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Permafrost thaw and resulting soil moisture changes regulate projected high-latitude CO₂ and CH₄ emissions

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Abstract

The fate of currently frozen permafrost carbon as high-latitude climate warms remains highly uncertain and existing models give widely varying estimates of the permafrost carbon–climate feedback. This uncertainty is due to many factors, including the role that permafrost thaw-induced transitions in soil hydrologic conditions will have on organic matter decomposition rates and the proportion of aerobic to anaerobic respiration. Large-scale permafrost thaw, as predicted by the Community Land Model (CLM) under an unmitigated greenhouse gas emissions scenario, results in significant soil drying due to increased drainage following permafrost thaw, even though permafrost domain water inputs are projected to rise (net precipitation minus evaporation >0). CLM predicts that drier soil conditions will accelerate organic matter decomposition, with concomitant increases in carbon dioxide (CO₂) emissions. Soil drying, however, strongly suppresses growth in methane (CH₄) emissions. Considering the global warming potential (GWP) of CO₂ and CH₄ emissions together, soil drying weakens the CLM projected GWP associated with carbon fluxes from the permafrost zone by more than 50% compared to a non-drying case. This high sensitivity to hydrologic change highlights the need for better understanding and modeling of landscape-scale changes in soil moisture conditions in response to permafrost thaw in order to more accurately assess the potential magnitude of the permafrost carbon–climate feedback.

1. Introduction

High latitude terrestrial warming is leading to soil warming and permafrost thaw, and is likely beginning to expose significant quantities of organic carbon (C) to decomposition by soil microbes (Schuur et al 2009, Schuur et al 2015). Continued warming will likely lead to large-scale losses of near-surface permafrost (Lawrence et al 2012, Koven et al 2013a, Slater and Lawrence 2013). Terrestrial carbon models indicate that the Arctic could shift from a sink to a source of carbon within the 21st century, as organic matter that was either previously frozen in permafrost or was effectively inert due to saturated soil conditions is decomposed and released to the atmosphere as carbon dioxide (CO₂) or methane (CH₄) (Koven et al 2011, Schaefer et al 2011, Burke et al 2012, Schneider von Deimling et al 2012, Koven et al 2015), thereby amplifying climate change. Though all models indicate that this carbon release is likely to be gradual and prolonged, uncertainty in the amplitude, timing, and form (CO₂ or CH₄) of the release remains high. The lack of agreement across models is due to the many poorly understood or unconstrained mechanisms and parameters that control permafrost thaw and subsequent organic matter decomposition. These uncertainties include: initial organic matter content (Hugelius et al 2013) and quality (Schädel et al 2014), nutrient dynamics (Bouskill et al 2014), depth dependence of decomposition rates (Koven et al 2015), vegetation feedbacks (Lloyd et al 2003, Lawrence and Swenson 2011, Bonfils et al 2012), heterogeneity in permafrost thaw and related hydrologic responses (Painter et al 2013), and the modulating role that...
changes in soil moisture and inundation will have on decomposition rates and partitioning into CO2 and CH4 emissions (Riley et al 2011, Elberling et al 2013, Melton et al 2013). Uncertainty related to the large-scale soil moisture response and its impact on decomposition rates is the subject of this study.

Permafrost exerts a strong control on soil hydrologic conditions, acting as a barrier to vertical water flow and thereby supporting perched water tables and saturated or near-saturated soils in the active layer. At the landscape scale, it remains poorly understood what will happen as the active layer and permafrost table deepen in response to warming. Multiple studies have detected changes in lake or wetland area which has been attributed alternatively to the melting of permafrost ice opening up pathways to the groundwater flow system (talk drainage, Smith et al (2005)), terrestrialization (floating mat encroachment, Roach et al (2011)), or natural precipitation variability (Plug et al 2008). Across several major Canadian and Alaskan Arctic rivers, there is compelling evidence for increases in groundwater contribution to base flow during low-flow conditions in winter, and these increases have been attributed to permafrost degradation (Walvoord and Striegl 2007, St. Jacques and Sauchyn 2009, Ye et al 2009). Bense et al (2009, 2012) showed, through idealized numerical experiments with a coupled hydro-thermal model, that disappearance of residual permafrost can cause a sudden establishment of deep groundwater flow paths and an increase in groundwater contribution to streamflow. However, permafrost thaw can also increase soil moisture in lowland regions as a result of localized ground slumping that results from melting ground ice. Even in upland areas that may experience large-scale soil drying, localized ground subsidence can result in collapsed areas that are saturated, interspersed with adjacent unsubsided and drier areas (Jorgenson et al 2001, Vogel et al 2009, O’Donnell et al 2012).

Changes in permafrost zone soil moisture are important because soil moisture (along with temperature) is a primary driver of Arctic carbon exchange (Oechel et al 1998, Shaver et al 2006, Oberbauer et al 2007). Field studies indicate that warming and drying leads to higher soil respiration rates than warming alone (Oberbauer et al 2007, Natali et al 2015). Elberling et al (2013) showed that permafrost soil carbon losses in laboratory incubations varied between 9% and 75%, depending on drainage conditions (higher drainage yielded greater carbon loss), indicating that the rate of carbon mobilization will depend strongly on future soil moisture conditions, and also that carbon in near-saturated conditions may remain largely immobilized. On the other hand, warming along with continued or increasingly saturated soil conditions (due perhaps to increasing precipitation, ground subsidence, or thermokarst) is likely to lead to increased CH4 emissions (Olefelt et al 2013), which would result in low total carbon loss, but a large impact on greenhouse gas (GHG) forcing driven by permafrost thaw because of the high methane global warming potential (GWP).

The purpose of this paper is to explore how projected changes in soil moisture, in response to permafrost thaw, could affect the total GHG forcing coming from the permafrost region. We utilize the Community Land Model version 4.5 (CLM4.5BGC). Under the strong projected warming in an unmitigated emissions scenario (figure 1, ∼+9°C by 2100 and > +15°C by 2300 averaged over the permafrost region), CLM4.5BGC predicts average soil warming at 1 m depth (Tsoil1m) of +7°C by 2100 and +13°C by 2300. The ‘permafrost region’ is defined here as the geographic area where CLM simulates at least one permanently frozen soil layer within the top 3 m of the soil column in 1850. Warming generates steadily deeper active layers and large declines of near-surface permafrost (~9 million km² lost by 2100; ~16 million km² by 2300). This deepening of the active layer and associated melting of permafrost soil ice allows water to drain deeper into the soil column, drying the near-surface soils, and effectively pulls the ‘bath plug’ from the soil column once permafrost melts to 3 m depth, allowing soil water to drain out of the soil column if the water balance does not support the presence of a water table within 3 m of the surface (figure 1). The large-scale soil drying occurs even with increasing net annual water inputs (trends in precipitation minus evaporation >0) and is the largest soil moisture change feature seen anywhere on the globe (figure S1).

The eventuality of CLM’s predicted large-scale soil moisture drying response to permafrost thaw is difficult to assess. But, CLM does produce a qualitatively similar increase in subsurface (groundwater) runoff and winter low flows after permafrost thaws (figure S2) as seen in observations (Ye et al 2009) and other models (Bense et al 2009). The changes in soil moisture have an impact on CLM’s simulated soil carbon turnover and methane production. In this study, we establish the extent to which permafrost-thaw induced soil drying, as predicted by CLM, alters GHG emissions from the permafrost zone and analyze whether drying is likely to lead to a stronger or weaker permafrost climate-carbon feedback.

2. Methods

2.1. The Community Land Model

We utilize the Community Land Model version 4.5 (CLM4.5BGC) under the same configuration and climate forcing as the permafrost climate-carbon feedback study of Koven et al (2015). CLM4.5BGC includes permafrost processes that allow simulation and projection of key thermal, hydrologic, and biogeochemical processes that are thought to be relevant for permafrost and its response to climate change (Lawrence and Slater 2007, Lawrence et al 2008, Riley...
et al 2011, Swenson and Lawrence 2012, Swenson et al 2012, Koven et al 2013b).

A complete description of CLM4.5BGC is included in Oleson et al (2013) and in the above references. Here, we introduce the soil hydrology and soil carbon turnover parameterizations that are of particular relevance to this study. We take advantage of recent improvements in the CLM representation of cold region hydrology (Swenson et al 2012) which generate an active layer hydrologic cycle that is qualitatively consistent with observations. Soils tend to be saturated after snow melt with a perched water table that slowly deepens throughout the summer. Migration of water towards the freezing front occurs in the autumn and an ice-rich transient layer builds at the base of the active layer, minimizing drainage to the regional groundwater system.

The saturated or near-saturated active layer and perched water table is maintained within CLM by impedance of vertical water flux through a strong reduction of hydraulic conductivity in the presence of soil ice. Specifically, hydraulic conductivity for each soil layer (K; mm s⁻¹) is calculated using an ice impedance factor with a power law form (Swenson et al 2012):

\[
K(z) = 10^{-F_{\text{ice}}(z)} K_{\text{sat}}(z) \left( \frac{\Theta_{\text{liq}}(z)}{\phi(z)} \right)^{B(z)+3},
\]

where \( F_{\text{ice}} = \frac{\Theta_{\text{liq}}}{\phi} \) is the ice-filled fraction of pore space in each soil layer, \( \Omega \) is a constant ice impedance factor, \( K_{\text{sat}} \) is saturated hydraulic conductivity (mm s⁻¹), \( \Theta_{\text{liq}} \) is liquid volumetric soil moisture content (mm³ mm⁻³), \( \phi \) is porosity (mm³ mm⁻³), and \( B \) is an exponent based on soil texture. As permafrost ice melts, \( F_{\text{ice}} \) decreases and hydraulic conductivity increases, allowing water to infiltrate deeper into the soil column. When the whole column to 3.8 m depth becomes ice free, water can drain into the unconfined aquifer or be diverted to sub-surface drainage.

Soil carbon turnover in CLM4.5BGC is based on vertical discretization of first-order multi-pool soil organic matter (SOM) dynamics (Koven et al 2013b) based on the Century decomposition model (Parton et al 1987):

\[
\frac{\partial C_i(z)}{\partial t} = R_i(z) + \sum_{j \neq i} \left( 1 - r_j \right) T_{ij} k_j(z) C_j(z) - k_i(z) C_i(z) + \frac{\partial}{\partial x} \left( D(z) \frac{\partial C_i}{\partial x} \right) + \frac{\partial}{\partial z} \left( A(z) C_i(z) \right),
\]
where $C_i$ is carbon in pool $i$ at soil level $z$, $R_i$ is the carbon input to that pool, $T_{ij}$ is the transfer of C due to decomposition from pool $j$ to pool $i$, $k_i$ and $k_j$ are the decay constants of pool $i$ and pool $j$, and $D$ and $A$ represent vertical carbon transport by diffusion and advection, respectively. $k_i$ is modified by the soil environment for all pools and is defined as

$$k_i = k_{0,i} r_T r_w r_O z,$$

where $k_{0,i}$ is an intrinsic pool-specific rate, $r_T$ is the direct temperature control ($Q_{10} = 1.5$), $r_w$ is the liquid moisture control, and $r_O$ is the oxygen control—when oxygen availability is insufficient to meet oxygen demands, respiration rates are scaled down, with a minimum oxygen limitation of 0.2 determined by decomposition rates in anaerobic soils. $r_z$ is a depth control, which is defined as

$$r_z = \exp \left( -\frac{z}{Z_r} \right)$$

with $Z_r$ a decomposition depth control parameter. This depth control of decomposition represents the net impacts of unresolved depth dependent processes including soil microbial population dynamics, pore-scale oxygen availability, mineral sorption, priming effects, and other unrepresented processes, which observations suggest reduce decomposition rates at depth beyond the limitations of temperature, moisture, and bulk oxygen availability. In this study, we primarily utilize $Z_r = 10$ m, which yields a weak additional depth dependence of decomposition beyond the environmental controls and, as discussed and evaluated relative to $Z_r = 1$ m and $Z_r = 0.5$ m in Koven et al. (2015), results in CLM permafrost-domain soil carbon stocks that are in closest agreement (1582 Pg for $Z_r = 0.5$ m, 1331 Pg for $Z_r = 1$ m, and 1032 Pg for $Z_r = 10$ m) with observed estimates (1060 Pg C to 3 m depth; Hugelius et al. 2013). We note that the relationship in equation (3), which implies multiplicative impacts of limitations to decomposition, is commonly applied in land biogeochemical models, but is quite uncertain.

The methane emission model includes representations of production in the anaerobic fraction of the soil, as well as oxidation within the soil column, aerenchyma transport, ebullition, aqueous and gaseous diffusion, and fractional inundation (Riley et al. 2011). Because CLM does not currently specifically represent wetland plant functional types or soil biogeochemical processes, gridcell-averaged decomposition rates are used as proxies. Thus, the upland (default) heterotrophic respiration is used to estimate the wetland decomposition rate after first dividing off the O$_2$ limitation. Although present-day simulated CH$_4$ emissions are within the wide range of inversion and bottom-up modeling estimates, they remain highly uncertain due to high parameter uncertainty, known structural limitations in CLM hydrology and biogeochemistry, and lack of relevant emissions such as those from lakes. Changes in CH$_4$ emissions do not account for changes in wetland or lake distribution that could occur due to thermokarst or other thaw-related landscape dynamics, (see Riley et al. (2011) for more discussion).

We utilize a model version with 30 ground vertical levels with a maximum depth of 45 m. The first soil layer has a thickness of 1.8 cm; below that, the thickness of the layers increase exponentially to a depth of 50 cm, after which they are held constant at 20 cm until 3.5 m depth. This increased resolution improves predictions of active layer and upper permafrost dynamics. The upper 22 layers down to 3.8 m are hydrologically active with the layers below representing bedrock.

2.2. Experimental design

We complete two offline CLM4.5BGC simulations for the period 1850–2300. The model is forced with time-varying meteorology, CO$_2$ concentration, nitrogen deposition, aerosol deposition, and land-use change as in Koven et al (2015). The historic atmospheric forcing data for the period 1901–2005 is taken from the CRUNCEP dataset (http://dods.ipsl.jussieu.fr/igcm/IGCM/BC/OOL/OL/CRU-NCEP/). The projection period forcing is calculated by applying monthly climate anomalies from a CCSM4 simulation (Meehl et al. 2012) for the Representative Concentration Pathway (RCP8.5, 2006–2100) and the Extension Concentration Pathway (ECP8.5, 2101–2300) scenarios on top of repeating 1996–2005 CRUNCEP meteorology. As noted in Koven et al. (2015), the CRUNCEP forcing dataset appears to be biased cold over the terrestrial Arctic compared to other reanalysis products and observed temperature products. The first simulation is a control simulation with the model configuration as described above. We label this simulation DRYSOIL to signify that permafrost zone soils dry out on average after permafrost thaws (figure 1).

The second simulation (WETSOIL) is identical to DRYSOIL except that we include changes to the model that prevent melted permafrost soil water from moving vertically or draining as permafrost ice melts. The objective of this experiment is to maintain initial soil moisture amounts in all permafrost layers throughout the length of the simulation while exerting minimal interference to other hydrologic or thermodynamic processes of the model. We achieve this by not allowing hydraulic conductivity (equation (1)) to rise above the annual maximum pre-industrial hydraulic conductivity. Specifically, we replace $F_{ice}$ from equation (1) with $F_{ice,imped}$:

$$F_{ice,imped}(z, t) = \max \left[ F_{ice}(z, t), F_{ice1850}(z) \right],$$

where $F_{ice1850}(z)$ is the annual minimum pre-industrial climatological ice fraction for the years 1850–1859 from the control (DRYSOIL) simulation. By implementing the constraint on vertical soil water...
flow in this way, we allow active layer hydrology to operate as normal throughout the annual cycle and throughout the simulation, subject to standard model processes that dictate surface soil water infiltration, soil evaporation, root water uptake, drainage from the perched water table, and seasonal melting and freezing of soil water.

An additional constraint is required to prevent water being lost via root uptake and transpiration. In the DRYSOIL experiment, if roots exist within any newly unfrozen soil layers, then plants are able to access and use this water for transpiration. This is undesirable in the WETSOIL experiment because by experimental design we are not allowing water to infiltrate down into newly unfrozen layers. Therefore, if water is drawn out from one of these deeper layers for transpiration, that water cannot be replenished by a vertical flux. Consequently, we impose an additional restriction in the WETSOIL experiment that roots can only draw water from the initial active layer.

By including these two constraints, we effectively maintain the initial soil water content in permafrost layers throughout the simulation (figure 2). The active layer, where soil hydrology is not modified, also tends to stay wetter in WETSOIL because water infiltration to deeper soil layers, which would normally occur after permafrost thaw, continues to be impeded throughout the simulation. Due to the coarse model resolution and the lack of processes such as lateral groundwater flow and explicit representation of uplands and lowlands, which will likely dictate the local soil moisture response to permafrost thaw, it is appropriate to consider these two simulations as approximate but not strict upper and lower bounds of soil moisture change in response to permafrost thaw.

Both simulations are initialized with the same 1850 initial conditions that were obtained via a multi-century spinup to equilibrium using repeated 1901–1920 meteorological forcing along with fixed pre-industrial CO2 concentrations, nitrogen and aerosol deposition, and land use, as per standard CLM spinup protocol (Koven et al 2013b).

3. Results

The projected changes in soil temperature, ice fraction, and soil moisture have a strong effect on the environmental scalars that control the model’s organic matter decomposition rates (figure 3). As discussed in Koven et al (2015), the direct temperature effect due to soil warming is relatively small. The stronger predicted changes in controls on decomposition are the liquid moisture availability, which is dependent on the
Figure 3. Left column: depth dependent trends in the temperature, moisture, and oxygen decomposition scalar trends and the product of all the decomposition scalars (including the depth decomposition scalar) and soil carbon for the DRYSOIL experiment. Right column: differences for each field (DRYSOIL–WETSOIL). Color bar in top row applies to first three rows. Results are averaged for the 1850 permafrost domain shown in figure 1.
unfrozen water content and therefore abruptly increases when soil ice melts, and oxygen availability, which becomes a weaker limitation when permafrost thaws and water drains from the soil. The overall impact on soil decomposition rates due to changes in soil environmental conditions is assumed in the model to be the product of these terms, along with the depth modifier \( r_d \) (bottom left, figure 3). Prior to the period in the 21st century when large-scale thawing of permafrost begins, decomposition rates at depths greater than 1 m are negligible (i.e., the product of scalar modifiers \( \sim 0 \)), while slow turnover occurs within the active layer. By the end of both simulations, decomposition rates have increased throughout the soil column, with the largest increases occurring at depths between 0.5 and 2.0 m. This increase in SOM decomposition rates leads to the losses of soil carbon shown in figure 1.

In the WETSOIL experiment, the prevention of soil drying alters soil decomposition scalar trends (figure 3). Wetter soils lead to slightly more soil warming, consistent with Subin et al (2013), but only by \( \sim 0.1 \, ^\circ C \) (not shown). Therefore, there is virtually no difference in the temperature decomposition scalar as a consequence of soil drying. But, soil drying strongly impacts both liquid moisture availability and oxygen availability. Drying drives down SOM turnover rates due to increased moisture limitation on decomposition, but this effect is countered at depth (>1 m) by an increase in oxygen availability due to reduced oxygen consumption by microbes and enhanced oxygen transport into the soil column prompted by greater pore space introduced when water infiltrates deeper into, and out of, the soil column.

The combined impact is for drying to increase SOM decomposition rates at depth (figure 3). Approximately 20 Pg C (9% out of 213 Pg C) more soil carbon is lost by 2300 in DRYSOIL compared to WETSOIL (figure 4). The increased soil carbon loss in DRYSOIL occurs mostly below 0.5 m depth and peaks around 2.5 m where the difference in decomposition rates is highest (figure 3). Near the surface, where the DRYSOIL simulation sees soil carbon gains (figure 1), the carbon gains are greater in DRYSOIL, despite smaller litter inputs, because drier surface conditions yield a stronger increase in moisture limitation than a reduction in oxygen limitation, which is not as dominant a constraint near the surface due to shorter oxygen diffusion paths. Higher soil carbon turnover rates under moist surface conditions are consistent with results found in litterbag studies (Hicks Pries et al 2013). Vegetation carbon stocks rise in both cases (DRYSOIL: +42 Pg C; WETSOIL: +48 Pg C) as vegetation responds to both higher atmospheric CO2 concentrations and more favorable environmental growing conditions. Vegetation carbon stocks do not rise as strongly in DRYSOIL because, towards the end of the simulation, drier soils begin to limit vegetation growth while also contributing to increased fire frequency and intensity. Increases in vegetation carbon stocks offset some of the losses in soil carbon such that the total ecosystem carbon change from the permafrost region in the DRYSOIL experiment is \(-167 \, Pg \, C\), which is a 26 Pg C (16%) greater loss than in WETSOIL. Changes in leaf and root litter inputs to the soil system between DRYSOIL and WETSOIL are minor compared to the changes in soil decomposition rates. For reference, changes in soil and vegetation carbon are provided in map form in figure S3.

For methane emissions, however, drying has the opposite effect. Soil drying slows increases in CH4 emissions through stronger reductions in CLM’s predicted inundated area as well as a general deepening of the water table that drives reduced anoxia in the upper levels of the unsaturated portion of the grid cell. In the DRYSOIL experiment, CH4 emissions from the permafrost zone rise by a factor of only \( \sim 1.5 \) (31 Tg CH4–C yr\(^{-1}\) in 2000, 45 Tg CH4–C yr\(^{-1}\) in 2300). Methane emission increases are 4.5 times greater in WETSOIL (figure 4). (Note: we remind the reader that the CH4 emissions and their response to permafrost thaw and soil moisture change are highly uncertain due to the complexity and heterogeneity of processes that dictate CH4 biogeochemistry).

From the perspective of total GHG contributions arising from permafrost thaw, relative soil drying
Table 1. Changes in CO2 and CH4 emissions. CO2e-C is the CO2-equivalent carbon emission.

| CO2 emission | DRYSOIL | WETSOIL | DRYSOIL–WETSOIL | DRYSOIL | WETSOIL | DRYSOIL–WETSOIL |
|--------------|---------|---------|------------------|---------|---------|------------------|
| Tg CO2e-C yr⁻¹ | +102 | +97 | +5 | +747 | +678 | +69 |
| [Tg CO2 yr⁻¹] | [+375] | [+357] | | [+2738] | [+2488] | |
| CH4 emission | | | | |
| Tg CO2e-C yr⁻¹ | +96 | +252 | −156 | +190 | +825 | −635 |
| [Tg CH4 yr⁻¹] | [+13] | [+33] | | [+25] | [+108] | |
| CO2 + CH4 | +198 | +349 | −151 | +916 | +1415 | −567 |
| Tg CO2e-C yr⁻¹ | | | | | |

(compared to the absence of drying) results in a much weaker increase in CH4 emissions compared to a slightly stronger increase in CO2 emissions. Assuming a GWP for CH4 of 28 (assuming 100 yr time horizon; Myhre et al. 2013) for the period 2050–2100, we find that soil drying strongly reduces total GHG emissions from the permafrost zone (presented here in units of Tg CO2 equivalent carbon per year). Combined CO2 and CH4 emissions are 198 Tg CO2e-C yr⁻¹ in DRYSOIL and ~75% higher at 347 Tg CO2e-C yr⁻¹ in WETSOIL (table 1). For the period 2250–2300, DRYSOIL emissions are 916 Tg CO2e-C yr⁻¹ while WETSOIL emissions are 55% higher (1415 Tg CO2e-C yr⁻¹, table 1).

4. Discussion

We have presented our results as large-scale averages for the entire permafrost domain, but the hydrologic responses to climate change and permafrost thaw are likely to be highly heterogeneous and dependent on many factors including the specifics of regional and local climate change, the presence or absence of local groundwater convergence or divergence and recharge, thermokarst activity and other thaw-related changes to the ground surface, subsurface geomorphology, surface slope and aspect, vegetation community response, and potentially several other factors (see Jorgenson and Osterkamp (2005) for discussion of some of the many complexities to be considered). The large differences in GHG emissions in DRYSOIL versus WETSOIL suggests that heterogeneous hydrological changes will lead to strongly divergent regional and local GHG emissions in response to permafrost thaw.

Beyond the many uncertainties related to model structure, scale, parameters, and the absence of potentially relevant processes such as landscape dynamics, vegetation successional processes, and excess ground ice there are relevant limitations related to how the simulations themselves were initialized and conducted. First, initial (frozen) soil moisture within permafrost is arbitrarily set at the beginning of the spinup simulation and remains static until permafrost thaws. For these simulations, volumetric soil moisture was initialized to 0.3 mm mm⁻³, which is the subjectively chosen soil moisture level utilized globally for standard CLM ‘cold start’ simulations. For most locations around the world, soil moisture levels at initialization is not critical since soil moisture reaches equilibrium within a few decades. But, with permafrost soil moisture the initial soil moisture level remains constant until permafrost thaws, potentially many centuries into a transient simulation. In the DRYSOIL experiment, this arbitrary choice of initial soil moisture is relatively unimportant since soil moisture in newly thawed layers quickly adjusts to a new equilibrium, though initial soil moisture will play a role in the timing of thaw via its effect on the latent heat required to thaw permafrost layers. However, in the WETSOIL experiment, these initial soil moisture levels are maintained throughout the simulation. Consequently, if permafrost soil moisture had been initialized at a higher level, we would likely have identified an even larger difference between the wet and dry soil simulations.

A second model setup choice that has an impact on the results is the value for the decomposition depth control parameter ($Z_\tau$). In the experiments described above, we use $Z_\tau = 10$ m (which effectively assumes that there is only weak explicit depth dependence on decomposition rates) because simulations using that value exhibited the closest agreement with observations for initial permafrost carbon stocks (Koven et al. 2015). But, $Z_\tau$ is a highly unconstrained parameter and there is evidence of a depth dependence on decomposition in some systems (Schmidt et al. 2011). To examine the sensitivity to this key uncertain parameter, we conducted a second set of simulations with $Z_\tau = 1$ m. The convolution of soil moisture change with different soil carbon distributions and deep soil decomposition rates yields quantitatively different results (table 2). As with the $Z_\tau = 10$ m, drying enhances soil respiration and limits increases in CH4 emissions. But, for the period 2050–2100 with $Z_\tau = 1$ m, soil drying actually shifts the permafrost domain from
Table 2. Same as table 1 except for experiments with $Z_r = 1\text{ m}$.  

|               | 2050–2100 | 2250–2300 |
|---------------|-----------|-----------|
|               | DRYSOIL   | WETSOIL   | DRYSOIL–WETSOIL |
| CO$_2$ emission | 2050–2100 | 2250–2300 |
| Tg CO$_2$e-C yr$^{-1}$ | $-71$    | $-171$    | $+125$    |
|              | $[-147]$  | $[-260]$  | $[+1948]$ |
| CH$_4$ emission | 2050–2100 | 2250–2300 |
| Tg CO$_2$e-C yr$^{-1}$ | $+64$    | $+434$    | $-559$    |
|              | $[+291]$  | $[+147]$  | $[+1948]$ |
| CO$_2$ + CH$_4$ | 2050–2100 | 2250–2300 |
| Tg CO$_2$e-C yr$^{-1}$ | $-6$     | $+820$    | $-434$    |
|              | $[+165]$  | $[+1255]$ | $[-260]$  |

being a net source of GHG emissions to a net sink (DRYSOIL: $-6$ Tg CO$_2$e-C yr$^{-1}$; WETSOIL: 165 Tg CO$_2$e-C yr$^{-1}$).

In some locations, permafrost soils are characterized by the presence of massive ice formations, ice wedges, and ice lenses. These potentially large water masses could affect local hydrological budgets if they melt. In these CLM simulations we ignore excess ice, but in a parallel modeling study using a version of CLM with a rudimentary parameterization of excess ice, it was shown that excess ice slightly slows permafrost thaw and increases soil moisture as excess ice melts. This increase in soil moisture, however, is only temporary, lasting a few decades at most before permafrost thaws to the bottom of the soil column and enhanced drainage removes the extra water (Lee et al. 2014). As with the prior caveats, including excess ice would quantitatively alter our results, but would likely not alter the overall conclusions.

In summary, we provide a first order estimate, utilizing a model in which observed permafrost hydrologic and biogeochemical processes are qualitatively captured, of the impact of soil moisture change on the amplitude of GHG flux response to permafrost thaw. Our results indicate that, as has been commonly anticipated, transitions in soil moisture conditions in response to permafrost thaw are likely to significantly alter subsequent CO$_2$ and CH$_4$ emissions. If permafrost thaw leads to large-scale drying of the landscape, as predicted by CLM, then we would expect higher soil respiration rates and higher CO$_2$ emissions, relative to a situation where soils do not dry. Increases in methane emissions, on the other hand, would be strongly suppressed if large-scale drying occurs. In the simulations examined here, we find that drying reduces by $>50\%$ the total GWP related to multi-century permafrost thaw, though uncertainties in the simulated soil moisture and CH$_4$ emission response to permafrost thaw are high. Depending on the parameter values assumed, disparate trends in soil moisture conditions can even change the sign from a weakly negative GWP with drying to a relatively strongly positive GWP without it for the end of the 21st century. This high sensitivity to soil moisture change highlights the need to better understand and model local and landscape-scale changes in Arctic hydrological conditions in order to more accurately predict the local, regional, and pan-Arctic carbon cycle response to permafrost thaw.

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