Estimating the Permafrost-Carbon Climate Response in the CMIP5 Climate Models Using a Simplified Approach

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ABSTRACT

Under climate change, thawing permafrost may cause a release of carbon, which has a positive feedback on the climate. The permafrost-carbon climate response ($\gamma_{PF}$) is the additional permafrost-carbon made vulnerable to decomposition per degree of global temperature increase. A simple framework was adopted to estimate $\gamma_{PF}$ using the database for phase 5 of the Coupled Model Intercomparison Project (CMIP5). The projected changes in the annual maximum active layer thicknesses ($ALT_{max}$) over the twenty-first century were quantified using CMIP5 soil temperatures. These changes were combined with the observed distribution of soil organic carbon and its potential decomposability to give $\gamma_{PF}$. This estimate of $\gamma_{PF}$ is dependent on the biases in the simulated present-day permafrost. This dependency was reduced by combining a reference estimate of the present-day $ALT_{max}$ with an estimate of the sensitivity of $ALT_{max}$ to temperature from the CMIP5 models. In this case, $\gamma_{PF}$ was from $-6$ to $-66$ Pg C K$^{-1}$ (5th–95th percentile) with a radiative forcing of $0.03$–$0.29$ W m$^{-2}$ K$^{-1}$. This range is mainly caused by uncertainties in the amount of soil carbon deeper in the soil profile and whether it thaws over the time scales under consideration. These results suggest that including permafrost-carbon within climate models will lead to an increase in the positive global carbon climate feedback. Under future climate change the northern high-latitude permafrost region is expected to be a small sink of carbon. Adding the permafrost-carbon response is likely to change this region to a source of carbon.

1. Introduction

Permafrost soils contain $\sim$1672 Pg of organic carbon (Tarnocai et al. 2009), much of which is permanently frozen and consequently relatively inert. Under increased temperature, permafrost degrades and a proportion of this old permafrost-carbon will become more vulnerable to decomposition and subsequent release into the climate system. Like fossil fuel burning, this is an irreversible process over time scales of hundreds of years and it has the potential to cause a further increase in greenhouse gases in the atmosphere, leading to a positive carbon–climate feedback (Schuur et al. 2008).

To help quantify the permafrost-carbon climate feedback, we need to know how much carbon is vulnerable to release, how fast it will be released, and whether it will be released as carbon dioxide (CO$_2$) or methane (CH$_4$). These processes are highly uncertain, spatially variable, and often hard to measure (Schuur and Abbott 2011). Burke et al. (2012) developed a simple framework to estimate the additional temperature increase caused by permafrost-carbon loss that includes the impact of many of these uncertainties. They suggest that, had permafrost-carbon been included within the Hadley Centre climate model HadGEM2-ES (note that all model names used in this paper are expanded in Table 1), there would be an additional temperature increase of $0.02^\circ$–$0.36^\circ$C (90% range). Schneider von Deimling et al. (2012) used an alternative simple approach, again included relevant uncertainties, and suggested there might be an additional warming of $0.04^\circ$–$0.23^\circ$C (68% range). MacDougall
et al. (2012), Koven et al. (2011), and Schaefer et al. (2011) developed their land surface schemes to include a simple representation of permafrost-carbon and quantified the permafrost-carbon lost, but they incorporated a more limited uncertainty assessment. MacDougall et al. (2012) coupled their land surface scheme with a global climate model of intermediate complexity and found an additional warming of $0.09^\circ-0.75^\circ C$ caused by the permafrost-carbon climate feedback. In general, the permafrost-carbon climate feedback is not yet included within coupled earth system general circulation models (GCMs).

In this paper, the permafrost-carbon climate response ($\gamma_{PF}$) is defined as the permafrost-carbon made vulnerable to decomposition per degree of global temperature increase. Negative values represent a loss of carbon from the land surface. This is a committed loss (i.e., the amount that may ultimately be lost if the temperatures were to stabilize). The majority of other studies estimate the amount lost over a specified time frame. This is a realized loss. The committed definition of $\gamma_{PF}$ was used because it significantly reduces the dependence of $\gamma_{PF}$ on the rate of change of global mean temperature. The amount of vulnerable permafrost-carbon can be quantified using the simple framework developed by Burke et al. (2012). They took the change in the maximum thaw depth output from a global climate model, quantified the amount of soil carbon in this depth range and, using an estimate of its quality, determined how much soil carbon was available for decomposition. Their approach is adopted here to quantify $\gamma_{PF}$ for the phase 5 of the Coupled Model Intercomparison Project (CMIP5) ensemble of global climate models.

Other carbon climate feedbacks currently represented within GCMs include the change in soil and vegetation carbon in response to climate change. The Coupled Climate Carbon Cycle Model Intercomparison Project (C4MIP) generation of models shows a global release of land carbon under increasing temperature and an uptake of land carbon under increasing $CO_2$ with a net positive feedback on to the global climate (Friedlingstein et al. 2006). In contrast, in the northern high latitudes, Qian et al. (2010) showed that the majority of the C4MIP models have a slight uptake of carbon in response to climate change. This suggests that, in this region and when permafrost-carbon is not included, the increase in primary productivity and litterfall outweighs any increase in soil respiration rates at the higher temperatures. Including permafrost-carbon processes may change the total high-latitude terrestrial response from sink to source (Koven et al. 2011). Here we make the simplifying assumption that the terrestrial carbon feedback at high latitudes can be split into two terms: a land feedback without permafrost and a feedback from permafrost with any interactions, a second-order effect. The land feedback ($\gamma_L; P_g C K^{-1}$) is quantified by the change in the global land carbon per degree of global mean temperature change. As before, negative values represent a loss of carbon from the land surface. Combining $\gamma_L$ with $\gamma_{PF}$ provides an estimate of the overall carbon–climate response including permafrost.

2. Models and methods
   a. Reference data

Large-scale observations of the annual maximum of the active layer thickness ($ALT_{max}$) are unavailable. A reference estimate of the present-day $ALT_{max}$ was obtained from the Joint UK Land Environment Simulator (JULES) land surface scheme (Best et al. 2011) driven by the Water and Global Change (WATCH) forcing data (Weedon et al. 2011) at a $2^\circ$ resolution (Burke et al. 2013). Using this reference model allows us to remove the considerable uncertainty associated with initial permafrost distributions across the CMIP5 models, while still allowing us to sample model uncertainty arising from other aspects of the CMIP5 ensemble, such as the climate sensitivity, arctic amplification, and sensitivity of active layer deepening to warming. This version of JULES calculates the soil temperature every 10 cm within the soil profile. A predecessor of the JULES land surface scheme the Met Office Surface Exchange Scheme (MOSES)/Top-Down Representation of Interactive Foliage and Flora Including Dynamics (TRIFFID) is included within HadGEM2, which overestimates the permafrost extent (Table 1) and underestimates the difference between the air and soil temperatures (Koven et al. 2013). However, the snow scheme within JULES is an updated multilayer snow scheme that increases the snow insulation effect and provides a more realistic estimate of permafrost extent (Burke et al. 2013).

b. CMIP5 global climate models

The data analyzed here were obtained from phase 5 of the Coupled Model Intercomparison Project (CMIP5) multimodel data archive. These CMIP5 global climate models support the Intergovernmental Panel on Climate Change Fifth Assessment Report (IPCC AR5). The experiments discussed include the “historical” experiments from the mid-1800s to the present day and two future scenarios for the twenty-first century, called rcp45 and rcp85. Representative concentration pathways (RCPs) 4.5 and 8.5 correspond to forcings of 4.5 and 8.5 $W m^{-2}$ by 2100 respectively, and represent intermediate and high warming scenarios (Moss et al. 2010). Data from 17 of the CMIP5 multimodel ensemble members that provided depth-resolved soil temperatures were used (Koven et al. 2013). This subset includes models with and without an
interactive carbon cycle and those models with an interactive carbon cycle do not include nutrient dynamics. The multimodel ensemble spans a range of climate sensitivities and contains several different representations of land surface processes. The CMIP5 models’ representation of permafrost dynamics is discussed in detail by Koven et al. (2013) and Slater and Lawrence (2013).

For comparison purposes, the sensitivity of land carbon (no permafrost) to climate change ($g_{L}$; Pg C K$^{-2}$1) was estimated from the climate models that include the interactive carbon cycle. Specialized simulations were used to derive $g_{L}$ in which the radiation code sees the increase in atmospheric CO$_2$ but the carbon cycle does not (Friedlingstein et al. 2006). In other words, the CO$_2$ is radiatively active and biogeochemically inert. In the CMIP5 experimental design “esmFdbck1” is such a set of simulations with the atmospheric CO$_2$ increasing by 1% yr$^{-2}$1. Data from these simulations are only available for a small subset of the models.

c. Estimating the maximum active layer thickness (ALT$_{max}$)

If the permafrost-carbon feedback had been included within the CMIP5 models, the physical state of the permafrost would have been quantified directly from the CMIP5 simulated soil temperatures with no corrections for biases. Therefore one estimate of $g_{PF}$ uses these uncorrected monthly mean soil temperatures to diagnose the thaw depth or active layer thickness (ALT). For each month and in each grid cell the ALT was defined as the deepest point in the soil column with soil temperature at or above freezing (Koven et al. 2013). The vertical resolution of the land surface schemes in many of the CMIP5 models is often poor, leading to possible biases in the estimate of the active layer thickness (see Burke et al. 2012 for details). The annual maximum of the active layer (ALT$_{max}$) was then defined from these monthly values. Permafrost was assumed to

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<table>
<thead>
<tr>
<th>Model</th>
<th>Historical mean PF extent (million km$^2$; 1995–2000)</th>
<th>RCP4.5 mean PF extent (million km$^2$; 2095–2100)</th>
<th>RCP8.5 mean PF extent (million km$^2$; 2095–2100)</th>
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be present in any given grid cell where ALT_{max} is shallower than 3 m. This approach provides an estimate of shallow permafrost and ties in with the observed distribution of soil organic carbon content that is available for the top 3 m of the soil. The value of ALT_{max} was defined annually for the permafrost region for the years between 1995 and 2100. The baseline ALT_{max} (ALT_{b, max}) is an estimate of the present-day value; ALT_{b, max} was defined as the maximum of the monthly ALT during the period 1995–2000. ALT_{max} was calculated for both the rcp45 and rcp85 simulations. Results using this set of ALT_{max} are denoted “uncorrected.”

The CMIP5 soil temperatures have considerable errors when compared with observations (Koven et al. 2013; Slater and Lawrence 2013), leading to biases in the diagnosed present-day permafrost state. Therefore, an alternative estimate of twenty-first-century values of ALT_{max} was made by combining the reference estimate of the present-day permafrost state (ALT_{b, max}) with an estimated rate of change of ALT_{max} with temperature sampled from the CMIP5 ensemble. For each CMIP5 model and each RCP scenario the rate of change of ALT_{max} per degree (ALT sensitivity) was calculated using a linear regression fit between ALT_{max} and the local near surface annual mean air temperature on a grid point by grid point basis for all grid points where there is permafrost in the top 3 m. If ALT_{max} becomes greater than 3 m all subsequent times are excluded from the analysis. This analysis assumes there is a linear relationship between ALT_{max} and local temperature. However, Koven et al. (2013) show that the sensitivity of ALT_{max} to local temperature tends to increase as temperatures approach the thawing point. This is emphasized by poor soil discretization in some of the CMIP5 models. This assumption means that the simulated permafrost thaws too early during the twenty-first century but has a smaller impact on errors in the depth of ALT_{max} by the end of the twenty-first century. The regression fits result in a distribution of ALT sensitivity that samples uncertainty from the different CMIP5 models.

In reality, ALT sensitivity is also spatially variable and dependent on, among other things, soil type, vegetation cover, and climatology. Because many of the CMIP5 models simulate a very low permafrost extent (Table 1, column 1; Koven et al. 2013; Slater and Lawrence 2013) and therefore indicate no permafrost in regions where the reference data suggests permafrost exists, it is not possible to estimate this spatial distribution for the observed permafrost extent. However, the 5th–95th percentile range of the distribution of ALT sensitivity (0.02–0.29 m K^{-1}) can be used as the upper and lower limits. While these values will not be representative of any particular model or region, they will encompass the plausible range of sensitivities within the northern high-latitude permafrost zone.

Time series of twenty-first century ALT_{max} were reconstructed for each grid point where there is permafrost in the present-day reference estimate by sampling ALT sensitivity from the 5th–95th percentile range and combining it with the local annual mean near-surface air temperature change for each CMIP5 model and each RCP scenario [Eqs. (1) and (2)]:

\[
ALT_{max}(t,x) = ALT_{max}(t-1,x) + ALT sensitivity[T_{local}(t,x)] - T_{local}(t-1,x),
\]

\[
ALT_{max}(t=0,x) = ALT_{b, max}(x),
\]

where \(T_{local}\) is the local temperature, \(t\) is the time, and \(x\) identifies each grid cell that has permafrost in the present-day reference estimate of the active layer. Here, the CMIP5 ensemble was used to sample both the uncertainty in the warming and the uncertainty in the response of ALT_{max} to this warming. Results using this derivation of ALT_{max} are denoted “bias-substituted.”

These two estimates of twenty-first century ALT_{max} can then be combined with the observed distribution of soil organic carbon content to provide two estimates of the permafrost climate response. The uncorrected \(\gamma_{PF}\) would have been obtained if the permafrost climate feedback were included in the current generation of climate models, and the bias-substituted \(\gamma_{PF}\) is an estimate using the permafrost susceptibility to air temperature derived from the CMIP5 ensemble referenced to the JULES-WATCH ALT_{b, max}. Both estimates have large associated uncertainties.

d. Estimating the permafrost-carbon climate response (\(\gamma_{PF}\))

The observed distribution of the soil organic carbon content (SOC) was taken from the Northern Circumpolar Soil Carbon Database (NCSCD; Tarnocai et al. 2009). In general, soil carbon in the NCSCD is severely undersampled, particularly at depths greater than 100 cm, and the uncertainties in any estimates of SOC are potentially large. Burke et al. (2012) used the information provided by Tarnocai et al. (2009) to provide an estimate of SOC along with its uncertainties in the 0–100-, 100–200-, and 200–300-cm depth ranges. The SOC at 100–200 cm is assumed to be less than that at 0–100 cm and less again for 200–300 cm. The SOC of the soil at depths shallower than ALT_{b, max} was assumed to be already active within the carbon cycle and is not considered here. During any year and for any grid cell that ALT_{max} is
greater than ALT$_{max}^b$ there is a quantifiable amount of thawed permafrost carbon. A proportion of this thawed carbon is assumed to be passive, very stable, and not released over the time scale of this study. The remainder of the carbon is assumed to all decompose over time scales of about 0–200 years. This is defined as the carbon vulnerable to decomposition (C$_{vul}$) and is the amount of carbon that could contribute to the permafrost-carbon climate response. The amount of carbon that is passive is uncertain. Dutta et al. (2006) use laboratory incubations on yedoma soils and estimate the passive soil carbon to be 18% of the total. Following a literature review, Falloon et al. (1998) suggested that the passive pool could be much larger and range between 15% and 60%. This uncertainty range was included in the analysis.

The permafrost-carbon climate response ($\gamma_{PF}$; PgC K$^{-1}$) was quantified as

$$\gamma_{PF} = \frac{C_{vul}}{\Delta T},$$

where $C_{vul}$ is the mean decomposable soil organic carbon (in PgC) made vulnerable by 2095–2100 and $\Delta T$ is the mean change in global mean temperature (in K) between 1995–2000 and 2095–2100. Any permafrost-carbon released will feed back onto the global mean temperature, which will be amplified in the northern high latitudes. This process is not included here and its absence may change $\gamma_{PF}$.

e. Uncertainty assessment

Uncertainties in this simple framework are large and can arise from differences between the RCP scenarios, uncertainties in the derivation of the simulated twenty-first-century ALT$_{max}$, uncertainties in both the amount of soil organic carbon present and how much of it is decomposable over the time scales of consideration, and deviations from the simplified linear analysis presented here. In the uncorrected estimates of $\gamma_{PF}$, ALT$_{max}$ was defined once for every CMIP5 model and both RCP scenarios. Five hundred Monte Carlo simulations were carried out for each model and scenario with the soil organic carbon distribution and the proportion of passive organic carbon randomly sampled from the range of plausible values using a Latin hypercube sampling (LHS; McKay et al. 1979) strategy. In the bias-substituted estimates of $\gamma_{PF}$, ALT$_{max}$ was derived by combining the JULES-WATCH estimate of ALT$_{max}^b$ with both ALT$_{sensitivity}$ and the location, model, and scenario-specific twenty-first-century near-surface air temperature change defined as the “local” temperature change. This results in 500 realizations of the time series of ALT$_{max}$ for each model and scenario. Additional uncertainties in these realizations arise from potential errors in the JULES-WATCH estimate of ALT$_{max}^b$, the range of ALT$_{sensitivity}$ discussed above, and the CMIP5 model and RCP specific local temperature change. Dankers et al. (2011) suggested that JULES simulations of ALT$_{max}$ are slightly too deep when compared with observations. Therefore the range of uncertainties on the JULES-WATCH estimates of ALT$_{max}^b$ was set to be between 50% too deep and 20% too shallow. For the bias-substituted simulations LHS sampling was used to sample the parameters required to estimate the time series of ALT$_{max}$, the soil organic carbon distribution, and the proportion of decomposable organic carbon.

The contribution of each uncertain parameter–process to the range of $\gamma_{PF}$ was determined as in Burke et al. (2012) by splitting the parameter–processes into a set of bins and calculating the mean of $\gamma_{PF}$ for each bin. This was then compared with the mean of $\gamma_{PF}$ for all of the simulations. If $\gamma_{PF}$ is sensitive to a parameter–process there will be notable differences between the mean of $\gamma_{PF}$ in each bin and that for all of the simulations. This can be quantified using the following equation:

$$S = \frac{1}{N} \sum_{i=1}^{N} \frac{(\mu_i - \mu)^2}{\sigma^2},$$

where $\mu_i$ is the mean of bin $i$, $\mu$ is the mean of all the simulations, and $\sigma$ is the standard deviation of all of the simulations.

f. Estimating the land carbon climate feedback ($\gamma_L$)

The sensitivity of land carbon (no permafrost) to climate change ($\gamma_L$; PgC K$^{-1}$) was found using esmFdbck1 where the atmospheric CO$_2$ increases by 1% yr$^{-1}$ in the following manner:

$$\gamma_L = \frac{\Delta C_L}{\Delta T},$$

where $\Delta C_L$ (PgC) is the change in the mean total land carbon and $\Delta T$ is the change in the mean global mean temperature (K). The change is defined for the mean of the first and last 5 years of the 140-yr simulation, which starts at preindustrial CO$_2$ and ends at 4 times preindustrial CO$_2$. These simulations do not include the confounding effects of changes in land use, non-CO$_2$ greenhouse gases, aerosols, etc., and so provide a controlled experiment with which to evaluate land carbon–climate interactions (Arora et al. 2013). It is assumed that there are no interactions between the carbon currently within the GCM’s carbon cycle and any new carbon released from thawed permafrost.

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3. Results

The mean CMIP5 permafrost extents during the period 1995–2000 are highly variable and range between 2.5 and 28.5 million km$^2$ (see Table 1, ordered by increasing permafrost extent). The observed extent is estimated to be somewhere between 12.2 and 17.0 million km$^2$ with the permafrost affected area estimated to be 22.8 million km$^2$ (Zhang et al. 2003). Very few models fall within this range, with the majority underestimating the extent and a few significantly overestimating it. The models (NorESM1-M, MRI-CGM3, and MIROC5) that fall close to this range are highlighted in italics in Table 1. Koven et al. (2013) suggest that, in general, the differences between models and observations are dominated by the parameterization of surface exchange, the snow scheme, and coupled thermal–hydrological dynamics within the soil column rather than by errors in the climate model simulation of the local mean air temperature. This was also demonstrated by Burke et al. (2013), who showed significant differences in the simulated present-day permafrost extent between two simulations identical except for their representation of snow. The simulated changes in the permafrost extent in response to a changing climate are also highly variable (Table 1). However, all of the models simulate a loss of permafrost over the twenty-first century leading to additional thawed carbon, which could decompose and be released into the atmosphere. As might be expected, this loss of permafrost is greater for RCP8.5 than for RCP4.5 because RCP8.5 has a bigger increase in global mean temperature over the twenty-first century.

Figure 1 shows the permafrost-carbon climate response ($\gamma_{PF}$) for the CMIP5 model simulations and the two RCP scenarios using the uncorrected derivation of ALT$_{max}$. This is the response that might have been obtained if permafrost-carbon were included directly in the global climate models—the majority of which have large biases in their simulation of permafrost extent. The box represents the inter quartile range (25%–75%), the mean of $\gamma_{PF}$ is the horizontal black line within the box, and the whiskers represent the 5th–95th percentile range. Points outside this range are shown with dots. There is a wide spread in $\gamma_{PF}$ with values ranging from $-3$ to $-123$ PgC K$^{-1}$ for RCP4.5 and from $-1$ to $-80$ PgC K$^{-1}$ (both 5th–95th percentile) for RCP8.5. The three models with permafrost extents similar to that observed (NorESM1-M, MRI-CGM3, and MIROC5) are highlighted in gray. Using just these models to estimate $\gamma_{PF}$ changes the lower limit of $\gamma_{PF}$ from $-1$ to $-17$ PgC K$^{-1}$ (5th percentile, both RCP scenarios included), which significantly increases the minimum likely positive feedback.

For many of the models, the permafrost-carbon response is larger for RCP4.5 than for RCP8.5. RCP8.5 has a larger twenty-first century temperature increase and therefore a deeper active layer by the end of the century. However, at the greater depths the observed soil organic carbon content is lower and therefore the rate of increase of vulnerable carbon per degree temperature change is reduced at higher temperatures when compared with lower temperatures. In some places this gives a nonlinear relationship between change in ALT$_{max}$ and vulnerable soil organic carbon and causes a reduced response for RCP8.5 compared with RCP4.5. However some models such as HadGEM2 have a similar values for RCP4.5 and RCP8.5 because ALT$_{max}$ is initially deeper and the thawed permafrost releases soil organic carbon from the deeper levels irrespective of scenario. This highlights both the scenario dependence of this definition of $\gamma_{PF}$ and the importance of observing the depth distribution of the soil organic carbon content.

The models in Fig. 1 are ranked in order of increasing mean $\gamma_{PF}$. This ranking is very similar to the ranking of present-day permafrost extent in Table 1. As might be expected, models with little present-day permafrost (and hence only a small amount of soil organic carbon) have a low $\gamma_{PF}$ and vice versa. There is a strong relationship between $\gamma_{PF}$ and the total soil organic carbon content in the top 3 m of the permafrost affected zone defined using the present-day extent for each CMIP5 model (Fig. 2). Although the RCP8.5 scenario is shown, results for RCP4.5 are qualitatively similar. If the permafrost-carbon climate feedback were included within the current generation of climate models, biases in the simulated permafrost extent would significantly bias the estimate of $\gamma_{PF}$ (Figs. 1 and 2).

The error bars on the points in Fig. 2 show two standard deviations both in the total soil organic carbon content and in $\gamma_{PF}$. The uncertainties in $\gamma_{PF}$ for any particular model are driven by uncertainties in the distribution and decomposability of soil organic carbon. For the models with the small permafrost extents, uncertainties are low because there is little soil carbon present initially in the permafrost zone. The vertical lines in Fig. 2 show the mean of the observed estimate of soil organic carbon within the permafrost zone where the permafrost zone is defined by either the International Permafrost Association (IPA) permafrost extent or the JULES-WATCH permafrost extent. This falls toward the center of the range of model simulations. Also shown on Fig. 2 is a linear least squares regression fit between the initial soil organic carbon content in the permafrost region and $\gamma_{PF}$ with the intercept set to zero. The thin dashed–dotted lines are the confidence intervals for this fit. For RCP8.5 the $R^2$ value is 91%. Using this relationship, the confidence
intervals for two standard deviations, and the observations as a constraint on $\gamma_{PF}$, $\gamma_{PF}$ could be between $-18$ and $-63$ PgC K$^{-1}$ (Fig. 2). The confidence in the regression fit for RCP4.5 (not shown) is lower, leading to a wider range of $\gamma_{PF}$—between $-5$ and $-105$ PgC K$^{-1}$. These latter estimates assume that all of the models simulate the same rate of loss of organic carbon as a function of temperature and the errors are solely driven by uncertainties in the initial soil organic carbon content. However, the rate of change of the active layer with temperature is spatially variable and dependent on CMIP5 model and its associated uncertainties need to be considered.

The other estimate of $\gamma_{PF}$ uses the bias-substituted $\text{ALT}_{\text{max}}$. In this case, $\gamma_{PF}$ is independent of biases in the CMIP5 estimates of present-day permafrost extent. The 5th–95th percentile range of the frequency distribution of $\gamma_{PF}$ is from $-6$ to $-66$ PgC K$^{-1}$ (Fig. 3). The two different RCP scenarios are pooled. This range is contained within the spread of values shown in Fig. 1. The median value of $-22$ PgC K$^{-1}$ falls toward the lower end of
the frequency plot. This is because the distribution is
skewed with a long tail of stronger feedbacks (larger
negative values). Although these larger permafrost feed-
backs have a relatively small likelihood of occurrence, they
cannot be ruled out. The 5th–95th percentile range
of carbon made vulnerable to release by 2100 is between
11 and 135 PgC for RCP4.5 and between 18 and 181 PgC
for RCP8.5. This is considerably lower than the estimate

**FIG. 2.** The relationship between $\gamma_{PF}$ calculated using the uncorrected $ALT_{max}$ for the RCP4.5 scenario and initial soil carbon content. The mean of the estimate of the initial soil carbon content for the JULES-simulated permafrost extent is shown by the vertical dark gray line and that of the IPA observations is shown by the vertical light gray line. Also shown is the linear regression (forced through zero) between the initial soil organic carbon content and $\gamma_{PF}$ (dashed dark line).

**FIG. 3.** Frequency distribution of $\gamma_{PF}$ estimated using the bias-substituted $ALT_{max}$. Both RCP scenarios are pooled. The value of $\gamma_{PF}$ is estimated using the reference estimate of present-day $ALT_{max}^{b}$ found from JULES-WATCH. The 5th–95th percentile range is shaded light gray.
by Harden et al. (2012) who found between 108 and 706 PgC may thaw by 2100 under RCP8.5.

As defined in this paper, $\gamma_{PF}$ is the committed change by the end of the twenty-first century. At this time not all of the thawed and decomposable soil carbon will have decomposed, but it is assumed that it will eventually decompose if the climate were to stabilize at that global mean temperature level. An alternative definition of the permafrost-carbon response, used by, for example Schneider von Deimling et al. (2012), is the amount of permafrost-carbon lost per degree temperature change—a realized $\gamma_{PF}$. This realized $\gamma_{PF}$ can be estimated for HadGEM2-ES using the methods developed by Burke et al. (2012). A comparison of the committed $\gamma_{PF}$ with the realized $\gamma_{PF}$ for HadGEM2-ES and the RCP4.5 scenario suggests that by the end of the twenty-first century approximately half of the vulnerable carbon will have been lost. Therefore, using this as a guide, a rough estimate of the committed response from Schneider von Deimling et al. (2012) is very similar for both RCP4.5 and RCP8.5 and between $-12$ and $-38$ PgC $K^{-1}$ (both 68% range). This is comparable with the committed $\gamma_{PF}$ estimated here ($-12$ to $-44$ PgC $K^{-1}$ also 68% range) with a similar uncertainty range. Using the Special Report on Emissions Scenarios (SRES) A1B high emissions scenario, Schaefer et al. (2011) estimated a realized loss of $110 \pm 40$ PgC and Koven et al. (2011) a realized loss of $62 \pm 7$ PgC by 2100. These can be roughly compared with the approximate realized carbon loss shown here of between 9 and 90 PgC (5th–95th percentile) by 2100 for RCP8.5.

Uncertainty assessment

Uncertainties in Fig. 1 are caused by our uncertain knowledge of the profile of soil organic carbon content, the spatial distribution of soil organic carbon content, the decomposability of the soil organic carbon content defined by the proportion of soil organic carbon that is active, and the model- and RCP-specific dependence of ALT$_{max}$ on temperature change. The relative contribution of each of these uncertainties to the overall uncertainty for $\gamma_{PF}$ derived using the uncorrected ALT$_{max}$ was estimated using Eq. (4) and is shown in Table 2 for the end of the twenty-first century. The continuous parameters were divided into four equally sized bins. There are only two bins representing the two RCP scenarios. Differences in the number of bins considered have a relatively minor impact on the relative contributions. Any differences between the CMIP5 models dominate the overall uncertainty in Fig. 1, contributing 64% to the total spread. The other main contributors are uncertainties in the soil organic carbon content at depths below 100 cm. Despite there being obvious systematic differences between the RCP scenarios in Fig. 1, they contribute a relatively small amount to the overall uncertainty.

Figure 4 explores the uncertainties by the end of the twenty-first century for the bias-substituted estimates of $\gamma_{PF}$. The black dashed line in Fig. 4 represents the mean for the whole ensemble (17,000 members) and the gray band is two standard deviations wide and centered on the mean. The black lines and error bars represent the means and standard deviations for subensembles that have been grouped by roughly equally sized bins around the values shown. Both the standard deviations and means change depending on subensemble (Fig. 4). In general, the bins that result in $\gamma_{PF}$ with the smallest magnitudes have narrower uncertainties and vice versa. This is because for these smaller permafrost-carbon responses there is little carbon made vulnerable and therefore less potential uncertainty on the total amount. Compared with Fig. 1 there are now small differences among the CMIP5 models. The permafrost-carbon response remains larger for RCP4.5 than for RCP8.5. Both of these differences arise because the rate of increase of vulnerable carbon per degree temperature change is nonlinear with temperature. It is generally lower at the larger temperature changes and deeper ALT$_{max}$. Schneider von Deimling et al. (2012) found similar values for both RCP scenarios because they did not explicitly represent the profile of soil carbon within the permafrost, which causes the notable differences here.

Table 3 shows the relative importance of the eight uncertainties addressed here. As in Table 2 they were calculated by dividing the continuous parameters into four equally sized bins. The SOCC below 100 cm now dominates the uncertainties in $\gamma_{PF}$. This causes approximately 59% of the overall uncertainty, with slightly more from the shallower soil than the deeper soil. Although the sensitivity of ALT$_{max}$ to local temperature change (ALT$_{sensitivity}$) and the potential biases in ALT$_{max}^{b}$ were both given a relatively wide range of values, they

<table>
<thead>
<tr>
<th>Uncertainty</th>
<th>Relative contribution to $\gamma_{PF}$ calculated using the uncorrected ALT$_{max}$ (Fig. 1) (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>RCP scenario</td>
<td>3</td>
</tr>
<tr>
<td>SOCC 0–100 cm</td>
<td>2</td>
</tr>
<tr>
<td>SOCC 100–200 cm</td>
<td>10</td>
</tr>
<tr>
<td>SOCC 200–300 cm</td>
<td>18</td>
</tr>
<tr>
<td>Decomposable proportion</td>
<td>3</td>
</tr>
<tr>
<td>of thawed carbon</td>
<td></td>
</tr>
<tr>
<td>CMIP5 model</td>
<td>64</td>
</tr>
</tbody>
</table>
only contribute \(-8\%\) of the overall uncertainty each. The RCP scenario, proportion of carbon that is decomposable, and the temperature sensitivities of the CMIP5 models each contribute 4\% or less to the total uncertainty. The larger overall contribution from the uncertainty in the SOCC compared with the smaller contributions from the diagnosis of the changes in ALT\(_{\text{max}}\) suggests that combining observational constraints with the relationship
Sensitivity of ALTmax to global CMIP5 temperature sensitivity from Friedlingstein et al. (2006).

Decomposable proportion SOCC (200–300 cm) 27
SOCC (100–200 cm) 32
SOCC (0–100 cm) 12

RCP scenario 4
2100 (Qian et al. 2010). Table 4 summarizes the estimated realized permafrost-carbon climate response for the RCP4.5 pathways. The value for γL for the C4MIP ensemble is taken from Friedlingstein et al. (2006).

4. Comparison with the land carbon climate response (γL)

Friedlingstein et al. (2006) found that the mean land carbon sensitivity γL (without permafrost) was −79 PgC K⁻¹ across the C4MIP generation of carbon cycle models under the high SRES A2 scenario for the period up to 2100 with a range of between −20 and −177 PgC K⁻¹. This represents a loss of carbon and results in a positive carbon climate feedback. This loss is dominated by extraboreal regions (Boer and Arora 2010) with the majority of models suggesting a net uptake in the northern high latitudes of up to 17 PgC by 2100 (Qian et al. 2010). Table 4 summarizes γL found using Eq. (5) from the available CMIP5 models for the present day.

5. Comparison with other biogeochemical feedbacks

To compare it with other feedbacks, γPF was converted to an equivalent radiative forcing by multiplying by φ.
(Arneth et al. 2010; Gregory et al. 2009), where $\phi$ is 0.0049 W m$^{-2}$ Pg C$^{-1}$ and represents the linear approximation of the increase in radiative forcing with increased atmospheric concentrations. Pooling all models for the RCP4.5 scenario and estimating the 5th–95th percentile range gives a $\gamma_{PF}$ between 0.03 and 0.45 W m$^{-2}$ K$^{-1}$. Similarly, for RCP8.5 $\gamma_{PF}$ is between 0.03 and 0.25 W m$^{-2}$ K$^{-1}$. As discussed previously, $\gamma_{PF}$ tends to be slightly lower for the RCP8.5 scenario. Arneth et al. (2010) estimated permafrost-carbon feedbacks to be similar in magnitude to those estimated here. Arneth et al. (2010) showed that other feedbacks such as wetland methane emissions, ozone emissions, and fire are likely to be smaller than $\gamma_{PF}$. This suggests the permafrost-carbon climate feedback is likely to be the largest biogeochemical feedback not currently included in coupled earth system GCMS.

6. Discussion and conclusions

The permafrost carbon–climate feedback is not yet included within the coupled earth system models used in the CMIP5 analysis. Therefore the technique discussed here provides a basic tool for quantifying the permafrost-carbon response and its uncertainty with respect to differing model structural and parametric elements. This paper uses a simple approach to provide a preliminary estimate of the carbon made vulnerable to decomposition through permafrost thawing per degree change in global mean temperature ($\gamma_{PF}$) for the CMIP5 global climate models. This is a committed response. Several estimates of $\gamma_{PF}$ are presented that suggest $\gamma_{PF}$ is likely to have a positive feedback on climate and is less than $-98$ Pg C $K^{-1}$ or $0.48$ W m$^{-2}$ K$^{-1}$. After correction for present-day biases in permafrost extent and active layer thickness, the best estimate of $\gamma_{PF}$ is between $-6$ and $-66$ Pg C $K^{-1}$ or between 0.03 and 0.29 W m$^{-2}$ K$^{-1}$ (5th–95th percentile range). This estimate is dependent on future emissions scenario because vulnerable organic carbon decreases with depth and is therefore nonlinearly related to global mean temperature change.

One of the main limitations of this study is that it uses highly simplified models to simulate the carbon cycle. In particular, it neglects interactions between the carbon currently within the GCM’s carbon cycle and any new carbon released from thawed permafrost. In addition, the GCMS do not simulate nutrient dynamics and cannot represent, for example, the impact of nitrogen availability on the carbon cycle. Northern high-latitude ecosystems are often nutrient limited and nutrients such as nitrogen contained within the permafrost may stimulate primary productivity once the permafrost thaws (Keuper et al. 2012). Another neglected process is thermokarst development. Bog formation might sequester some of the thawed permafrost-carbon and render it inert. Other additional missing processes include formation of ice wedges within the permafrost, cryoturbation, the presence of peat soils, mosses as a vegetation type, and the impact of nutrient availability on vegetation productivity. Time scales of decomposition and release of carbon to the atmosphere are not considered in this paper. Rates of change will impact the amount of carbon lost from the northern high latitudes. For a low rate of thaw the ecosystem is more likely to respond and utilize the thawed nutrients and carbon. If the thaw is quick, the carbon and nutrients are more likely to be washed into aquatic systems and lost. Field experiments are regularly carried out to develop understanding of many of these processes locally and in different ecosystems (e.g., Schuur et al. 2009), but increasingly upscaling methods are required to improve their representation at the larger scales, such as that of a GCM grid cell.

The current generation of coupled earth system GCMS represented by the CMIP5 model ensemble have large biases in their estimate of present-day permafrost extent (Koven et al. 2013; Slater and Lawrence 2013). These biases need to be significantly reduced while including the permafrost-carbon climate feedback within global climate models. Therefore, future work should focus on improving the accuracy of the coupling between the air temperature and the soil temperature in the northern high latitudes and fully evaluating the land surface schemes using methods such as those discussed in Burke et al. (2013) and Koven et al. (2013). This will help characterize the spatial variability in the active layer thickness and its sensitivity to a changing climate.

An assessment of the impact of uncertainties on the estimate of $\gamma_{PF}$ demonstrates the importance of knowing the amount of soil carbon in the soil below the present-day maximum thaw depth. Our knowledge of this can be significantly improved through improved observational-based analyses, such as that of Harden et al. (2012).

All of the Monte Carlo simulations discussed here show a loss of permafrost-carbon in the future. This loss is likely to be large enough to change the high latitudes from a potential sink to a source of carbon under future climate change. The permafrost-carbon climate feedback is probably the biggest biogeochemical feedback not currently included within coupled earth system GCMS.

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