flux		CH ₄ flux	
to atmospheric CO ₂ (unitless)	$\beta_C = 1.11$			With dynamic wetland	$\beta_{C \rightarrow M} = 0.0142$
				Without dynamic wetland	$\beta_{C \rightarrow M}^f = 0.0155$
to atmospheric CH ₄ (unitless)		$\beta_M = -0.0040$			
to climate (in GtC K ⁻¹)	$\gamma_C = -82.3$			Constant	Variable
				T_{mean}	T_{mean}
		$Q_{10} = 3$	With dynamic wetland	$\gamma_M = -1.83$	$\gamma_M = -3.27$
		$Q_{10} = 3$	Without dynamic wetland	$\gamma_M^f = +1.27$	$\gamma_M^f = -0.84$
		$Q_{10} = 5.5$	With dynamic wetland	$\gamma_{M-Q10} = -1.51$	$\gamma_{M-Q10} = -4.85$
		$Q_{10} = 5.5$	Without dynamic wetland	$\gamma_{M-Q10}^f = +5.37$	$\gamma_{M-Q10}^f = -0.17$
Climate sensitivity					
to atmospheric CO ₂ (in K GtC ⁻¹)		$\alpha_C = 0.0029$			
to atmospheric CH ₄ (in K GtC ⁻¹)		$\alpha_M = 0.0840$			

When T_{mean} varies in time, changing the Q_{10} tends to increase γ_M in high latitudes (not shown) due to the activation of some soil layers. At the global scale, this effect is hidden by small changes in contribution of the different latitudes bands to the total emissions from simulation with $Q_{10} = 3$ to simulation with $Q_{10} = 5.5$ (not shown).

3.2 Literature based estimates of $\beta_{C \rightarrow M}$ and γ_M

As mentioned before and despite some remaining uncertainty (e.g. because of interaction with nitrogen cycle), the CO₂ flux sensitivities to climate and atmospheric CO₂ have been already studied and estimated (notably the C⁴MIP intercomparison; Friedlingstein et al., 2006). In the present study, most uncertainty concerns terms relative to the sensitivity of wetlands CH₄ emissions.

A pair of previous studies investigated the future changes in CH₄ emissions from wetlands. Although these studies did not quantify the CH₄ emissions sensitivity to climate and atmospheric CO₂ (respectively γ_M and β_{CM}) one can use their results to derive these quantities.

Neither of the previous studies (SWF04, GCH04, EMA08 and V08) explicitly accounted for changes in CH₄ concentration and its effect on CH₄. To our knowledge there are also no site-level manipulative experiments with increased CH₄ concentration conditions. Therefore we cannot provide

a literature based estimate of β_M . However, we found this term to be negligible (see previous section).

A first-order estimate of γ_M is possible from SWF04, GCH04 and EMA08. Their approach does not account for the fertilizing effect of high CO₂ atmospheric level on wetland CH₄ emissions but only for the effect of the climate change induced by it. Thus, they allow for a direct estimate of γ_M . SWF04 estimate a rise of 78% of wetlands CH₄ emissions (from 156 to 277 Tg yr⁻¹) under a transient 2 × CO₂ climate with a global warming of 3.4 °C. The calculation of γ_M needs the time evolution of the wetland CH₄ emission, as it is the ratio of the cumulated emissions divided by the related warming. Not having this time evolution, we assume here that the growth of CH₄ emission is linear, as the warming is close to linear in such transient 2 × CO₂ climate simulations (e.g. Cubasch et al., 2001). This gives a value of 0.93 GtC K⁻¹ for γ_M . The same estimate of γ_M can be done with GCH04 where they simulate a wetland CH₄ emission increase of 255 Tg yr⁻¹ over 110 years for a warming of 4.2 K (in their reference case, CTRL). This gives a value of 2.70 GtC K⁻¹ for γ_M .

EMA08 find an increase of 130–140 to 170–200 Tg yr⁻¹ under the SRES-A2 scenario (+3.4 °K) but the accounted effects of climate warming on wetland CH₄ emissions are only relative to temperature dependency of methanogenesis and to change in soil depth when permafrost thaws. Both wetland extent and water table depth are constant during their

simulation time and the methanogenesis substrate is constant. Such an increase gives a γ_M value of $\sim 1.9 \text{ GtC K}^{-1}$.

V08 obtained a wetland CH₄ emissions increase of +40% from 20th to the end of 21st century (from 240 Tg yr⁻¹ to 340 Tg yr⁻¹) under a warming of 3.5 °C (scenario A1B). The results cannot be easily delineated into the theoretical framework we have developed in this paper: CO₂ fertilization effect on wetland CH₄ emissions seems to be accounted for (the paper uses NPP to approximate substrate) but is not discussed. Thus we cannot estimate the γ_M and $\beta_{C \rightarrow M}$ effects based on this study.

We also note that the evolution of the global temperature in these previous studies is not the same as one simulated by the ISPL-CM4 OAGCM under SRES-A2. It has been shown for the C-CO₂ feedback that the rate of perturbation has an impact on the estimate of γ_C (Gregory et al., 2009). The same certainly applies for the C-CH₄ feedback and the value of γ_M considered here.

Finally, wetland CH₄ emissions sensitivity to temperature derived through warming manipulation on sites (e.g. Updegraff et al., 2001; White et al., 2008) could not be used to estimate γ_M because this term represents the overall global response of wetlands to climate and not just to temperature.

The published global modeling approaches do not give an estimate of the relationship between atmospheric CO₂ level and global wetlands CH₄ emissions. However, there are wetlands site level manipulative experiment where emissions are measured under ambient and elevated CO₂ (e.g. Dacey et al., 1994; Megonigal and Schlesinger, 1997; Kang et al., 2001; Vann and Megonigal, 2003; Pancotto et al., 2010). The measured response varies between 0% (Pancotto et al., 2010) and 136% under 2 × CO₂ (Megonigal and Schlesinger, 1997) according to the wetland type and the experimental conditions. ORCHIDEE-WET simulated wetland CH₄ emissions increase by +80% (respectively $\sim +50\%$ with no evolution of wetland extent) when atmospheric CO₂ concentration given by SRES-A2 scenario grows from 355 to 716 ppm. The speed of the CO₂ perturbation of the manipulations experiment is totally different from that under the SRES-A2 scenario. Moreover, the relationship between atmospheric CO₂ concentration and wetland CH₄ emissions estimated at sites is extrapolated to global scale with difficulty. Thus we retain only the ORCHIDEE-WET based $\beta_{C \rightarrow M}$ in the following. In Sect. 4, the effect of the different interactions on atmospheric CH₄ and CO₂ will be added successively. The effect without CO₂ fertilization on CH₄ emissions could be seen as lowest boundary of the uncertainty range for accounting for $\beta_{C \rightarrow M}$.

In summary, we find an estimate for β_M , based on ORCHIDEE-WET, of -0.0040 and for $\beta_{C \rightarrow M}$ of 0.0142 . For γ_M , we find, based on both ORCHIDEE-WET and literature based estimates, a range from -1.83 to $+2.70$. The value of γ_M chosen as representative of ORCHIDEE-WET simulation corresponds to the best estimate (i.e. accounting for variation in wetland extent, $Q_{10}=3$ and a constant T_{mean}). ORCHIDEE-WET gives a negative value for γ_M while the

estimates based on GCH04 and SWF04 give a positive γ_M . An analysis on the reasons for the uncertainty on the sign of γ_M will be given in the discussion section.

3.3 Other terms

In order to estimate the gains, we finally have to calculate the climate sensitivity to CO₂ and to CH₄ (α_C and α_M respectively) as well as μ , the atmospheric OH sink scaling term (Eq. 4b).

For α_C , we used the transient global warming of the IPSL-CM4 model from an idealized simulation with changes in atmospheric CO₂ only (CMIP 1 % yr⁻¹) (Fig. 4f). This gives a α_C of $0.0029 \text{ K ppm}^{-1}$. For α_M we have no parallel climate simulation with change in CH₄ concentration only. Thus, we used the standard CO₂ radiative forcing equations (IPCC, 2001, Table 6.2) to derive a climate sensitivity to changes in radiative forcing (ΔRF) from the previous simulation. The same standard radiative forcing equations, but for CH₄, allow us to go from ΔRF to an equivalent of atmospheric CH₄ concentration, if we assume the same climate sensitivity for ΔRF whether it is due to CO₂ or CH₄. We can also estimate the warming due to CH₄ only (Fig. 4f). This gives a α_M of $0.0840 \text{ K ppm}^{-1}$.

We compute μ , the atmospheric OH sink scaling term, as the ratio of the cumulative changes of atmospheric CH₄ along the SRES-A2 scenario to its change at the end of the scenario. We find a μ of 0.322. Anthropogenic emissions of CH₄, F_{MF} , come from the EDGAR database (<http://www.sec.gov/edgar.shtml>) for the historical period and from the SRES-A2 scenario for the 21st century. The CH₄ atmospheric lifetime, τ , is assumed constant with a value of 9 years (IPCC, 2001). This lifetime is sensitive to the atmospheric composition (e.g. COV, see Valdes et al., 2005) and in particular to CH₄ concentration itself (e.g. V08), leading to a feedback. IPCC (2007) gives an adjustment time (Lelieveld et al., 1998) of 12 years to account for indirect effects of increase in CH₄ emissions. Not using of a coupled climate-chemistry model, we cannot account for dependency of τ on CH₄ concentration and on climate.

4 Estimation of the feedbacks' gains and their interaction

4.1 C-CH₄ and C-CO₂ feedbacks' gains

Once the sensitivity terms are estimated, we computed the gains of the C-CO₂ and C-CH₄ feedbacks when each gas is considered alone (Eqs. 7a and b), as well as the interaction between these feedbacks as defined in Eqs. (10) and (11). Combining our range of γ_M (from -1.83 to 2.70 GtC K^{-1} ; see above) we find the C-CH₄ feedback gain, g_M , when CH₄ is considered alone, ranging between -0.016 and 0.024 by 2100, respectively obtained for the best ORCHIDEE-WET estimated γ_M and literature estimated γ_M . The sign

of the gain is controlled by the γ_M . Negative gains are due to the negative wetland emission sensitivity to climate found in ORCHIDEE-WET. For the C-CO₂ feedback gain, using our ORCHIDEE-WET simulations, we find a value of 0.113, slightly higher than the value found in Friedlingstein et al. (2006).

Going back to Eqs. (10) and (11), we can now calculate the cross-gains terms when not accounting for the fertilization interaction. For CH₄ concentration changes (Eq. 10), the g_M gain is augmented by $g_C \cdot g_M / (1 - g_C)$, the additional gain due to the interaction between CO₂ and CH₄. Using a γ_M of -1.83 GtC K^{-1} , this cross-gain is equal to -0.0017 which represents a correction of $\sim 10\%$ of the initial gain g_M . Similarly, for CO₂ (Eq. 11), the cross-gain term $g_C \cdot g_M / (1 - g_M)$ amounts to -0.0015 which represents only $\sim 1.5\%$ of g_C . The CO₂ contribution to CH₄ is larger than the reciprocal because climate has a larger absolute effect on the net CO₂ flux than on the CH₄ emissions from wetlands. If we use the upper estimate of g_M , the cross-gains due to the interactions between C-CO₂ and C-CH₄ feedbacks have similar effects on g_M and g_C (cross-gain $\sim 13\%$ of g_M and $\sim 3\%$ of g_C).

4.2 Effect on atmospheric CO₂, CH₄ and global temperature

We compute the changes of CH₄ in the atmosphere between future (2100) and pre-industrial time, ΔCH_4 , in the case of C-CH₄ feedback alone, and then with cross-feedbacks accounted, for using Eq. (10) (or equations from Appendix B for the most general case). ΔCH_4 and ΔCO_2 are expressed in the following in ppbv and ppmv, respectively. Figure 5a shows the incremental changes in the calculated ΔCH_4 when accounting for successive gains. $\Delta\text{CH}_4^{\text{unc}}$ is the change in CH₄ in the absence of any retroaction, as given by Eq. (6). Then we account successively for the climate-CH₄ feedback (i.e. CH₄ emissions dependence on CH₄ induced temperature change), the *climate interaction* as explained in Sect. 2.2 (i.e. temperature dependence to atmospheric CO₂; $\alpha_C \neq 0$) and the *fertilization interaction* (i.e. CH₄ emissions dependence to atmospheric CO₂; $\beta_{C \rightarrow M} \neq 0$). Figure 5b shows the same calculation, but for ΔCO_2 . For each gas, we plotted both the case with (solid line) and without (dashed line) anthropogenic emissions of the other gas. All calculations were done with the ORCHIDEE-WET based estimated $\beta_{C \rightarrow M}$. However, given the high uncertainty on γ_M , we plotted also the case with the positive γ_M derived from literature (in grey) or with the negative γ_M based on the best ORCHIDEE-WET estimation (in green). We add also the case where T_{mean} is variable in time (in blue). As a first check on our framework, we compared the uncoupled estimates of $\Delta\text{CH}_4^{\text{unc}}$ and $\Delta\text{CO}_2^{\text{unc}}$ to the values given by the SRES-A2 scenario, where none of the feedbacks presented here were accounted for. We find a CH₄ concentration increase by 2100 of 3030 ppbv and a CO₂ concentration change of 496 ppm, not far from the SRES-A2 concentration changes (2925 ppbv for CH₄ and

5450 ppm for CO₂; IPCC, 2001). This indicates that, for CH₄, the assumption of a constant lifetime and the use of the scaling parameter μ are appropriate. We note that the CO₂ concentrations given in IPCC (2001) already accounts for a C-CO₂ feedback.

Concerning CH₄ (Fig. 5a), accounting for the different feedbacks does not have a large effect on the calculated CH₄ concentration as long as anthropogenic CO₂ emissions are neglected (dashed lines). This is because the climate effect of CH₄ anthropogenic emissions alone is too weak to generate a non-negligible C-CH₄ or C-CO₂ feedback. Only the CO₂ emissions induced climate change leads to a large effect on CH₄ emissions and CO₂ sinks and hence modify the calculated CH₄ concentration. This is clearly different for the CO₂ (Fig. 5b) for which its own feedback with the climate explains most of the ΔCO_2 (at least 80 % of the ΔCO_2 given with all interactions).

When anthropogenic CO₂ emissions are included, the large change in atmospheric CO₂ affects CH₄ emissions through (i) the climate effect on CH₄ emissions and (ii) the fertilization effect on substrate. Figure 5a shows that these two terms are important. Depending on the sign of γ_M , the CO₂ induced climate effect will enhance (grey line) by 457 ppbv or reduce (green line) by 310 ppbv the calculated CH₄ concentration. This *climate interaction* includes temperature change due modification of ΔCO_2 caused by both anthropogenic emissions and C-CO₂ feedback. The CO₂ fertilization effect is always positive and, depending on the previous value of γ_M , further increase (red line) or compensate (blue line) the CO₂ induced climate effect.

When accounting for all interactions between CO₂ and CH₄, ΔCH_4 is 1400 ppbv larger than the uncoupled $\Delta\text{CH}_4^{\text{unc}}$ in the case of positive γ_M and 475 ppbv larger in the negative case (with constant T_{mean}). Variable T_{mean} leads to an increase of only 190 ppbv.

For CO₂ (Fig. 5b), as mentioned above, most of the change in the calculated concentration comes from the C-CO₂ gain, where ΔCO_2 rise from 495 to 560 ppm. Accounting for the interactions with CH₄ slightly changes this value and can lead to an increase of 15 ppm. The climate change induced by anthropogenic CH₄ emissions is preponderant to the C-CO₂-CH₄ effect on the ΔCO_2 (comparison between dash and plain lines) while the C-CH₄ feedback induced climate change has only a small effect. In all the cases, the increase in wetland CH₄ emissions induced by the *fertilization interaction* has a little effect on ΔCO_2 (~ 3 ppm).

The change in global temperature, ΔT that would follow these different changes in atmospheric CO₂ and CH₄ can be estimated using Eq. (2b). Several combinations of ΔCH_4 and ΔCO_2 are possible, according to the interactions that are accounted for. Here, we limit the ΔT computation to specific cases. In the absence of anthropogenic CH₄ emissions, accounting for C-CO₂ feedbacks leads ΔT to rise from 3.05 to 3.44 K. Accounting for anthropogenic CH₄ emissions in addition leads ΔT to rise to 3.98 K. In the case where γ_M

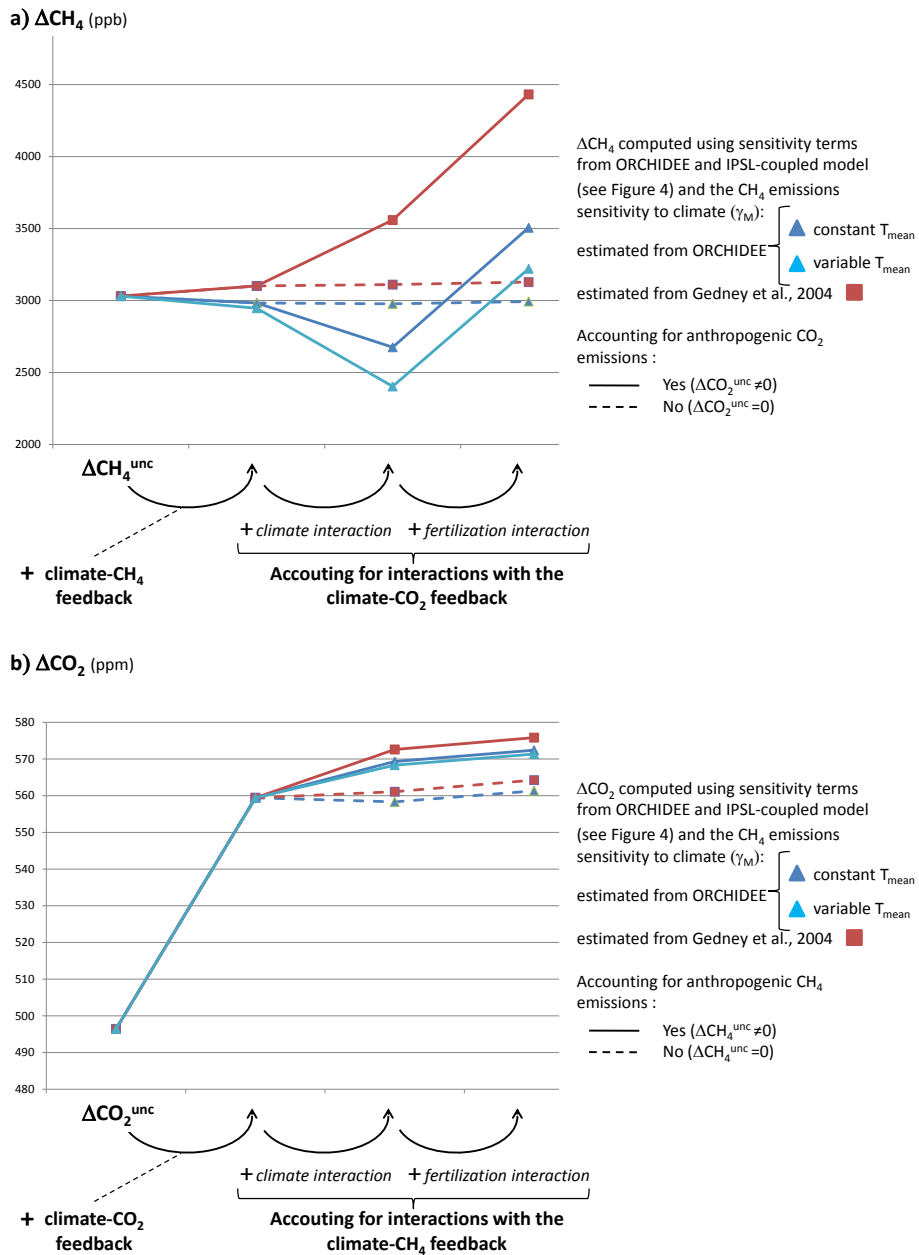


Fig. 5. Evolution of the difference in CH₄ (a) and CO₂ (b) atmospheric concentration between future and pre-industrial time accounting or not for feedback with climate and interactions with the other feedback. “Climate interaction” and “Fertilization interaction” report to definitions into Sect. 2.2 of the manuscript. To reminder, concerning CH₄ (respectively CO₂), “climate interaction” means that climate is also a function of CO₂ (respectively CH₄). Accounting for “fertilization interaction” consists in accounting for wetland CH₄ emissions dependence to atmospheric CO₂.

is negative, adding the C-CH₄ feedback and its interactions with C-CO₂ leads to a ΔT of 4.14 K. The same addition with positive γ_M leads to a ΔT of 4.33 K. The estimate of ΔT when accounting for the C-CH₄ feedback and its interaction with CO₂ is non-negligible (in comparison with the warming directly due to anthropogenic CH₄ emissions) but strongly depends on the sensitivity of wetland CH₄ emissions to climate.

5 Discussion

In the above calculations, the highest uncertainty comes from the sensitivity of wetland CH₄ emissions to climate (γ_M). Contrary to SWF04, GCH04 and EMA08, we find a negative value for γ_M . In our ORCHIDEE-WET based estimation, the climate-driven change in wetland extent plays a large role in the overall emission reduction. As explained

above, EMA08 did not account for change in wetland extent. GCH04 obtained also a weak wetland reduction, while SWF04 simulates a small increase in wetland extent. However, as both GCH04 and our study use a mechanistic approach (TOPMODEL) as opposed to the empirical approach used by SWF04, we have a higher confidence in a reduction of wetland surfaces. Regardless, large uncertainty remains on the representation of wetland extent (e.g. Bohn and Lettenmaier, 2011; Ringeval et al., 2011). Despite a decrease in wetland extension, GCH04 shows an increase in overall emissions. That is to say their emission rate increases and compensates for the reduction of emitting surface, a feature we do not find with ORCHIDEE-WET.

GCH04 tested a large range of methanogenesis Q_{10} values with an upper range higher than the standard value we used in this study. Even if there is a clear evidence that methanogenesis rates increase with temperature (e.g. Conrad, 1989), much uncertainty about the Q_{10} value remains (Valentine et al., 1994). To investigate the role played by this parameter and to test the case in which a higher Q_{10} can counterbalance the decrease in wetland extent, we performed an additional simulation with a Q_{10} of 5.5 in accordance with greatest value of GCH04 (see Sect. 3.2). In the case where T_{mean} is variable, we obtain a more negative value for γ_M . An indirect effect of a Q_{10} increase is a little change in the preindustrial latitudinal distribution of wetland emissions (not shown). When T_{mean} is considered as a variable, changing the Q_{10} has a very little impact on the methanogenesis rate which cannot counterbalance the effect of change in latitudinal emissions. Thus, the γ_M is more negative (-4.85) in this case. In the case where T_{mean} is constant (a configuration more directly comparable to GCH04), we find a smaller but still negative value for γ_M at the global scale, contrary to GCH04.

In ORCHIDEE-WET, the evolution of CH₄ flux density is explained by a balance between an increase of methanogenesis rate due to its temperature dependence and a decrease of substrate. In our results, with the exception of some locations where the increase due to temperature dependence seems to be the predominant factor, the decrease of substrate contributes to limit this increase. Neither GCH04 nor SWF04 and EMA08 account for a climate-induced change in substrate, they only account for a methanogenesis dependence on temperature. For the sake of comparison, we calculate the change in wetland CH₄ emissions which ORCHIDEE would simulate by 2100 if soil carbon pools were unchanged (i.e. held at the initial pre-industrial value). We estimate roughly this for each simulation by multiplying the CH₄ flux densities at each year and at each grid-cell by the ratio of pre-industrial active soil carbon to the stock of the simulation considered. The different effects are summarized in Table 3. In the $Q_{10} = 5.5$ case, removing the variability in soil carbon switches the climate effect on the difference between wetland CH₄ emissions in 2100 and pre-industrial time from a decrease of 10 % to an increase of 84 %. Hence, we find that

Table 3. CO₂ alone, climate alone and combined effect on difference in global wetland CH₄ emissions between 2099–2090 and 1860–1869. Results are done for simulations with a constant T_{mean} . CH₄ emissions with pre-industrial soil carbon (last line of the table) correspond to a sensitivity test done a posteriori (cf. discussion). Results are done in percent of global pre-industrial emissions. In our study, CO₂ effect and climate alone are considered independent and are derived from only 3 simulations (cf. Table 1) thus last row of the Table 3 is the sum of the two previous rows.

CH ₄ Flux densities		Wetland extent	CO ₂ effect	Climate effect	CO ₂ + Climate effect
Q_{10}	Soil carbon				
3	T	T	+134 %	−64	+69
3	T	PI	+83	+64	+148
3	PI	T	+34	+25	+59
3	PI	PI	+3	+134	+137
5.5	T	T	+138	−10	+128
5.5	T	PI	+85	+209	+294
5.5	PI	T	+33	+84	+117
5.5	PI	PI	+1	+274	+275

PI: Pre-industrial; T: Transient over 1860–2100

if we assume no change in substrate, we also find a positive γ_M as in GCH04.

Thus, a crucial question for understanding changes to the CH₄ flux density is whether methanogenesis substrate will change in the future as a response to global warming. In ORCHIDEE-WET, we account for this change, which we model as the active soil carbon pool, whereas GCH04 and SWF04 do not account for any change.

The ORCHIDEE modeled reduction of active soil carbon pools by future warming is driven by a change in inputs (NPP) and outputs (CO₂ heterotrophic respiration). The active soil carbon used as a CH₄ production substrate is the total active carbon stock of all natural plant functional types in each grid-cell. The current parameterization may not capture realistically productivity and decomposition processes in northern wetlands soil (Ise et al., 2008; Bridgman et al., 2006). In particular, in some regions, NPP decreases under future climate change because of a decrease of plant water availability. This might not be realistic for the water-saturated fraction of such grid-cell.

Regarding heterotrophic respiration, in a wetland, the rate of soil organic carbon decomposition is lower due to anoxic conditions. In fact, data from permanently inundated sites shows a slow-down of decomposition processes (Freeman et al., 2002) yielding to carbon accumulation in the soil, i.e. peat growth (Clymo et al., 1998). In those wetlands which are saturated throughout the year, the direct respiration of soil carbon into CO₂ is thus strongly inhibited. Despite this inhibition, the decomposed carbon in a wetland is mainly turned into CO₂ and not to CH₄. The observed range of CH₄/CO₂ ratios in anaerobic conditions is large (from 0.0001 to 1.7, see e.g. Wania et al., 2010; Updegraff et al., 2001;

Rinne et al., 2007). Neither the inhibition of decomposition into CO₂ nor the effect of CH₄ decomposition on soil carbon pool is accounted for in ORCHIDEE-WET.

Lastly, CO₂ heterotrophic respiration in our model does increase with temperature in the model ($Q_{10}=2$). Again, we might overestimate this effect over wetlands, as one would expect the wetland soils decomposers to be less responsive to temperature because of the anoxic conditions. Flooding contributes to decrease the apparent temperature sensitivity of decomposition (Davidson and Janssens, 2006). Nevertheless, it seems unlikely that the methanogenesis substrate in wetland soils would not respond at all to climate as in GCH04, SWF04 and EMA08.

Our results point to the necessity of being able to accurately simulate the changes in methanogenesis as well as its available substrate as it was suggested by Kaplan (2002) over other time periods. Having wetland-specific plant functional types with their own productivity and soil decomposition parameters as done by Wania et al. (2009) seems necessary. It enables the model to also account for changes in wetland vegetation composition under future climate change (Ström et al., 2003). Regardless, large uncertainties remain on how to represent the methanogenesis substrate in global models, mainly because of the challenge of upscaling local information on, for example, substrate available for methanogenesis (Limpens et al., 2008; Zona et al., 2009) to large scale quantities such as productivity or soil carbon active pool production (Christensen et al., 2003).

The difference of results obtained for simulations performed with constant or variable T_{mean} underlines the effect of the uncertainty in the micro-organisms response to change in environment on global wetland CH₄ emissions. As underlined by lots of studies (e.g. Rainey and Travisano, 1998), microorganisms are likely to adapt to changing conditions. This adaptation could be done either by mutation or by change in communities, and implies that the microbial community is already highly optimized for a given site and can thus benefit (or suffer) less from changing climate than if adaptation is not accounted for. But we found that accounting or not for this adaptation will not change the sign of the climate effect on wetland CH₄ emissions as much as the variation in wetland extent accounts for.

Another uncertainty relates to the wetland CH₄ emissions' sensitivity to atmospheric CO₂ ($\beta_{C \rightarrow M}$). Large increases in CH₄ emissions in response to elevated CO₂ can occur in a wide variety of wetland ecosystems (Vann and Megonigal, 2003). One hypothesis to explain this increase in CH₄ emissions is the rise of photosynthates that become available for fermentation through root exudation or rapid root turnover via enhanced photosynthesis (Dacey et al., 1994; Vann and Megonigal, 2003). In herbaceous dominated wetland, another possibility is the increase of the plant-mediated transport via an increase in plant biomass and thus in tiller number/stem weight at maturity (Vann and Megonigal, 2003). In our modelling approach, we represent increase in substrate

for methanogenesis only through variation in active carbon pool. We do not represent roots exudates in ORCHIDEE. The increase in plant transport is indirectly accounted for via the CO₂-induced increase in LAI modelling by ORCHIDEE. Further investigation is needed to quantify the contribution of each of these processes to the response to CO₂ we simulate and how they agree with observations. Moreover, experimentation had underlined many uncertainties linked to interactions with nutrient cycle, which are not accounted currently for in the model. Indeed, direct fertilization effects of CO₂ could be balanced by their effects on the substrate quality and thus on the decomposition rate (Pancotto et al., 2010). Other uncertainties are linked to the change in wetland plant physiology (e.g. modification of the turnover rate under high CO₂ level; Megonigal and Schelsinger, 1997) or to modification of oxidation rate in the case of change in transport by plant (higher supply of oxygen transport into the rhizosphere; Kang et al., 2001).

Another limitation of our approach is that we did not account for permafrost carbon decomposition and associated CO₂ and CH₄ emissions. The release of CH₄ by decomposition of thawed deep soil carbon under increase of active layer (Khvorostyanov et al., 2008) could dramatically increase to the CH₄ emissions sensitivity to climate.

6 Conclusions

In this study, we have generalized the theoretical analysis of Friedlingstein et al. (2003) for the interplay between the climate-carbon cycle and the climate-CH₄ feedback. These two feedbacks are not independent, instead they interact through two processes. The first is that a warming due to a CO₂ release would have an effect on wetland CH₄ emissions through changes in available substrate, methanogenesis rate, and the extent of wetland areas. Reciprocally, a CH₄-induced warming would affect carbon storage and hence atmospheric CO₂. The second is that increased atmospheric CO₂ would increase the amount of available organic substrate for methanogenesis (via enhancement of plant photosynthesis), in the absence of other limitations or dynamic vegetation responses, and modify the plant-mediated transport intensity and hence increase CH₄ emissions from wetlands. Our theoretical approach makes it possible to express the additional gains arising from these interactions and to quantify the effect on atmospheric CH₄ and CO₂ concentrations. High uncertainty remains, even for the sign and amplitude of the C-CH₄ feedback gain essentially due to the lack of knowledge about wetland extent evolution as well as the representation of wetland soil carbon dynamics in global models. Nevertheless, we find that, when each gas is considered alone, the gain of the C-CH₄ feedback (−0.016 to 0.024) is much lower than the C-CO₂ feedback gain (~0.113). Concerning the interaction between the two feedbacks, because of the much larger radiative forcing associated with CO₂ than

CH₄ (in the scenario used here), the cross feedback effects are only significant on atmospheric CH₄ concentration when anthropogenic CO₂ emissions are included. The different interactions between the two feedbacks can offset or add up, based on the sign of the C-CH₄ feedback gain. Thus, ΔCH₄ could be increased between 475 and 1400 ppb due to feedbacks, with various effects on ΔT.

Still, large uncertainties remain in the C-CH₄ feedback gain. These mainly arise from (i) the sensitivity of both the wetland extent and methanogenesis substrate to climate as well as (ii) the CO₂ fertilizing effect on the wetland CH₄ emissions. Our results suggest that, in particular, the representation of methanogenesis substrate and its specific local-scale response to the larger scale global change is an area that deserves further development.

Appendix A

Getting of Eq. (9) using equations relative to C-CO₂ feedback

The only one modification of the following equations as compared as Friedlingstein et al. (2003) is the CH₄-dependance of ΔT. In the following equations, F_{CF} is the integral of anthropogenic CO₂ emissions and F_{C^{add}}(GtC) is the integral of the change in natural net fluxes between surface and atmosphere. We had brought together ocean and continental surface.

$$\begin{cases} \Delta\text{CO}_2 = F_{CF} - F_C^{\text{add}} \\ F_C^{\text{add}} = \beta_C \Delta\text{CO}_2 + \gamma_C \Delta T \\ \Delta T = \alpha_C \Delta\text{CO}_2 + \alpha_M \Delta\text{CH}_4 \end{cases} \quad (\text{A1})$$

$$\Rightarrow (1 + \beta_C + \gamma_C \alpha_C) \Delta\text{CO}_2 + \gamma_C \alpha_M \Delta\text{CH}_4 = F_{CF}$$

Appendix B

Equations in the most general feedback calculation framework (i.e. with fertilization interaction)

$$\begin{aligned} \Delta\text{CH}_4^{\text{COU}} & \quad (\text{B1}) \\ &= \frac{1}{1 - \left[g_M - \frac{\beta_{C \rightarrow M} \alpha_M \gamma_C}{(1 + \beta_C + \gamma_C \alpha_C) (1 + \frac{\mu}{\tau} \Delta t - \beta_M)} + \frac{g_C g_M}{1 - g_C} \right]} \Delta\text{CH}_4^{\text{UNC}} \\ &+ \left(\frac{\beta_{C \rightarrow M}}{(1 + \frac{\mu}{\tau} \Delta t - \beta_M)} + \frac{\alpha_C}{\alpha_M} g_M \right) \cdot \frac{1}{(1 - g_C)} \\ &\cdot \frac{1}{1 - \left[g_M - \frac{\beta_{C \rightarrow M} \alpha_M \gamma_C}{(1 + \beta_C + \gamma_C \alpha_C) (1 + \frac{\mu}{\tau} \Delta t - \beta_M)} + \frac{g_C g_M}{1 - g_C} \right]} \Delta\text{CO}_2^{\text{UNC}} \end{aligned}$$

$$\begin{aligned} \Delta\text{CO}_2^{\text{COU}} & \quad (\text{B2}) \\ &= \frac{1}{1 - \left[g_C + \frac{\beta_{C \rightarrow M} \alpha_M}{\alpha_C (1 + \frac{\mu}{\tau} \Delta t - \beta_M - \gamma_M \alpha_M)} g_C + \frac{g_C g_M}{1 - g_M} \right]} \Delta\text{CO}_2^{\text{UNC}} \\ &+ \frac{\alpha_M}{\alpha_C} \cdot \frac{g_C}{1 - g_M} \\ &\cdot \frac{1}{1 - \left[g_C + \frac{\beta_{C \rightarrow M} \alpha_M}{\alpha_C (1 + \frac{\mu}{\tau} \Delta t - \beta_M - \gamma_M \alpha_M)} g_C + \frac{g_C g_M}{1 - g_M} \right]} \Delta\text{CH}_4^{\text{UNC}} \end{aligned}$$

Appendix C

Details about the way to compute wetland extent dynamic

In ORCHIDEE-WET, wetland area dynamics were computed using the TOPMODEL (Beven and Kirkby, 1979) approach of Decharme and Douville (2006). For each grid-cell, using both topographic heterogeneities and soil moisture computed by ORCHIDEE-WET, the TOPMODEL subroutine computes a sub-grid saturated fraction. The simulated space-time distribution of saturated soils was evaluated globally against data from a suite of satellite observations from multiple sensors (Prigent et al., 2001, 2007) interpolated at 1° resolution. Multi-satellite data gives information about inundated fraction (i.e. water-logged soil) whereas our ORCHIDEE-WET model gives the saturated fraction. Thus, the two variables are not comparable in absolute value; saturated area being not necessary free-water surface/stagnant water-logged (over 1993–2000, mean Prigent et al. (2007) areas = 2.8 % of global surface, mean ORCHIDEE-WET areas = 11.2 %). Moreover, absolute values of Prigent et al. (2007) data is prone to some uncertainties: multi-satellite approach has difficulty to catch small, isolated water patches in areas with large dry fraction; as well as small dry patch in areas with large wet fraction. The product could be also affected by ocean contamination on the coast. That is why we focus our evaluation step only comparing normalized variability of Prigent et al. (2007) data and one of ORCHIDEE-WET areas. We show in Fig. C1 comparison between year-to-year variability for these two distributions for the period 1993–2000 for three large regions. More details about this evaluation will be found in Ringeval et al. (2010, 2011).

Saturated areas as simulated by ORCHIDEE-WET do not correspond necessarily to water-logged soil and emitting wetland areas. Gedney and Cox (2003) used also a TOPMODEL approach to diagnose wetland area and raised the same issue. They introduced a global scaling factor in order to simulate a global wetland extent in agreement with observations taken from Aselmann and Crutzen (1989). Here, we opted for a different method and used a climatology (1993–2000) constructed from the Prigent et al. (2007) dataset as a

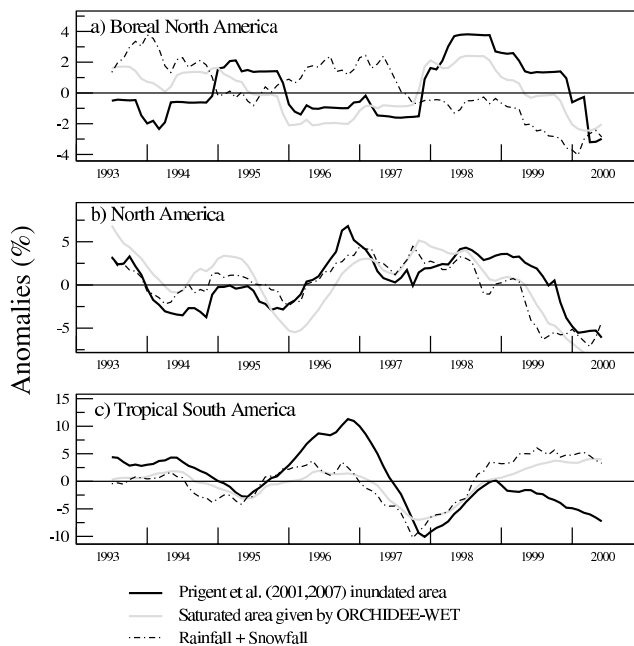


Fig. C1. Comparison between year-to-year variability for inundated area of Prigent et al. (2007) (black curve) and saturated area simulated by ORCHIDEE-WET (grey curve) over 1993–2000 for three regions for instance (Boreal North America, North America and Tropical South America). Regions definitions come from the TRANSCOM atmospheric inversions intercomparison project (Gurney et al., 2003). Anomalies of precipitations from Sheffield et al. (2006) climate forcing data are also added (dashed curve).

baseline for our present day estimate. Future simulated wetland extent was then calculated from the ORCHIDEE-WET simulations, corrected to subtract the systematic biases between the present day simulated saturated area and observed wetland distributions. Only the mean climatology (average of 1993–2000) of the modelled wetland extent is normalized to the same climatology from Prigent et al. (2007) data. As underlined by sensitivity tests (not shown), the way in which the bias is removed (absolute or relative anomalies; Francois et al., 1998) has no influence on the role played by wetland extent in the following (see Sect. 3.1.3).

Appendix D

Incorporation of Walter et al. (2001) model into ORCHIDEE

As mentioned in Sect. 3.1.1, when including the Walter et al. (2001) CH₄ emission model in ORCHIDEE, we made the same following modification as in Ringeval et al. (2010): substrate for methanogenesis is computed from active soil organic carbon computed by ORCHIDEE rather using linear regression versus soil temperature and Net Primary

Productivity (NPP) based on 6 sites as it was done by Walter et al. (2001). Contrary to Ringeval et al. (2010), ORCHIDEE-WET was run here not assuming absence of water stress for vegetation. Thus ORCHIDEE carbon stocks are different in the two studies and new optimization for the base rate of methanogenesis parameter is necessary (α_0 in Eq. 2 of Ringeval et al., 2010). To do so, same approach as in Ringeval et al. (2010) is used: simulated CH₄ fluxes with ORCHIDEE-WET were performed on 3 sites (Abisko, Jackowicz-Korczynski et al., 2010; Michigan, Shannon and White, 1994 and Panama, Keller, 1990) using the monthly NCEP climate forcing data corrected by CRU (n. Viovy et al., personal communication, 2009, <http://dods.extra.cea.fr/data/p529viovy/cruncep/readme.htm>) and compared to the site level observations. Optimized values are respectively: 3.51 , 2.63 and $15.77 \times 10^{-6} \text{ m}^{-1} \text{ month}^{-1}$. Only three sites are chosen because we did not optimize the metanotrophy-related parameters and hence we restrict our calibration only to the flooded sites (i.e. sites where the water table depth reaches soil surface) and flooded period. Contrary to Ringeval et al. (2010), identification of each grid-cell to a wetland type (i.e. sharing the same optimized parameter as Abisko, Michigan or Panama) is not based on latitudinal criteria yet but on a criteria of vegetation type.

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