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Significant feedbacks of wetland methane release on climate change and the causes of their uncertainty

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Emissions from wetlands are the single largest source of the atmospheric greenhouse gas (GHG) methane (CH₄). This may increase in a warming climate, leading to a positive feedback on climate change. For the first time, we extend interactive wetland CH₄ emissions schemes to include the recently quantified, significant process of CH₄ transfer through tropical trees. We constrain the parameterisations using a multi-site flux study, and biogeochemical and inversion models. This provides an estimate and uncertainty range in contemporary, large-scale wetland emissions and their response to temperature. To assess the potential for future wetland CH₄ emissions to feedback on climate, the schemes are forced with simulated climate change using a ‘pattern-scaling’ system, which links altered atmospheric radiative forcing to meteorology changes. We perform multiple simulations emulating 34 Earth System Models over different anthropogenic GHG emissions scenarios (RCPs). We provide a detailed assessment of the causes of uncertainty in predicting wetland CH₄–climate feedback. Despite the constraints applied, uncertainty from wetland CH₄ emission modelling is greater than from projected climate spread (under a given RCP). Limited knowledge of contemporary global wetland emissions restricts model calibration, producing the largest individual cause of wetland parameterisation uncertainty. Wetland feedback causes an additional temperature increase between 0.6% and 5.5% over the 21st century, with a feedback on climate ranging from 0.01 to 0.11 Wm⁻² K⁻¹. Wetland CH₄ emissions amplify atmospheric CH₄ increases by up to a further possible 25.4% in one simulation, and reduce remaining allowed anthropogenic emissions to maintain the RCP2.6 temperature threshold by 8.0% on average.

Introduction

CH₄ is the second largest contributor to the anthropogenic greenhouse gas (GHG) effect after carbon dioxide (CO₂), contributing about 21% of the increased radiative forcing since pre-industrial times (Myhre *et al* 2013). Over a 100 year time period CH₄ global warming potential is about 28 times that of CO₂ (Myhre *et al* 2013). Surface atmospheric CH₄ concentrations reached 1810 ppb in 2012 (Saunio *et al* 2016), which is about 2.5 times as large as pre-industrial concentrations. In the late 1990s and early 2000s there

was a near decade of minimal growth in atmospheric CH₄, which has recently been followed by a renewed and sustained period of increase (Nisbet *et al* 2014).

Natural wetland emissions currently contribute up to approximately 40% of the global CH₄ emissions total (Saunio *et al* 2016), and are thought to cause much of the yearly atmospheric CH₄ concentration variability (McNorton *et al* 2016). These emissions are expected to produce a positive feedback on climate change (Gedney *et al* 2004, Arneth *et al* 2010, Ciais *et al* 2013, Melton *et al* 2013, Stocker *et al* 2013) but are assessed to have a ‘very low’ scientific confidence (Arneth *et al* 2010).

The Sauniois *et al* (2016) review shows that present-day wetland CH₄ global emissions remain highly uncertain, though there is some consistency between estimates using top-down (atmospheric inversions which incorporate observations into dynamic and chemistry models) and bottom-up (process-based land models) approaches, with ranges of 127–202 TgCH₄yr⁻¹ and 153–227 TgCH₄yr⁻¹, respectively. The range in bottom-up estimates of Sauniois *et al* (2016) is smaller than reported in similar earlier studies (Kirschke *et al* 2013, Melton *et al* 2013), which is mainly due to the use of a common wetland area map across models.

Amazon basin observations have recently found that tropical trees act as significant conduits for wetland CH₄ emissions (Pangala *et al* 2017), transferring about half of the total wetland flux here. These measurements have resolved the previously large discrepancy between CH₄ flux estimates from scaled-up flux measurements and top-down approaches over this large regional source (Pangala *et al* 2017). However, lack of detailed knowledge of present-day wetland extent remains a large contributor to current emissions uncertainty in process-based models, as it is often used for scaling up (Kirschke *et al* 2013, Sauniois *et al* 2016). Uncertainties in atmospheric chemistry and transport, and non-wetland CH₄ budgets contribute significantly to inversion model estimate uncertainty (Sauniois *et al* 2016).

From a limited number of studies over different climate change scenarios, global wetland CH₄ emissions are projected to increase by up to 78% (with simulated climate change between present day and doubling of atmospheric CO₂), and enhanced future feedback raises radiative forcing by up to 0.1 Wm⁻² K⁻¹ (Arneeth *et al* 2010, Ciais *et al* 2013). To extend this we take a more rigorous approach in assessing CH₄ wetland feedback under climate change. We consider separate parameterisations of the two primary wetland CH₄ generation processes, include CH₄ transfer through trees and a new representation of tropical soils, and adopt better constrained parameters. We then apply this in a framework of multiple emissions scenarios and 34 different Earth System Models (ESM) climate change estimates.

JULES wetland model

CH₄ consumption (methanotrophy) in the non-saturated zone above the water table is far more efficient than methanogenesis occurring within the saturated zone (Roulet *et al* 1992). Hence we assume that CH₄ produced in the soil saturated zone can only reach the atmosphere directly when the water table reaches the soil surface (incorporated via the grid-box wetland fraction; Gedney *et al* 2004). CH₄ in the soil saturated zone may still be transferred to the atmosphere via vegetation when the water table is lower

however. With emissions being so dependent on water table depth, it is necessary to simulate hydrology and wetland extent adequately.

Large-scale surface models traditionally use soil hydraulic properties based on sand, silt and clay percentages (Best *et al* 2011). In order to improve the hydrological depiction in the tropics, which is a major contributor to the global wetland CH₄ budget, we extend our soil properties to include tropical soils (feralsols, which are weathered soils with micro-aggregated particles), as these are poorly represented by standard particle distribution functions (Marthews *et al* 2014).

JULES (Clark *et al* 2011) simulates the wetland fraction fw by combining simulated grid-box mean water table depth and sub-grid topographic distribution to produce a sub-grid water table depth distribution $fw'(z)$ (supplementary appendix SA is available online at stacks.iop.org/ERL/14/084027/mmedia). fw is defined as the grid-box fraction where the water table is at or just above the surface z_{wet} , i.e. $fw = \int_0^{z_{wet}} fw'(z) dz$ (Gedney *et al* 2004, Gedney and Cox 2003). Thus fw can be compared against observation-based inundation products.

To incorporate CH₄ transfer through vegetation, including the significant flux via tropical trees (Pangala *et al* 2017), we extend the standard modelled fw to include an *effective* wetland extent fw_e . We assume this flux through vegetation can occur where roots exist below the water table. The flux originating at soil depth z is therefore proportional to the relative occurrence of water table depth at z within the grid box $fw'(z)$, and the fractional number of roots below z , $P(z)$. Root distribution decays exponentially up to total soil depth z_{soil} , according to root depth z_r :

$$P(z) = \frac{\int_{z_{soil}}^z e^{-\frac{z}{z_r}} dz}{z_r(1 - e^{-\frac{z_{soil}}{z_r}})} \quad (1)$$

Integrating $P(z)fw'(z)$ gives:

$$\int_{z_{soil}}^{z_{wet}} P(z)fw'(z) dz = \int_{z_{soil}}^0 P(z)fw'(z) dz + fw$$

as $\int_{z_{soil}}^0 P_i(z) dz = 1$ by definition. Summing over vegetation fractions fv_i , we obtain an *effective* vegetation wetland fraction fw_e , which encompasses vegetation transfer:

$$fw_e = \sum_i fv_i A \left[\int_{z_{soil}}^0 P_i(z)fw'(z) dz + fw \right]$$

Here i represents each vegetation type. Efficiency factor A relates to how effectively CH₄ is removed from the soil via vegetation. Over vegetation when the water table is at or above the surface, we assume that any soil CH₄ not lost through vegetation is emitted directly into the atmosphere. Hence the effective wetland fraction for soil is:

$$fwe_s = \sum_i f v_i \cdot (1 - A) f w + \left(1 - \sum_i f v_i \right) \cdot f w.$$

Due to lack of detailed, available measurements, such as how vegetation flux varies with water table depth, A can only be calculated by ensuring the total fraction of CH_4 emitted via tropical Amazon trees agrees with that observed (~ 0.5 ; Pangala *et al* 2017). Since Pangala *et al* (2017) is the only comprehensive, large-scale study on this (Covey and Megonigal 2019) we assume that the tuned ‘ A ’ value is valid over all vegetation types. Applying this ‘ A ’ value to different plant functional types is non-ideal, however its impact on total flux over other vegetation types is small partially due to shallower rooting depths. (Figure S5 shows that including vegetation transfer has little effect outside tropical South America). This finding is consistent with McNorton *et al* (2016), which shows that incorporating additional processes, such as CH_4 transport via short vegetation, does not improve overall CH_4 flux simulation.

JULES CH_4 parameterisation

CH_4 production in wetlands occurs via two main pathways: acetoclastic and hydrogenotrophic methanogenesis. Acetic acid from root exudates (related to net primary productivity NPP; Bridgham *et al* 2013) is the substrate for the former, whereas CO_2 , from the soil carbon store is a source for the latter. There is considerable uncertainty about the relative importance of the two substrates in CH_4 generation at the large-scale however, causing speculation that this is a large contributory factor in future wetland CH_4 flux uncertainty (Bridgham *et al* 2013). To place bounds on this uncertainty, JULES is run with two CH_4 parameterisations, where the source for wetland CH_4 flux into the atmosphere (F_{CH_4}) is based on either soil carbon C_s (kgm^{-2}) or NPP ($\text{kgm}^{-2} \text{s}^{-1}$). Due to the inclusion of vegetation-mediated flux, total CH_4 fluxes become dependent on an effective wetland fraction, fwe (where $fwe = fwe_s + fwe_v$):

$$F_{\text{CH}_4}(\text{Cs}) = fwe \cdot K_{\text{CS}} \cdot C_s \cdot Q10_{\text{CS}}(T) \frac{T - T_0}{10}, \quad (2)$$

$$F_{\text{CH}_4}(\text{NPP}) = fwe \cdot K_{\text{NPP}} \cdot \text{NPP} \cdot Q10_{\text{NPP}}(T) \frac{T - T_0}{10}, \quad (3)$$

where T (K) is the mean top 1 m soil temperature and T_0 a reference temperature (273.16 K). $F_{\text{CH}_4}(\text{NPP})$ is restricted from going below zero, as would otherwise occur if NPP becomes negative. (Soil uptake is included within the atmospheric methane lifetime). The choice of two substrates is important, as they are likely to have different climate sensitivities: $F_{\text{CH}_4}(\text{Cs})$ is proportional to a large pool, which we assume is time invariant; $F_{\text{CH}_4}(\text{NPP})$ is sensitive to a change in flux, with initial soil CH_4 emissions occurring less than twelve hours after photosynthetic assimilation (Megonigal *et al* 1999). Both schemes are sensitive to

climate through their Q10 temperature relationships and effective wetland fraction (fwe), with $F_{\text{CH}_4}(\text{NPP})$ having an additional substrate sensitivity to climate and atmospheric CO_2 . These two schemes are chosen to see if uncertainties in the temporal production of available substrate is important in the climate- CH_4 feedback response.

K_{CS} and K_{NPP} are tuned to produce the appropriate global total wetland flux to cover the total range for all models in Saunio *et al* (2016), 127–227 $\text{TgCH}_4\text{yr}^{-1}$. (K_{CS} varies from $3.2\text{--}5.0 \times 10^{-12} \text{s}^{-1}$ and K_{NPP} $4.1\text{--}8.0 \times 10^{-3}$ (unitless) from UPP to LOW). Q10(T) factors in equations (2) and (3) describe the amounts by which reaction rates increase with a 10 K temperature increase. These factors are themselves a function of temperature, so as to directly follow the Arrhenius equation which describes the temperature dependence of a biological process. Thus using a temperature dependent Q10 allows us to fit the Arrhenius equation beyond a specific climatic region. Hence $Q10(T) = Q10_{T_0}^{T_0/T}$, where $Q10_{T_0}$ is the Q10 temperature sensitivity at T_0 . $Q10_{\text{CS},T_0}$ and $Q10_{\text{NPP},T_0}$ are specific to the methanogenesis pathways considered.

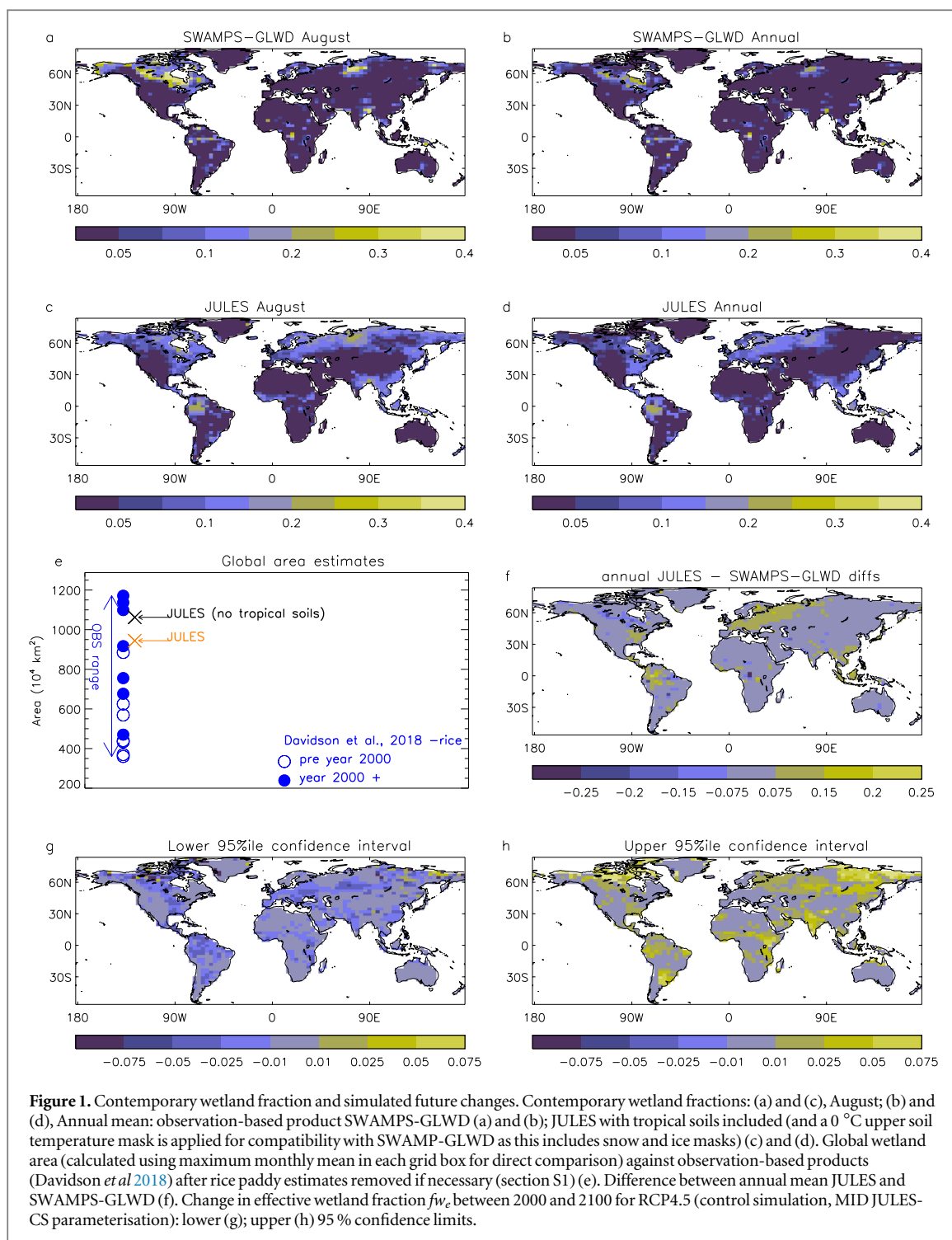
The above scheme does not include some more detailed processes, such as CH_4 suppression due to sulphate deposition, some transport mechanisms, and multiple CH_4 pools. McNorton *et al* (2016) demonstrate that their inclusion does not improve overall model performance, and there are insufficient observations to adequately constrain many such processes (e.g. redox and pH, Riley *et al* 2011). Instead we focus on constraining temperature sensitivity using a large number of observations over an extensive number of field sites (Turetsky *et al* 2014), as uncertainty in temperature response has the largest impact on net CH_4 emissions (Riley *et al* 2011).

JULES wetland and CH_4 calibration and validation

In order to calibrate the JULES wetland CH_4 relations (JULES-CS, JULES-NPP) off-line runs are carried out using observation-based meteorology (Weedon *et al* 2014) at 0.5° resolution for all land points.

The wetland fraction simulated by JULES compares well spatially (figure 1) and temporally (figure S2) with the SWAMPS-GLWD (Poulter *et al* 2017) observation-based product, and is within observational spread for global total wetland area (figure 1(e), Davidson *et al* 2018). A detailed assessment of simulated wetland uncertainty and its impact on the magnitude of wetland CH_4 feedback on climate change is provided in the supplementary material SI sections S1 and S7.

To calibrate the soil carbon-based emission relation JULES-CS, temperature sensitivity $Q10_{T_0,\text{CS}}$ is determined by fitting against Q10's in the multiple wetland site data analysis (Turetsky *et al* 2014, $Q10_{\text{TUR}}$,



section S2). To directly map $Q_{10_{To,CS}}$ onto $Q_{10_{Tur}}$, JULES-CS must use *fixed* (observed) soil carbon (Nachtergaele and Batjes 2012), which is equivalent to assuming a highly recalcitrant and/or large soil carbon pool that changes negligibly over time. Using this fitting procedure we produce lower (LOW), mid (MID) and upper (UPP) estimates of $Q_{10_{CS,To}}$ of 3.0, 3.7 and 4.7 (equivalent to poor and rich fens, and bogs in Turetsky *et al* 2014), respectively (SI table S1).

The intermediate $Q_{10_{CS,To}}$ value of 3.7 obtained from the above bottom-up approach here is very close to that found in Gedney and Cox (2003) (3.4–3.7),

where the CH_4 parameterisation was calibrated against global inter-annual variations in atmospheric CH_4 (a top-down approach). Although Turetsky *et al* (2014) focuses on bogs and fens, Yvon-Durocher *et al* (2014) demonstrate that for multiple eco-systems spanning the globe, CH_4 emissions have a consistent mean temperature dependence (0.96 eV with a 95% confidence of 0.86–1.07 eV). This is numerically equivalent to $Q_{10_{To}} = 4.5$ and range of 3.8–5.3 (as 1 eV $\sim 1.6 \times 10^{-19}$ J), and consistent with our estimates of 3.0–4.7. The resulting regional distribution of JULES CH_4 fluxes compare well against other

estimates (Saunois *et al* 2016, section S4). The Amazon basin emissions and uncertainty ranges produced by JULES also compare well with estimates extrapolated from a large-scale measurement campaign (JULES: 24.6–53.9, Pangala *et al* 2017: 24.4–53.5 TgCH₄yr⁻¹; table S2). (As we have tuned the tree-mediated fluxes to be 50% of the total, and our flux totals agree with observations, it follows that there is also good agreement between the tree-mediated fluxes: JULES 12.3–27.0, Pangala *et al* 2017 11.5–26.2 TgCH₄yr⁻¹). The aforementioned comparisons demonstrate consistency across very different calibration approaches and different spatial and temporal scales, indicating that Q10_{CS,T₀} is appropriately calibrated for a global-scale study.

We are able to use the regionally-calibrated Q10_{CS,T₀} globally because CH₄ fluxes are observed to have very similar temperature dependencies for a wide range of eco-systems and regions (Yvon-Durocher *et al* 2014). The above mapping approach cannot be directly applied to the NPP substrate-based CH₄ emissions relation (JULES-NPP) however, because NPP is itself dependent on soil moisture and temperature. Instead, we calibrate JULES-NPP by comparing against regional fluxes across the optimised JULES-CS (section S2). The resulting values of Q10_{NPP,T₀} for poor and rich fens, and bogs are 1.3, 1.6 and 2.3, respectively (SI figure S3(c), table S1).

From these comparisons with multiple measurements and models over different spatial and temporal scales, we demonstrate that both the JULES wetland extent and CH₄ emission parameterisations respond appropriately to temperature and precipitation and are therefore suitable for the analysis of large-scale CH₄ emissions in future climate scenarios.

Climate change simulations

To analyse the potential for wetland CH₄ emissions to feedback on climate change, multiple simulations are performed using JULES coupled to a climate change impacts model (IMOGEN, section S5). IMOGEN (Huntingford *et al* 2010) is calibrated against 34 CMIP5 ESM climate simulations (table S6). We perform simulations for 3 RCP GHG scenarios: 2.6, 4.5 and 8.5 (Collins *et al* 2013), which range from a high mitigation strategy to ‘business-as-usual’. All GHG concentrations, except for CH₄, are prescribed to RCP values, so that interactive wetland CH₄ emissions can modify atmospheric CH₄ concentration and radiative forcing (section S5, Etminan *et al* 2016). To assess the strength of this wetland CH₄-climate feedback, separate simulation sets are carried out with wetland emissions either fixed in time (REF), or able to respond to changes in meteorology.

The atmospheric lifetime of CH₄ includes losses from tropospheric OH, stratospheric loss and soil sink, and is also dependent on CH₄ concentration as

increasing CH₄ reduces tropospheric OH (CH₄-OH factor *s*, section S5). Myhre *et al* (2013) (3.SM.2) quote contemporary lifetime $\tau_o = 9.25 \pm 0.6$ years and $s = 0.25 \pm 0.03$. For our control experiment (CTL) we set $\tau_o = 9.25$, $s = 0.25$. To cover uncertainty we add two sensitivity experiments with $\tau_o = 8.65$, $s = 0.22$ and $\tau_o = 9.85$, $s = 0.28$ (section S5).

Both CH₄ substrate relations, their Q10_{T₀}'s, and present-day global wetland emissions estimates and their uncertainties, are all considered in the control experiment ensemble (table S1, CTL). (For the purposes of climate simulations here, the lower, median and upper 2000–2009 average wetland flux from current best estimates (Saunois *et al* 2016), are used as the year 2000 LOW, MID and UPP global wetland flux totals, respectively). To investigate uncertainty in modelled physical processes further, additional model ensembles are run without the JULES vegetation-mediated CH₄ transfer or tropical soils. We also investigate the impact of: modelled wetland extent errors using an observation-based mask; reduced substrate availability with soil depth; soil carbon uncertainties and; limiting the vegetation flux to Amazonia (sections S3, S4, S7). (Amazonia is the only region where the vegetation flux is studied comprehensively, and measurements from other regions suggest this flux may be smaller elsewhere, Covey and Megonigal 2019).

Analysing the main feedback drivers and sources of uncertainties

To rigorously assess the relative importance of the drivers of CH₄ emission changes, JULES-CS and JULES-NPP, equations (2) and (3) are differentiated as functions of temperature and wetland (and NPP for JULES-NPP) giving:

$$dF_{\text{CH}_4}(\text{Cs}) \sim F_{\text{CH}_4}(\text{Cs}) \left(\frac{dfwe}{fwe} + 0.1 \ln(Q10_{\text{CS}}(T)) dT \right). \quad (4)$$

(As we assume a time invariant soil carbon is the main substrate source, Cs is approximated as changing little over the time period considered). For JULES-NPP, equation (3), NPP is not invariant in time, so this differentiates to:

$$dF_{\text{CH}_4}(\text{NPP}) \sim F_{\text{CH}_4}(\text{NPP}) \left(\frac{dfwe}{fwe} + 0.1 \ln(Q10_{\text{NPP}}(T)) dT + \frac{d\text{NPP}}{\text{NPP}} \right). \quad (5)$$

Changes in wetland fraction and precipitation are strongly related (Papa *et al* 2010) so equations (4) and (5) are approximated to:

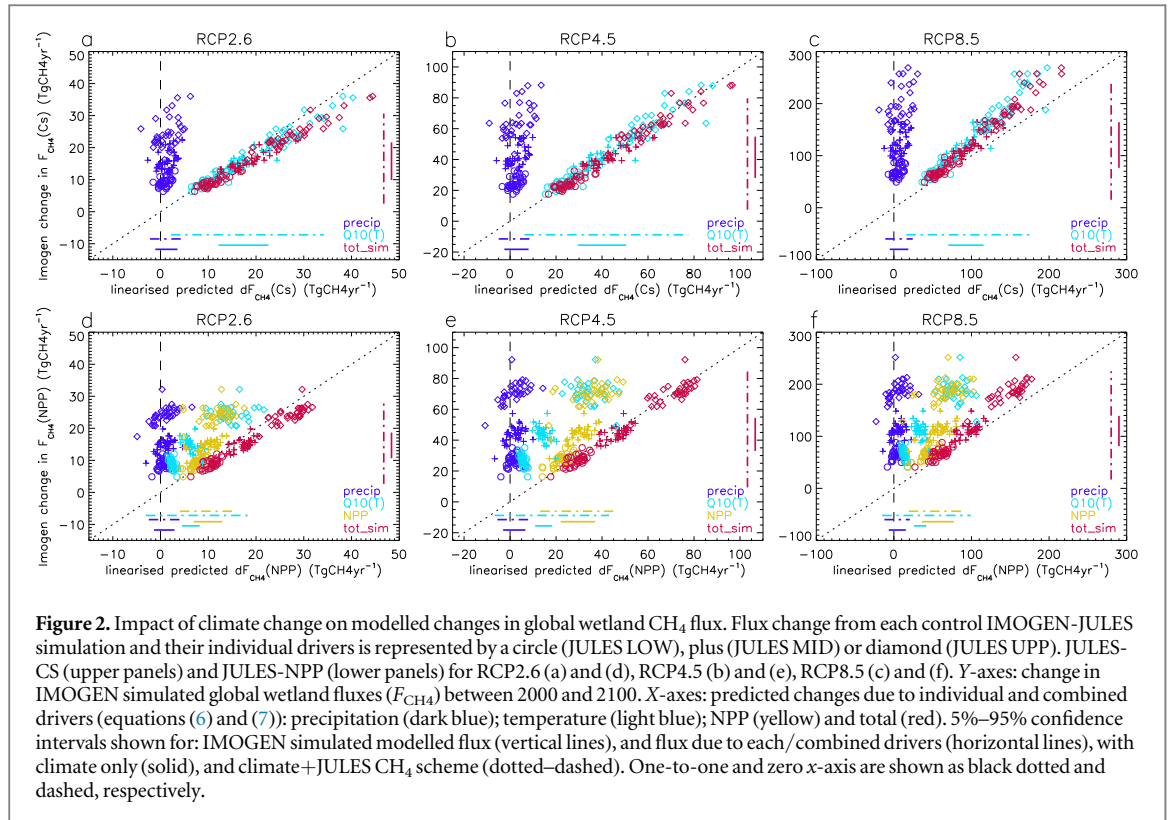


Figure 2. Impact of climate change on modelled changes in global wetland CH_4 flux. Flux change from each control IMOGEN-JULES simulation and their individual drivers is represented by a circle (JULES LOW), plus (JULES MID) or diamond (JULES UPP). JULES-CS (upper panels) and JULES-NPP (lower panels) for RCP2.6 (a) and (d), RCP4.5 (b) and (e), RCP8.5 (c) and (f). Y-axes: change in IMOGEN simulated global wetland fluxes (F_{CH_4}) between 2000 and 2100. X-axes: predicted changes due to individual and combined drivers (equations (6) and (7)): precipitation (dark blue); temperature (light blue); NPP (yellow) and total (red). 5%–95% confidence intervals shown for: IMOGEN simulated modelled flux (vertical lines), and flux due to each/combined drivers (horizontal lines), with climate only (solid), and climate+JULES CH_4 scheme (dotted–dashed). One-to-one and zero x-axis are shown as black dotted and dashed, respectively.

$$dF_{\text{CH}_4}(\text{Cs}) \sim F_{\text{CH}_4}(\text{Cs}) \left(\frac{d\text{precip}}{\text{precip}} + 0.1 \ln(Q10_{\text{CS}}(T)) dT \right), \quad (6)$$

$$dF_{\text{CH}_4}(\text{NPP}) \sim F_{\text{CH}_4}(\text{NPP}) \left(\frac{d\text{precip}}{\text{precip}} + 0.1 \ln(Q10_{\text{NPP}}(T)) dT + \frac{d\text{NPP}}{\text{NPP}} \right). \quad (7)$$

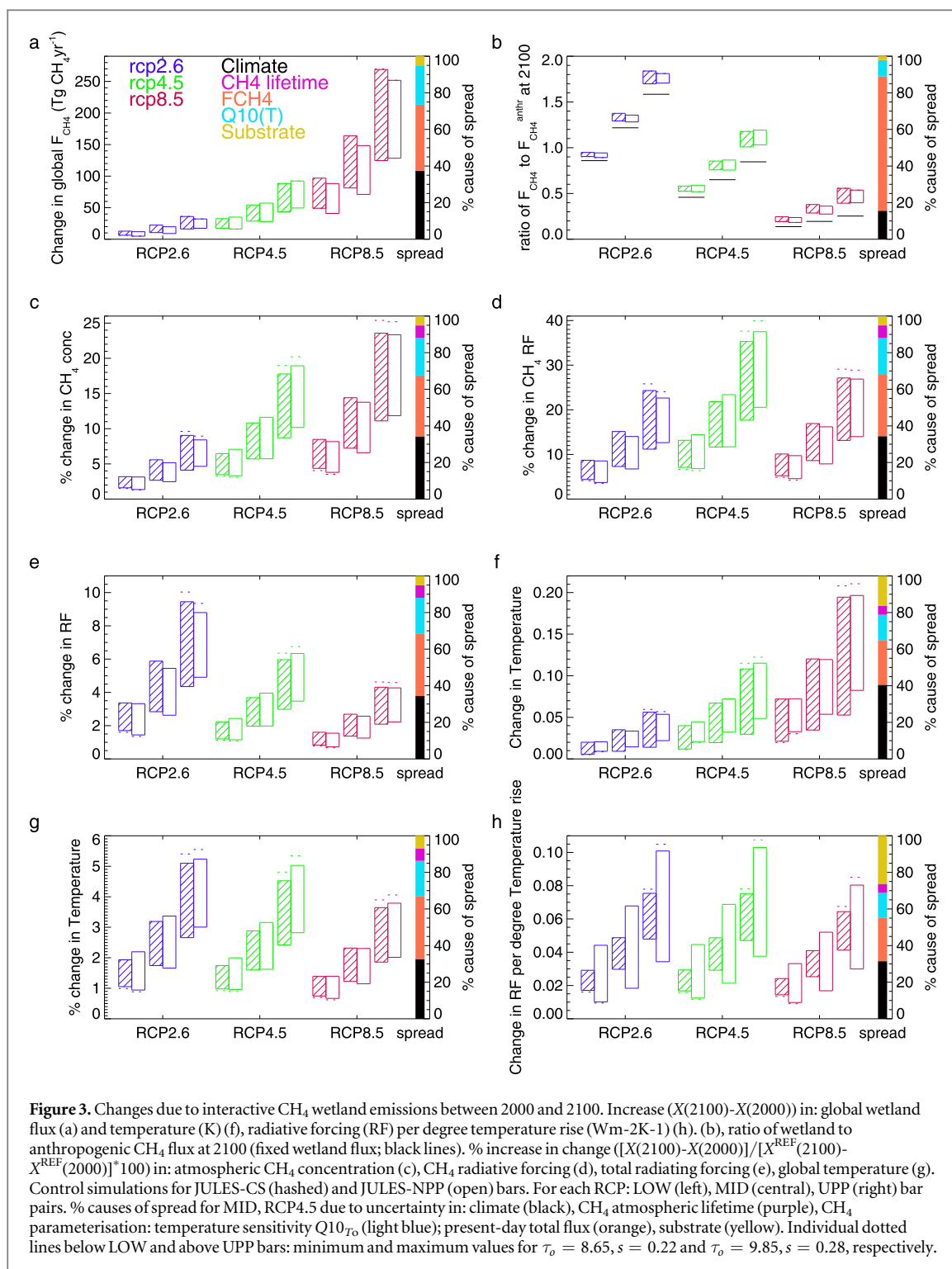
Results

For the highest GHG scenario RCP8.5, global wetland CH_4 emissions are projected to increase from $173 \text{ TgCH}_4\text{yr}^{-1}$ at 2000 to $254\text{--}337 \text{ TgCH}_4\text{yr}^{-1}$ and $244\text{--}321 \text{ TgCH}_4\text{yr}^{-1}$ by 2100 for MID JULES-CS and JULES-NPP, respectively. A maximum flux of $495 \text{ TgCH}_4\text{yr}^{-1}$ is predicted at 2100 for the ‘UPP’ estimate (or $494 \text{ TgCH}_4\text{yr}^{-1}$ if only considering the best estimate atmospheric lifetime, table S4, figures 2, 3(a)). Fractional increases in the tropics and extratropics are similar (figure S8). Hence future global growth in wetland CH_4 flux is dominated by the tropics as this is the main source of present-day emissions.

To understand the causes of these flux changes and their uncertainties, we present the impact of temperature, precipitation and NPP (light and dark blue and yellow symbols, respectively; figure 2) on global CH_4 flux changes (equations (6) and (7)). Wetland extent responses are dominated by differences in precipitation

projections (as demonstrated by comparing figures 2 and S7) which are themselves highly uncertain. Consequently the simulations do not predict a consistent expansion or contraction over most wetland regions (figures 1(g) and (h)). There is even a lack of consensus as to whether the global total wetland area is likely to increase or decrease in the future. (For example, for RCP8.5 simulations it varies from a 2% decrease to a 19% increase between 2000 and 2100). Hence projected changes in precipitation result in simulated CH_4 flux changes of uncertain sign. The resulting flux changes are of relatively small magnitude however. Projected temperature change is a more important driver than precipitation, and results in both a larger CH_4 flux response and a greater uncertainty (figure 2). For JULES-NPP, some of the temperature response is within the NPP term. As well as being dependent on temperature, NPP is also strongly driven by atmospheric CO_2 (through increased CO_2 fertilisation), which is in itself highly correlated with temperature.

In both substrate parameterisations the spread in projected wetland CH_4 flux by year 2100 is dominated by uncertainty in the wetland emissions parameterisation (figure 3). The uncertainty due to simulated climate change (within a specific RCP) plays a smaller, but still significant role (63%–65% versus 35%–37% over the three RCP scenarios—only RCP4.5 shown in figure 3(a)). Within the wetland scheme, uncertainty due to present-day global wetland flux estimate is larger than that from temperature sensitivity ($Q10_{T_0}$, figure 3(a)). Moreover the projected spread due to global present-day wetland emissions uncertainty alone is



comparable to that from simulated climate change. Critically, in terms of CH₄ generation process uncertainty, the impact of substrate source used (i.e. Cs versus NPP), is relatively small (between 5% and 13% for the different scenarios). We also find that the wetland physics sensitivity experiments have little impact on the projected future wetland flux (section S7).

Enhanced wetland CH₄ emissions cause substantial 21st century increases in atmospheric CH₄ concentration of up to 25.4% above that with fixed

wetland emissions (and 23.6% without including the CH₄ atmospheric lifetime sensitivity experiments) (figures 3(c) and 4). Depending on the scenario, the corresponding 21st century increases in total and CH₄ radiative forcings are further enhanced by between 0.7%–10.0% and 3.5%–40.0%, respectively. The percentage changes in CH₄ radiative forcing are higher for lower RCP scenarios (figures 3(c) and (d)) because CH₄ radiative forcing is nonlinearly dependent on concentration (Etminan *et al* 2016). This radiative

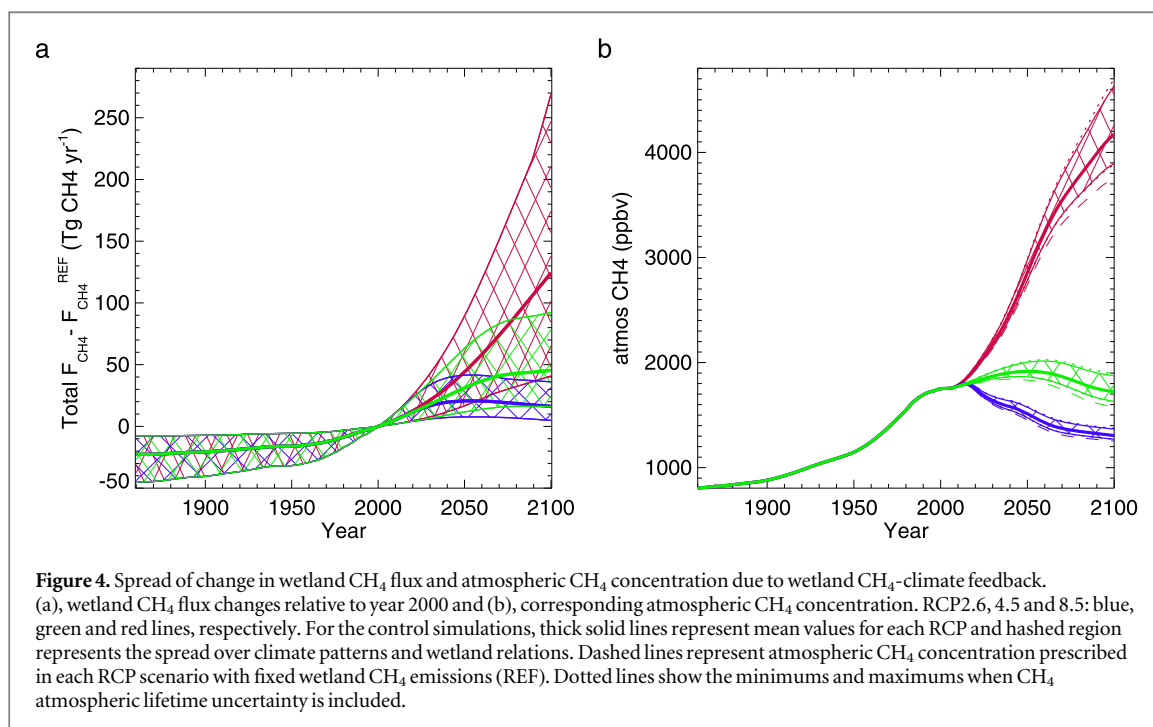


Figure 4. Spread of change in wetland CH_4 flux and atmospheric CH_4 concentration due to wetland CH_4 -climate feedback. (a), wetland CH_4 flux changes relative to year 2000 and (b), corresponding atmospheric CH_4 concentration. RCP2.6, 4.5 and 8.5: blue, green and red lines, respectively. For the control simulations, thick solid lines represent mean values for each RCP and hashed region represents the spread over climate patterns and wetland relations. Dashed lines represent atmospheric CH_4 concentration prescribed in each RCP scenario with fixed wetland CH_4 emissions (REF). Dotted lines show the minimums and maximums when CH_4 atmospheric lifetime uncertainty is included.

Table 1. Reduction in allowed anthropogenic (fossil fuel and land use related) emissions between 2020 and 2100 due to wetland CH_4 -climate feedback. Average, and range of 68% probability (mean and ± 1 standard deviation) shown in brackets. The values are based on the control and CH_4 atmospheric lifetime sensitivity experiments.

	RCP2.6	RCP4.5	RCP8.5
Remaining anthr emissions (without wetland- CH_4 feedback)	268.2	661.4	1809.3
Reduction (GtC)	21.4 (12.4–30.4)	36.7 (21.6–51.9)	79.1 (45.0–113.2)
Reduction as % of remaining	8.0 (4.6–11.3)	5.6 (3.3–7.8)	4.4 (2.5–6.3)

forcing enhancement leads to an additional increase in global air temperature of up to 0.21 K at 2100 (and land air temperature of up to 0.27 K -not shown).

The relative causes of uncertainty in the enhanced total radiative forcing and percentage temperature changes are similar to those in wetland emissions (figure 3). In addition, atmospheric CH_4 lifetime uncertainty contributes around 7% of the total spread in radiative forcing and temperature changes.

Due to wetland emissions feedback, significant reductions in allowed anthropogenic emissions are required to maintain projected temperature changes between 2020 and 2100 (appendix SB, table 1). For RCP2.6 these reductions are 21.4 GtC, with a likely value (greater than 68% probability) of between 12.4–30.4 GtC (4.6%–11.3%). This is consistent with the Comyn-Platt *et al* (2018) estimate of 19.6 GtC for the comparable 1.5 °C stabilisation threshold. (This is despite Comyn-Platt *et al* 2018 using a different substrate generation scheme, thereby further demonstrating that the impact of substrate used is relatively small). Near-term implications are significant with the RCP2.6 mid 21st century global temperature peak reached nine years earlier (4–16 years) (not shown).

CH_4 emissions from wetlands increase significantly from present day to 2100 in all simulated

climate change patterns and RCP scenarios. In the higher sensitivity wetland emission relation, they may more than double between 2000 and 2100, with an associated change in atmospheric CH_4 concentration of up to 952 ppbv (figure 4). The subsequent feedback on radiative forcing ranges from 0.01 to 0.11 $\text{Wm}^{-2} \text{K}^{-1}$. Despite the improved physical representation in JULES, this has little impact on the feedback magnitude of the wetland emissions on climate at the global scale (section S7).

Discussion and conclusions

There is also considerable uncertainty in other inland water CH_4 source emissions (Saunio *et al* 2016) and these are also likely to respond to climate change (Ciais *et al* 2013). Present-day rice agriculture emissions are estimated to be around 36 $\text{TgCH}_4\text{yr}^{-1}$ which is between 16% and 30% of natural wetland emission estimates (Saunio *et al* 2016). Moreover, even without rice agriculture expanding, their associated CH_4 emissions are likely to increase in response to temperature (Khalil *et al* 1998), adding an additional radiative feedback.

We have included new physical processes: the CH_4 transfer through tropical trees and the inclusion of tropical-specific soils, to improve wetland representation

in this important region. This allows us to make more accurate assessments of the feedback magnitude between wetland CH₄ release and climate change. We have determined this feedback uncertainty by constraining wetland parameterisations, using data from multiple sources over local to regional scales. Despite this, there remains a sizeable feedback uncertainty, which is dominated by relatively poorly known wetland extent and CH₄ parameters, rather than that due to climate change. Our more rigorous and constrained approach helps explain the IPCC's assessment of 'low confidence' in the magnitude of the wetland CH₄-climate feedback (Ciais *et al* 2013). Furthermore, we identify the main causes of the uncertainty range, which can be further constrained through future measurement campaigns focusing on detailed wetland processes, including those determining the vegetation transport of CH₄ (Covey and Megonigal 2019). Given that the remaining lack of knowledge in contemporary global wetland CH₄ flux is the single largest cause of wetland scheme uncertainty, improved estimates of present-day global wetland emissions remain a research priority.

The CH₄ feedback on climate may be enhanced or reduced through interactions with other parts of the earth system through the intricate coupling of biogeochemical cycling and reactive atmospheric chemistry. For instance, we do not consider the interaction with the carbon cycle in which changing atmospheric CO₂ and climate may alter the size and distribution of vegetation and soil carbon, and in turn the available substrate for methanogenesis. Understanding these interactions, and their potential nonlinearities, will become possible as the processes analysed in this study are routinely incorporated in to the next generation of ESM, and we suggest this should be a priority.

Under a range of future socio-economic assumptions, from high GHG emission mitigation to 'business-as-usual' scenarios, we find that wetland CH₄ feedbacks in a warming climate will significantly augment atmospheric CH₄ concentrations by up to 25.4% over the 21st century. This raises CH₄ radiative forcing significantly beyond that caused by direct human activity, producing a positive feedback that further enhances climate change. Over the scenarios considered, the wetland feedback amplifies the 21st century CH₄ radiative forcing on average by 14.4%, with uncertainty estimates indicating values as high 40.0% for the RCP4.5 scenario. This generates an average additional warming of 2.4%. Under the RCP2.6 emissions scenario (which gives an approximately 1.5 °C increase in temperature from pre-industrial to 2100), the 21st century temperature rise is amplified by up to 5.5%, and furthermore rises more quickly in the near term. This corresponds to a likely reduction in remaining allowed anthropogenic emissions of between 4.6%–11.3% in order to maintain the same temperature profiles. The more rapid warming and carbon budget reduction, imply that a combination of enhanced short-term emission cuts and longer-term

increased use of negative emission technology, are required to be consistent with the Paris climate targets.

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