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Analysis of Permafrost Thermal Dynamics and Response to Climate Change in the CMIP5 Earth System Models

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ABSTRACT

We analyze global climate model predictions of soil temperature (from the Coupled Model Intercomparison Project 5 (CMIP5) database) to assess the models' representation of current-climate soil thermal dynamics, and their predictions of permafrost thaw during the 21^{st} century. We compare the models' predictions to observations of active layer thickness, air temperature, and soil temperature, and to theoretically-expected relationships between active layer thickness and air temperature annual mean and seasonal cycle amplitude. Models show a wide range of current permafrost areas, active layer statistics (cumulative distributions, correlations with mean annual air temperature and amplitude of seasonal air temperature cycle), and ability to accurately model the coupling between soil and air temperatures at high latitudes. Many of the between-model differences can be traced to differences in the coupling between either near-surface air and shallow soil temperatures, or between shallow and deeper (1m) soil temperatures, which in turn reflect differences in snow physics and soil hydrology. We compare the models to observational datasets to benchmark the permafrost-relevant physics of the models. The models show a wide range of predictions for permafrost loss: 2-66% for RCP2.6, 15-87% for RCP4.5, and 30-99% for RCP8.5. Normalizing the amount of permafrost loss by the amount of high-latitude warming in the RCP4.5 scenario, the models predict an absolute loss of 1.6 ± 0.7 million km^2 permafrost $^\circ C^{-1}$ high-latitude warming, or a fractional loss of 6-29 % $^\circ C^{-1}$.

1. Introduction

Permafrost is a critical component of high-latitude land and determines the character of the hydrology, ecology, and biogeochemistry of the region. There is widespread interest in the use of coupled atmosphere-ocean-land surface models to predict the fate of permafrost over the next centuries because: (1) permafrost contains the largest organic C reservoir in the terrestrial system (Tarnocai et al. 2009); (2) permafrost stability is primarily dependent on temperature; and (3) global warming is expected to be relatively larger over the permafrost domain due to arctic amplification processes (Holland and Bitz 2003). Thawing of permafrost soils over the next century (Lawrence and Slater 2005) may contribute a powerful greenhouse gas feedback due to microbial decomposition and release as CO_2 and CH_4 of the frozen soil C to the atmosphere (Koven et al. 2011; Schaefer et al. 2011). This feedback may also have operated during prior climate warmings (Ciais et al. 2012; DeConto et al. 2012).

Here, we analyze output from a set of Earth System Models (ESMs) (Table 1) that participated in the Coupled Model Intercomparison Project (CMIP5), (Taylor et al. 2009), to evaluate the permafrost model predictions against observations and theoretical expectations, and to compare the predicted fate of permafrost under warming scenarios. Because the models participating in this exercise do not include critical process representation needed to calculate the permafrost C budget itself, which at a minimum includes sufficient belowground vertical resolution in their biogeochemical component to distinguish between permafrost and active layer carbon pools (Koven et al. 2009, 2011), we do not attempt to calculate a permafrost C feedback here; instead we focus on the soil thermal environment and thaw predictions, which are represented in these models and can thus serve as a basis for calculating the possible range of feedback strength (Schneider von Deimling et al. 2012; Harden et al. 2012; Burke et al. 2012).

Th purpose of this paper is twofold: (1) to document the behavior, in comparison to observations, of the permafrost-relevant aspects of these models in the current climate, and (2) to compare the model predictions of future changes to permafrost under climate change. By providing a framework for assessing realism of the models, we hope to lay a foundation for benchmarking the frozen-soil physics of these models, and which can then serve to inform future development (Luo et al. 2012). By doing this in the context of an intercomparison of future predictions, we seek to analyze how model differences that can be seen in the current climate affect the future response.

A number of authors have developed high-latitude-specific models of the exchange of energy and water to study the behavior of soil freeze and thaw processes. These models were initially developed for local and regional studies (Romanovsky and Osterkamp 1997; Hinzman et al. 1998; Shiklomanov et al. 2007; Rinke et al. 2008; Nicolsky et al. 2009). Many of the relevant processes, including the specific thermal and hydrological properties of organic soils, have been incorporated into global models (Nicolsky et al. 2007; Lawrence and Slater 2008; Schaefer et al. 2009; Koven et al. 2009).

We focus on the CMIP5 models as a representative set of global coupled models that are being used as an integral component of the Intergovernmental Panel on Climate Change Fifth Assessment Report (IPCC-AR5). The CMIP5 project included a large number of simulation experiments, including testing model response to a range of forcings, decadal predictability experiments, control scenarios, and paleoclimate experiments. To evaluate the high-latitude thermal predictions of the models, we analyze three future warming scenarios, RCP2.6, RCP4.5, and RCP8.5, which correspond to 2.6, 4.5, and 8.5 W m^{-2} forcing by 2100, respectively, and thus represent low, intermediate, and high warming scenarios (Taylor et al. 2009). We examine the ability of the CMIP5 models to simulate relevant aspects of the currently frozen soil thermal dynamics, and how these dynamics may change under the set of warming experiments. While many such numerical experiments have been conducted using regional permafrost models forced by atmospheric dynamics, it is useful to look to this large, state-of-the-art sample of ESMs, which includes a broad set of climate sensitivities, arctic amplification factors, and detailed land-atmosphere coupling, along with a clearly prescribed experimental design and forcing perturbation, to better understand the range of possible model-predicted permafrost fates under different global warming scenarios. A similar analysis, focusing on the changes to the distribution of climatological metrics known to influence permafrost extent, is being conducted by Slater and Lawrence (In review).

A simplified schematic of temperature dynamics for northern soils (fig. 1), shows that the soil temperature annual cycle is driven by changes in the radiative forcing and surface heat exchange, such that the amplitude of the seasonal cycle is greatest in the air, decreases across the air-soil interface, and decreases further with depth into the soils following a roughly exponential profile. The active layer in permafrost soils is defined as the maximum depth at which the annual temperature wave causes the soil to thaw on a regular basis (at least every other year). Coupling between environmental conditions, thermal properties, phase change, ground ice, and cryoturbation make the actual temperature dynamics of permafrost soils more complex than can be represented by simple diffusive energy transport. Across the air-soil interface, snow acts to insulate during the winter but not during the summer, leading to thermal rectification and warmer mean soil temperatures than mean air temperatures. Within the soil column, the low thermal diffusivity of organic soil horizons and the large amount of latent heat required to freeze and thaw moisture in the active layer water leads to rapid attenuation of the annual temperature wave. In addition, the differences between frozen and thawed soil thermal conductivities, particularly for organic soils which are good insulators in the summer but allow heat to escape during the winter, lead to further change in the mean temperatures with depth, though with a cooling rather than a warming effect(Romanovsky and Osterkamp 1997).

The CMIP5 models represent these processes very differently, both conceptually and numerically. For example, snow insulation may be treated either as a separate layer or layers existing above the soil column ("bulk" or "multi-layer" snow schemes in the classification of (Slater et al. 2001)) or as a transient replacement of the upper soil column with snow-like properties ("composite" or "implicit" schemes (Slater et al. 2001)). The representation of soil physical properties differ greatly as well, with some models including the effects of organic mater (e.g., those with CLM4 as their land model: CCSM4, CESM1-CAM5, and NorESM), while the majority analyzed here use only mineral soil properties. The coupling between thermal and hydrologic states in the models differs as well, with some models not including a latent heat term for soil moisture freeze-thaw processes. The model vertical discretization for soil thermal calculations varies widely between these models as well, as do the mechanics of coupling between the land surface and atmosphere. However, rather than enumerating the differences between the models, our focus here is on diagnosing the net behavior of the different models under current conditions, and how that behavior is linked to their predictions of permafrost thaw over the 21st century.

2. Methods

a. Analysis of CMIP5 models

We calculate the active-layer thickness (ALT) from the model predictions using monthly-mean soil temperatures (T_s). Some models in the CMIP5 experiment do not report depth-resolved soil temperatures and thus we do not include those models in this analysis. We calculate monthly-mean thaw depth as the deepest point in the soil column of a given gridcell at a given month with soil temperature at or above freezing. Given the coarse vertical discretization of land-surface models, thaw depth can be defined multiple ways, e.g., as the lower edge of the deepest thawed layer, (Lawrence and Slater 2005) or alternatively by interpolating soil temperature between model level centers and calculating the depth that the interpolated line intersects the freezing point (Lawrence et al. 2012). Here, we use the former (level edge) approach, and define the freezing point as 0 °C. The use of this single temperature threshold may introduce errors in some models due to artifacts in their latent heat parameterizations; this will be discussed in more detail below. We then calculate annual ALT as the maximum monthly thaw depth for a given year. We define permafrost to be present in a gridcell if the maximum annual ALT is shallower than either 3 m or the deepest model soil level, whichever is less; this approach therefore gives a metric of "near-surface" permafrost (Lawrence and Slater 2005).

In order to diagnose controls on permafrost distribution within the models, we compare modeled ALT with the local climate. Here we use the monthly-mean modeled surface air temperature, and examine two quantities that we hypothesize control permafrost distribution in the models: the annual mean temperature and the amplitude of its seasonal cycle. In particular, we are interested in how the propagation of energy in the soil leads to vertical differences in the annual mean and seasonal cycle amplitude of soil temperatures. To examine the amplitude of the seasonal cycle, we use a fourier analysis to calculate the amplitude of the annual frequency of the monthly-mean surface air temperature and soil temperature at 0 m and 1 m depth. At mid and high latitudes, the majority of the variance is contained in the annual wave (Stine et al. 2009), so we neglect higher frequency components. For all model runs, we use the first 10 years of the RCP4.5 climate scenario (2006-2015) for this analysis in order to compare against recent observations, and average across multiple ensemble members where possible.

For each model, we calculate the change in mean temperature and the amplitude of the seasonal cycle across two vertical gradients: the atmosphere to shallow soil interface and the change from 0 m to 1 m depth. Separating the atmosphere to deeper soil thermal connection into these two gradients has the advantage that we can isolate the processes operating across each region. The seasonal cycle response across the air to soil surface interface is mediated by snow insulation, radiative processes, and coupling between the atmospheric boundary-layer and the soil surface. The shallow to deeper soils gradient is dominated by soil hydrology, latent heat, and thermal properties. An exception to this is for some models which place snow insulation effects within the soil column. Similarly, while the mean temperature and the amplitude of the seasonal cycle will be linked at a given position along these vertical gradients, varying process representation in different models may lead to different levels of thermal rectification associated with the multiple processes operating across each gradient.

The soil vertical grids differ between models. So, to compare them, we need to interpolate predicted T_s to a uniform reference depth. To do this, we assume that the seasonal-cycle

T_s amplitude will attenuate roughly exponentially (fig. 1), following standard Fickian dynamics, while vertical differences in mean T_s will be roughly linear. Thus, we log-transform the amplitude so that it will be roughly linear with depth, then interpolate to the 1 m reference depth, and take its exponential. For mean temperatures we perform a simple linear interpolation to 1 m depth. The mean temperature differences are calculated as:

$$\Delta \bar{T}_{0m-1m} = \bar{T}_{1m} - \bar{T}_{0m} \tag{1}$$

$$\Delta \bar{T}_{air-soil} = \bar{T}_{0m} - \bar{T}_{air} \tag{2}$$

where $\bar{T}_{1m}$, $\bar{T}_{0m}$, and $\bar{T}_{air}$ are the mean temperatures at 1 m, the soil surface, and air, respectively. We report the seasonal cycle amplitude attenuations, α_{0m-1m} and $\alpha_{air-soil}$, as:

$$\alpha_{0m-1m} = \frac{\hat{T}_{1m}}{\hat{T}_{0m}} \tag{3}$$

$$\alpha_{air-soil} = \frac{\hat{T}_{0m}}{\hat{T}_{air}} \tag{4}$$

where $\hat{T}_{1m}$, $\hat{T}_{0m}$, and $\hat{T}_{air}$ are the corresponding amplitudes of the seasonal cycle.

b. Analysis of site observations

To compare modeled active layers with observations, we use two ALT datasets: the Circumpolar Active Layer Monitoring Network (CALM, (Brown et al. 2000)), and a separate analysis of historical ALT derived from soil temperature measurements at 31 Russian sites (Zhang et al. 2006). We also compare modeled soil temperatures directly with observations of soil temperatures using two datasets: (1) the International Polar Year - Thermal State of Permafrost (IPY-TSP, Romanovsky et al. (2010); Romanovsky (2010)), and (2) the Historical Russian Soil Temperature (HRST, Gilichinsky et al. (1998); Zhang et al. (2001)).

The IPY-TSP data are measured at multiple depths; here we use only sites that have at least one complete annual cycle at 3 depths between the surface and 1.5 meters. The HRST data are measured at a variety of depths, but the majority of sites have 20 cm as their shallowest depth. The mean temperatures generally show a linear and the seasonal cycle amplitudes an exponential relationship with depth, allowing interpolation to the reference levels. For both soil temperature datasets where temperature is not reported at the levels of interest (0 m and 1 m), we perform a linear regression of the mean temperatures as a function of depth and project it to 0 m and to 1 m. For the annual cycle amplitudes, we use the same approach but with log-transformed amplitudes.

An important caveat needs to be taken into account with regards to the HRST data. These measurements were generally made on bare soils in which surface organic layers had been removed (Gilichinsky et al. 1998), thus we expect these observations to underestimate the magnitude of the seasonal cycle attenuation and cooling with depth for these soils. These patterns are evident in the means for the two data collections: the mean $\Delta\bar{T}_{0m-1m}$ for the IPY-TSP data is -0.66 °C vs -0.31 °C for the HRST data, while the mean α_{0m-1m} is 0.38 for the IPY-TSP data and 0.45 for the HRST data. However, we use both datasets here because their spatial domains are different and complementary: the HRST observations are all in Russia, while the IPY-TSP data have panarctic coverage but are focused in Alaska. The difference in spatial coverage is similar for the two active layer datasets: CALM has a broad coverage but little representation of interior Siberia, while the Zhang et al. (2006) data are focused on interior Siberia.

We compare both the active layer and soil temperature data to atmospheric temperature data. Many of the IPY-TSP sites report measurements of the local surface air temperature.

We use these data where available, and otherwise use climatological means and seasonal cycles in air temperature from the corresponding gridcell of the CRU TS3.1 surface air temperature climatology (Mitchell and Jones 2005).

3. Results and Discussion

a. Comparison of high-latitude soil thermal dynamics under current climate

Simulated current-climate permafrost extent varies widely across the models, and generally between the models and the observation-based map of Brown et al. (1998) (fig. 2). Since the models generate their own atmospheric climatology, which could partly explain differences in permafrost area, we also show the position of the zero-degree isotherm in the surface mean annual air temperatures (MAAT, blue line) for each of the models and in the observations (final) panel, the CRU data. The permafrost distributions of Brown et al. (1998) contain 11.0 and 4.3 million km^2 of continuous and discontinuous permafrost, respectively, for a total of approximately 15.3 million km^2. The models calculate widely divergent total permafrost area under current climate (Table 2). If it were the case that the permafrost differences were caused by differences in predicted climate, the models would show a similar spacing between the permafrost edge and the zero-degree isotherm. Instead, this spacing varies widely between the models, indicating that the differences in permafrost extent lie fundamentally in the modelled soil thermal regimes or in the atmosphere to soil energy exchanges.

In addition to total permafrost area, another crucial component of a given model's permafrost dynamics are the predicted active layer depths. The simple permafrost temperature schematic (fig. 1), suggests that a model's predicted active layer at a given location should be controlled by the mean annual air temperature and the amplitude of the seasonal cycle, with warmer locations or larger seasonal cycles corresponding to deeper active layers. Fig. 3 shows these relationships for each of the models, and also for the combined active layer (Brown et al. 2000; Zhang et al. 2006) and atmospheric climatology (Mitchell and Jones 2005). The majority of models show a positive relationship between warmer climate and larger-amplitude seasonal cycles with deeper active layers, although the slopes of these relationships, as well as the fraction of total ALT variance explained by climate, differs between the models and between the models and the observations. As a simple first-order approximation of the relative role of climate in determining ALT between the models, we regress the variables assuming a relationship of the form:

$$Z_{thaw} = a\bar{T}_{air} + b\hat{T}_{air} + c \tag{5}$$

where Z_{thaw} is the active layer thickness, $\bar{T}_{air}$ and $\hat{T}_{air}$ are the mean and seasonal cycle amplitudes of surface air temperature. While the observations are consistent with the general relationships (Table 3), the air temperature accounts for a much smaller fraction of total variance (r^2=0.13) than it does for the simulations (r^2=0.22-0.84, with mean of 0.5). Thus the observations support the idea that factors other than climate, likely including soil conditions and fine-scale hydrology, account for a large fraction of this variance, while the models, which do not include these fine-scale controls, attribute too much of the ALT variance to climate. However, for this analysis, we have restricted the ALT observations only to high-latitude sites, which may bias our results away from a climate control since doing so excludes the low-latitude, high altitude ALT sites, which do show a stronger climate control. Several of the models show a convex-downwards trend to the active layer thickness with increasing temperature, or a bimodal regime with a shallow cold-permafrost slope and a steeper warm permafrost slope. Those with a distinct bimodal slope regime (e.g. CanESM, HadCM3, HadGEM2) all have relatively fewer model levels, suggesting that this pattern is an artifact of their limited vertical resolution.

Analytical solutions exist for the 1-D heat conduction with phase change subject to periodic upper boundary conditions problem, given simplifying assumptions: the Stefan equation (which assumes frozen soil is initially at 0 °C, and that the latent heat of fusion dominates the heat budget), and the Kudryavstev equation (which allows an initial permafrost mean annual temperature less than 0 °C) (Romanovsky and Osterkamp 1997; Riseborough et al. 2008). For qualitative comparison, we include panels in fig. 3 with the predicted active layer from these two equations given a single set of reasonable soil physical parameters: (unfrozen soil conductivity = 0.6 W m^{-1} K^{-1}, porosity = 0.25, assuming saturated soils), and climate parameters (we use a simplified climate representation for this exercise: CRU climatological $\bar{T}_{air}$ and $\hat{T}_{air}$, assuming a uniform 3 degree thermal offset $\Delta\bar{T}_{air-soil}$, no thermal offset $\Delta\bar{T}_{0m-1m}$, and a uniform attenuation coefficient $\alpha_{air-soil}$ of 0.6). The Kudryavstev equation shows the same basic pattern that the numerical models are capturing, i.e., active layers increase steeply near the permafrost edges, a concave-downward profile between ALT and MAAT, and an increasing ALT with increasing seasonal cycle amplitude. For the set of parameters applied here, the Kudryavstev ALT are shallower than predicted in the CMIP5 models, although other parameter choices can lead to deeper ALT, while maintaining the same functional form. In contrast, the Stefan equation predicts generally larger ALT than predicted by the Kudryavstev equation and a slightly concave-upward profile. Thus, the more complex Kudryavstev model supports those models which predict (1) smooth concave-downward profiles and (2) clear impacts of the temperature seasonal cycle amplitude and mean on ALT. Further, the Kudryavstev model supports the large observed variability in ALT resulting from site-specific differences in soil and snow physical properties.

The distributions of ALT and permafrost area under historical, current, and future climates vary greatly between the models (fig. 4). This statistic of cumulative ALT distributions is relevant to calculating the C feedback effect associated with permafrost thaw, as the difference between successive curves under a climate warming scenario is proportional to the soil volume transferred from permafrost to active layer, and thus the quantity of organic carbon made vulnerable to decomposition as a result of thawing (Harden et al. 2012). Most of the models show ALT distributions under future climate scenarios with a shape roughly the same (though with smaller total magnitude) as under current climate, though some models (e.g., GFDL, MIROC5) show an increase in the relative abundance of deeper active layers, presumably related to a slowed transient downwards thaw. Almost all models predict that some permafrost has already thawed during the 20th century (the difference between 2005-2010 and 1850-1859 curves at 3 m depth in fig. 3), varying between 3% gain (HadGEM2-CC) and 49% loss (BCC-CSM1-1) in permafrost area. While it is not clear how much permafrost thaw has occurred during the 20th century, observations do not support permafrost losses on the high end of this spectrum (Burn and Nelson 2006).

Although the available ALT observations are not evenly distributed throughout the permafrost region and thus are not available as a quantitative comparison against model predictions, we include the cumulative distributions from the observational datasets as a qualitative reference for the overall shape of these distributions. The observations show a broad range of active layer depths, though unlike the models, none of the sites have mean ALT less than ~20 cm—however this may again be due to sampling bias avoiding the coldest environments where ALT approaches zero.

Many of the models show step-like distributions in ALT (fig. 4), associated with the boundaries between the model levels of their finite-difference discretizations. These step-like patterns result from a tendency of a given model level to get stuck at 0 °C due to the large latent heat threshold required to transform the entire soil level's mass across the freeze/thaw boundary. In addition, some of the models show unrealistic behavior with respect to shallow active layers, with either too much (e.g., INM-CM4) or too little (e.g., MPI-ESM-LR, CanESM2, IPSL-CM5) of the permafrost area having shallow active layers. As discussed below, these are related to differences in the prescription of the latent heat of fusion of soil water for the models. Differences between the models' ALT predictions can be roughly quantified by calculating the model's median ALT, which varies from almost zero to > 3 m (Table 3). While the non-random spatial distribution of ALT observations does not allow us to calculate a rigorous observational constraint, the definition of gelisols (permafrost-affected soils) used in USDA Soil Survey Staff (1999) of less than 1 m (or 2 m if other evidence such as cryoturbation is present), rules out plausible values far outside of this range, such as .015 m in INM-CM4 at one extreme or 2.6-3.2 m for the IPSL-CM4 or MPI-ESM models at the other extreme.

The models occupy different subsets of the possible phase space between climate and ALT (fig. 3). If a given model's ALT equilibrates rapidly to a change in climate, then time trajectories of ALT in individual gridcells as a function of climate would have comparable slopes to the relationship between comparable climatic conditions across space (i.e., a "space for time" relationship). We qualitatively searched for such a relationship in the models by plotting lines connecting the predicted current (2005-2010) and future (2090-2099 for RCP4.5) ALT (fig. 5). Across the models, the slopes of these time trajectories are similar to those across space under current climate (fig. 3), suggesting that the models rapidly equilibrate their predicted ALT to a new climate, at least in comparison to the centennial timescale used to calculate these differences. The implication of this is that models which show a high sensitivity of current-day climatic control of ALT will also show a high sensitivity of ALT to warming. Since the models tend to overestimate, relative to the observations, both the slope of the spatial MAAT-ALT relationship and the fraction of ALT variance explained by climate (Table 3), it may be that the models are thus too sensitive in their predicted ALT response to climate change.

In order to diagnose the thermal dynamics responsible for the differences between the CMIP5 models under current climate (2006-2015), and to evaluate which models have a more realistic permafrost response to climate warming, we next discuss an analysis of how the models propagate temperature from the air through the upper soils, using the metrics discussed above: $\Delta\bar{T}_{air-soil}$, $\Delta\bar{T}_{0m-1m}$, $\alpha_{air-soil}$, and α_{0m-1m} in figs. 6, 7, 8, and 9.

The majority of models show a positive thermal offset ($\Delta\bar{T}_{air-soil}$, fig. 6) over the high-latitude region, which is linked to the strong attenuation of the seasonal cycle amplitude ($\alpha_{air-soil}$, fig. 7). As discussed above, the warming and attenuation are primarily due to the presence of snow, which can be seen in the models since this effect is confined mainly to the boreal and arctic regions, with a maximum that generally spans the boreal belt. The magnitude of the warming differs between models, with mean $\Delta\bar{T}_{air-soil}$ at the gridcells corresponding to observations of -0.2 to 8.8 °C and mean $\alpha_{air-soil}$ of 0.29 to 1.05. The two statistics of $\Delta\bar{T}_{air-soil}$ and $\alpha_{air-soil}$ are highly correlated between the models, with an r^2 of 0.8. Across the air-soil interface, the observations also show a pronounced warming (mean $\Delta\bar{T}_{air-soil}$ of 6.2 °C) in the mean temperatures for the shallowest soils relative to the surface air temperatures, and significant attenuation (mean $\alpha_{air-soil} = 0.57$) of the annual cycle.

Between the shallow and deeper soils, the models show a much smaller temperature gradient ($\Delta\bar{T}_{0m-1m}$, fig. 8), and twelve of nineteen show a general cooling with depth. The change in amplitude through the top meter of soil (α_{0m-1m}, fig. 9) also shows a strong attenuation (i.e. low values of α_{0m-1m}) throughout the high latitude region, although its magnitude and spatial distribution varies between models. This signal also shows strongest attenuation across the boreal belt, though the location of the minimum in α_{0m-1m} is typically slightly to the south of the maximum in the air-soil case. In principle, this attenuation should be strongest where the thermal diffusivity (the ratio of the thermal conductivity to the heat capacity) is lowest; if we assume, following the Stefan equation, that the latent heat of fusion of soil water dominates the heat capacity term, then the attenuation should be strongest where the most water changes phase, and thus strongest where active layers are as deep as the reference depth and porosity is high. Given that the change in mean temperature with depth through the soils is mostly due to the differing thermal conductivity between frozen and unfrozen soils (Romanovsky and Osterkamp 1997), this term will be very model dependent, but should be strongest (i.e. most negative values of $\Delta\bar{T}_{0m-1m}$) where soil porosity and water content is highest. Unlike for the air-soil interface, the multi-model means in $\Delta\bar{T}_{0m-1m}$ and α_{0m-1m} are only weakly correlated across the models, with an r^2 of 0.15. A less strong mean attenuation across the models is also correlated with deeper median active layer thickness

($r^2 = 0.22$, $p = 0.05$). The observations show a pronounced attenuation throughout the permafrost region, and a small cooling in the mean soil temperature.

In both the air-to-soil and the 0 m-to-1 m temperature changes, many of the models fail to reproduce the observed behavior, with many showing less attenuation in the annual cycle amplitude with depth (larger values of α_{0m-1m}), and larger temperature changes (with both warming and cooling with depth predicted in different models). Several of the models, such as the MOSES/TRIFFID land model in HadGEM2, the Sechiba/ORCHIDEE land model in IPSL-CM4 (the CMIP5 version of which predates the frozen soil developments in Poutou et al. (2004); Koven et al. (2009)), and the JSBACH land model in MPI-ESM-LR, show very little warming from the air to the soil, or even cooling, over the high northern latitudes ($\Delta\bar{T}_{air-soil}$, fig. 6), limited attenuation from air to the soil ($\alpha_{air-soil}$, fig. 7), and warming with depth through the soil instead of cooling ($\Delta\bar{T}_{0m-1m}$, fig. 8). For the HadGEM2, IPSL-CM4, and MPI-ESM models, this lack of attenuation in the $\alpha_{air-soil}$ term is due to the implicit and composite snow treatments, respectively (in the sense of Slater et al. (2001)), which leads to the models inserting the snow thermal effects between the shallow and deeper soils rather than between the atmosphere and shallow soils. However, differences even among these simplified models are evident—HADGEM2 replaces only the top layer with snow properties, while IPSL-CM4 uses snow thermal properties to the depth of the calculated snow thickness, leading to the very different $\Delta\bar{T}_{0m-1m}$ and α_{0m-1m} responses, and consequently the different permafrost extent and ALT distributions (figs. 3 and 4), between these two models. One advantage in separating the modelled temperature responses into the four components used here ($\Delta\bar{T}_{air-soil}$, $\Delta\bar{T}_{0m-1m}$, $\alpha_{air-soil}$, and α_{0m-1m}) is that there are possible tradeoffs, such as too-cold soils with too-large amplitudes at depth, that could lead to the same ALT; the separation described here allows for a clearer sense of what controls the permafrost extent and ALT.

A major difference between the models is in their treatment of the latent heat of fusion of soil water (table 1 and fig. 10), which includes: omission, apparent heat capacity over a discrete temperature range, or more detailed thermodynamic treatments of supercooled moisture. A result of these differences can be seen graphically (fig. 10) by plotting histograms of soil temperatures across a range spanning the soil freezing point. Models that omit latent heat (IPSL-CM5 and MPI-ESM) have a flat distribution; those that include latent heat terms would be expected to show a higher frequency of occurrence through the range that the latent heat term is applied as a given model gridcell soil level should get stuck at those temperatures for a longer duration during the passage of the seasonal cycle—this is one result of the "zero curtain effect" (Outcalt et al. 1990). The depth at which the zero-curtain effect occurs most strongly also differs between the models, and between models and observations. The IPA-IPY and HRST observations, averaged to monthly values and aggregated across all sites within each dataset, show an asymmetric, negatively-skewed and relatively gradual peak mainly below the freezing point, as significant unfrozen water exists and continues to freeze well below 0°C. Some of the models (e.g., CLM4) follow this pattern, while others show more sharply peaked distributions centered at a given temperature (which ranges from -2°C to just above 0°C), or multiple peaks corresponding to the boundaries of a discrete range over which an apparent heat capacity is applied (e.g., MRI-CGCM3). For the analysis throughout this paper, we have used a cutoff of 0°C for defining the boundary between frozen and unfrozen soil and therefore for permafrost; however this approach may lead to artifacts in models that apply the latent heat term away from zero, such as BCC-CSM1.1

which clusters just above 0°C, or the GFDL models which apply freezing at -2°C.

The amount of coupling between the atmosphere and soil surface, and the mediation of this coupling by snow, has a large impact on the differences between the modelled soil thermal environments. All of the models that simulate high current-climate permafrost extent ($> 17 \times 10^6$ km^2) also have low values (< 2 °C) of $\Delta \bar{T}_{air-soil}$, suggesting a first-order control on the modelled permafrost distribution by the air-to-soil thermal offset.

b. Model Evaluation

The widely divergent model behavior for these comparisons reflects several underlying causes, including (1) the level of process detail represented in the models; (2) parameter choices; (3) and degree of model calibration. The CMIP5 models are a suite of global atmosphere-ocean-land climate models, which must of necessity represent the huge complexity across the Earth System. We note that many of these models were not specifically developed to represent permafrost systems, although most modeling groups are actively working to improve this aspect of model performance.

Comparing observations against model predictions for the metrics defined above indicates no clear ranking of the models (Tables 2 and 3). For the soil thermal comparisons (Table 3), the rankings based on the modelled mean response at the observation sites are largely the same as the rankings based on the RMS error, showing that it is the inter-model differences in the mean response that dominates the RMS term rather than the model differences in the inter-site correlation. This conclusion is supported by the fairly uniform high-latitude signals in the thermal metrics (figs. 6, 8, 7, and 9).

c. *Comparison of modeled permafrost response to warming*

While all of the models show some loss of permafrost under 21^{st} century warming, the range of responses is large for all RCP scenarios (fig. 11, table 2). There are two ways of looking at the changes to modelled permafrost extent: as absolute changes to the permafrost area or, given that there are such large differences in the initial permafrost distributions between the models, to look at the fractional changes to permafrost area. If we look at the absolute changes, then they range from 0.1-5.2 million km^2 for RCP2.6, 0.6-8.0 million km^2 for RCP4.5, and 0.9-15.7 million km^2 for RCP8.5. The fractional loss in permafrost extent between 2005-2100 ranges from 2-66% for RCP2.6, 15-87% for RCP4.5, and 30-99% for RCP8.5. Despite these large ranges and their implied model uncertainty, the range of model responses can be used to offer some implications for permafrost under climate change: e.g., these models predict much more drastic losses in permafrost under the high warming scenarios, RCP4.5 and especially RCP8.5, than under the low warming RCP 2.6 scenario.

Given that the CMIP5 models are all fully coupled land-atmosphere-ocean models, and our analysis only covers the surface air-soil domain, we expect large inter-model differences associated with other climate forcings, such as a model's overall climate sensitivity and degree of arctic amplification. To separate these differences from those associated with the surface air-soil domain, we calculate absolute and relative permafrost vulnerability indices as the ratio of the absolute or fractional extent of permafrost loss to the total high-latitude climate temperature change. Here we define high-latitude climate change as change in MAAT over land, oceans, and ice poleward of 60°N. Using the RCP4.5 scenario, the CMIP5 models have absolute permafrost vulnerability indices of 0.2-3.5 million km^2 permafrost °C^{-1} high-latitude warming, and fractional vulnerability indices that range from 6 to 29 % $°C^{-1}$ (table 2). For the absolute vulnerability index, much of this range is set by outlier models at either end of the sensitivity range: for the models as a set the mean and standard deviations of this value is 1.6 ± 0.7 million km^2 permafrost $°C^{-1}$ high-latitude warming, and for the fractional loss the ensemble mean is 13% ± 6% permafrost loss $°C^{-1}$ high-latitude warming.

Ideally, in this type of multi-model climate change analysis, one would like to find a metric that is both observable within the current climate and has predictive power over the system's response to transient climate change (Hall and Qu 2006). One such relationship, at the scale of individual gridcells, can be seen by the similarity between figs. 3 and 5: the models' predictions of climate control of ALT at the present day inform their predictions of ALT response to changing climate. At the pan-arctic scale, the choice of metric dictates the control variable, because across the set of models, the two vulnerability indices (absolute and fractional) are not correlated ($r^2 < 0.001$). The absolute permafrost vulnerability is largely controlled by the initial permafrost area ($r^2 = 0.41$, $p < 0.01$), while the fractional vulnerability index is negatively correlated with initial permafrost area ($r^2 = 0.48$, $p < 0.01$), and positively correlated with median present-day active layer thickness ($r^2 = 0.27$, $p < 0.05$).

While many processes (such as those listed in table 1) are treated differently, or with varying degrees of complexity, in these models, they share many common characteristics. In addition to what they share in terms of resolved processes, they share a lack of other processes known to be important in permafrost dynamics, including (1) processes that could act to accelerate permafrost loss with warming, such as thermokarst and the lateral thaw associated with fine-scale coupling of thermal and hydrologic properties, and (2) processes that could act to slow permafrost loss with warming, including the presence of massive ground ice which would need to melt in order to substantially deepen active layers.

The lack of representation of these critical processes in any of the CMIP5 models in combination with (1) the wide range of model predictions of ALT and permafrost extent under current and projected climate and (2) relatively poor comparisons with observed permafrost thermal properties (for example, none of the models performed well in every comparison with the data), lead us to conclude that, as a group, the current suite of CMIP5 model projections of permafrost loss and dynamics over the coming century is very uncertain. Given that these dynamics are closely linked to prediction of the potential CO_2 and CH_4 emissions and resulting atmospheric feedbacks, we argue that model projections of high-latitude C cycle climate feedbacks over the next century based on the physics in these models, (even beyond the fact that a permafrost C cycle is not represented in any of these simulations) are also very uncertain.

4. Summary and Conclusions

We compare permafrost thermal dynamics for a set of models participating in the CMIP5 project, to evaluate their behavior under the current climate and assess the range of model predictions for permafrost extent under transient global warming experiments. The models show a wide range of behaviors under the current climate, with many failing to agree with fundamental aspects of the observed soil thermal regime at high latitudes.

Under future climate change, the models differ in their degree of warming, both globally and at high latitudes, and also in the response of permafrost to this warming. All of the models show some loss of permafrost, but a wide range of possible magnitudes in their responses, from 6-29% permafrost loss $^{\circ}C^{-1}$ high-latitude warming. Several of the models predict that substantial permafrost degradation has already occurred (ranging from 3% gain to 49% loss relative to 1850 conditions), though the majority of models at the high end of relative 20th century permafrost loss also show unrealistically small preindustrial permafrost extent; given that such high rates of permafrost loss are not observed, this indicates a too-high sensitivity for those models predicting such losses.

Given the large complexity and number of differing components of the CMIP5 models, we find that a useful approach to understand the model differences is to break down the thermal communication between the surface air and deeper soil by examining changes to the mean and amplitude of the annual temperature cycle across the air to shallow soil and shallow soil to deeper soil interfaces. The available soil temperature observations at high latitudes allow such an observational constraint, and demonstrate that different model representations lead to better or worse agreement with different aspects of the observed soil thermal climate.

Much of the disagreement in modelled mean soil temperatures can be traced to the representation of thermal connection between the air and land surface, and in particular its mediation by snow in winter. There is wide model disagreement on the value of $\Delta\bar{T}_{air-soil}$, the difference in mean temperatures across the air-soil interface, with several of the models predicting the wrong sign for this statistic. Similarly, there is wide model disagreement in the changes of mean and amplitude of soil temperatures with depth; some of which can be tied to differences in modelled soil physical properties and coupling between soil temperature and hydrology. This appears to be particularly the case for the representation of organic layers; even models that do incorporate organic material do so using a mixture of organic and mineral properties, instead of representing organic soils as separate units, with their own dynamics distinct from mineral soils. Models that show deep active layers under the current climate are more likely to have larger fractional reductions in their fractional permafrost extent with warming.

Given that the high latitude soil C pool is the single largest component of the terrestrial carbon cycle that could respond directly to climate change on timescales of centuries, it is important for ESMs to accurately predict how the permafrost soil climate may respond to warming. With this analysis, we show that widespread disagreement exists amongst this generation of ESMs. All CMIP5 models predict some loss of permafrost, and increasing loss under higher warming scenarios, but the magnitude of this loss is still highly uncertain.

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Model Name	Modelling Center	Land Model	Mult. Snow Layers	Snow betw. soil and atmosphere	Latent Heat of soil water	Differing Froz./ Unfroz. Soil Therm. Cond.	Org. Matt.	Reference
BCC-CSM1-1	BCC	BCC_AVIM1.0	Yes	Yes	Yes	Yes	No	Ji (1995)
CCSM4	NCAR	CLM4	Yes	Yes	Yes	Yes	Yes	Lawrence et al. (2011)
CESM1-CAM5	NCAR	CLM4	Yes	Yes	Yes	Yes	Yes	Lawrence et al. (2011)
CanESM2	CCCMA	CLASS	No	Yes	Yes	Yes	No	Verseghy (1991)
GFDL-ESM2G	GFDL	GFDL LM3.0	Yes	Yes	Yes	Yes	No	Dunnea et al. (2012)
GFDL-ESM2M	GFDL	GFDL LM3.0	Yes	Yes	Yes	Yes	No	Dunnea et al. (2012)
GISS-E2-R	GISS	GISS LS	Yes	Yes	Yes	Yes	Yes	Rosenzweig and Abramopoulos (1997)
HadCM3	MOHC	MOSES	No	No	Yes	Yes	No	Cox et al. (1999)
HadGEM2-CC	MOHC	MOSES2	No	No	Yes	Yes	No	Essery et al. (2003)
HadGEM2-ES	MOHC	MOSES2	No	No	Yes	Yes	No	Essery et al. (2003)
INM-CM4	INM	INM-CM4	Yes	Yes	Yes	Yes	Yes	Volodin et al. (2010)
IPSL-CM5A-LR	IPSL	ORCHIDEE	Yes	No	No	No	No	Krinner et al. (2005)
IPSL-CM5A-MR	IPSL	ORCHIDEE	Yes	No	No	No	No	Krinner et al. (2005)
MIROC-ESM-CHEM	JAMSTEC	MATSIRO	Yes	Yes	Yes	No	No	Takata et al. (2003)
MIROC-ESM	JAMSTEC	MATSIRO	Yes	Yes	Yes	No	No	Takata et al. (2003)
MIROC5	JAMSTEC	MATSIRO	Yes	Yes	Yes	No	No	Takata et al. (2003)
MPI-ESM-LR	MPI-M	JSBACH	Yes	No	No	No	Yes	Raddatz et al. (2007)
MRI-CGCM3	MRI	HAL	Yes	Yes	Yes	Yes	No	Yukimoto et al. (2012)
NorESM1-M	NCC	CLM3	Yes	Yes	Yes	Yes	Yes	Lawrence et al. (2011)

TABLE 1. List of models used in this analysis, the modelling groups that developed them, model attributes and references. The model attributes listed here are relevant to soil physics at high latitudes, including: Whether the model incldues a multi-layer snow model, whether the snow acts to insulate between the soil and atmosphere, the inclusion of soil water latent heat and differing frozen and unfrozen soil thermal conductivity, and whether soil physical properties are effected by geographically-varying soil organic matter.

		Historical		RCP 4.5									RCP 2.6		RCP 8.5	
model name	PF area 2005 (*10^6 km^2)	PF area 1850 (*10^6 km^2)	Fraction PF lost 1850 - 2005 (unitless)	PF area 2050 rcp4.5 (*10^6 km^2)	Fraction PF remaining 2005-2050 rcp4.5 (unitless)	PF area 2100 rcp4.5 (*10^6 km^2)	Fraction PF remaining 2005-2100 rcp4.5 (unitless)	Global warming 2005-2100 RCP4.5 (°C)	High-lat warming 2005-2100 RCP4.5 (°C)	Arctic Amplification (unitless)	total loss PF / degree high-lat warming (*10^6 km^2 / °C)	fractional loss PF / degree high-lat warming RCP4.5 (%/°C)	PF area 2100 rcp2.6 (*10^6 km^2)	Fraction PF remaining 2005-2100 rcp2.6 (unitless)	PF area 2100 rcp8.5 (*10^6 km^2)	Fraction PF remaining 2005-2100 rcp8.5 (unitless)
BCC-CSM1-1	1.4	2.7	0.47	0.9	0.66	0.8	0.58	1.2	2.7	2.3	0.2	15.4	–	–	0.5	0.36
CCSM4	**10.5**	13.1	0.20	8.0	0.76	5.7	0.54	1.2	2.1	1.7	2.3	21.8	7.7	0.73	2.7	0.26
CESM1-CAM5	**12.3**	13.8	0.11	9.1	0.74	6.1	0.49	2.0	4.2	2.1	1.5	12.1	7.9	0.64	2.4	0.19
CanESM2	3.2	5.7	0.43	0.9	0.29	0.5	0.14	1.9	4.1	2.2	0.7	20.8	1.1	0.34	0.0	0.01
GFDL-ESM2G	25.6	27.3	0.06	23.5	0.92	22.0	0.86	0.7	1.5	2.2	2.3	9.1	25.1	0.98	15.3	0.60
GFDL-ESM2M	24.3	27.0	0.10	21.1	0.87	20.0	0.82	0.8	1.2	1.5	3.5	14.3	22.1	0.91	14.8	0.61
GISS-E2-R	4.5	8.2	0.45	2.6	0.58	2.5	0.57	0.9	1.4	1.5	1.4	31.0	4.4	0.98	1.0	0.23
HadCM3	27.3	28.6	0.05	–	–	–	–	–	–	–	–	–	–	–	–	–
HadGEM2-CC	25.1	24.6	-0.02	20.6	0.82	17.1	0.68	2.0	5.2	2.6	1.5	6.2	–	–	9.4	0.38
HadGEM2-ES	22.3	23.4	0.05	17.8	0.80	14.5	0.65	2.1	5.3	2.6	1.5	6.5	17.5	0.78	7.0	0.31
INM-CM4	**14.0**	15.6	0.11	12.8	0.92	11.5	0.82	1.1	2.3	2.2	1.1	7.7	–	–	9.8	0.70
IPSL-CM5A-LR	**12.6**	18.5	0.32	8.5	0.68	7.0	0.56	1.8	3.4	1.9	1.7	13.2	9.8	0.78	2.0	0.16
IPSL-CM5A-MR	**10.1**	14.9	0.32	6.5	0.64	4.6	0.46	1.8	3.5	1.9	1.6	15.7	8.1	0.80	1.4	0.14
MIROC-ESM	**11.5**	12.6	0.08	7.3	0.63	4.7	0.40	2.1	5.0	2.4	1.4	11.8	6.4	0.55	0.6	0.05
MIROC5	**17.4**	19.0	0.08	14.5	0.83	12.3	0.70	1.3	3.3	2.5	1.5	8.9	14.2	0.81	6.5	0.37
MPI-ESM-LR	8.6	13.7	0.37	6.2	0.72	4.0	0.46	1.3	2.6	2.0	1.7	20.3	6.9	0.80	0.5	0.06
MRI-CGCM3	**15.5**	16.7	0.08	13.2	0.86	10.9	0.70	1.5	3.0	2.0	1.5	9.8	12.7	0.82	7.7	0.50
NorESM1-M	**13.3**	15.1	0.11	10.2	0.77	7.3	0.55	1.5	4.1	2.8	1.5	11.1	9.7	0.73	3.5	0.27
Observations	15.0	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–

TABLE 2. Modelled current and future permafrost extent in the upper 3 m of soil. For qualitative comparison, the 50% of models with the least bias for the present time period are noted in boldface type.

Model Name	Max Depth (m)	Median active layer thickness (m)	Regression constant of trivariate regression (m)	Slope of ALT-MAAT (m/°C)	slope of ALT-ampl (m/°C)	r^2 of trivariate regression (unitless)	Mean T offset 0-1m (°C)	Mean atten. 0m-1m (unitless)	Mean T offset air-0m (°C)	Mean atten. air-0m (unitless)	RMS err: T offset 0-1m (°C)	RMS err: atten. 0-1m (unitless)	RMS err: T offset air-0m (°C)	RMS err: atten air-0m (unitless)
BCC-CSM1-1	3.4	1.75	**0.16**	0.253	0.48	**0.41**	**-0.33**	**0.35**	**8.78**	0.29	**0.88**	0.28	**4.28**	**0.33**
CCSM4	43.7	1.20	1.23	**0.101**	**0.09**	**0.46**	-1.54	0.25	**7.20**	**0.53**	1.62	0.25	**3.13**	**0.23**
CESM1-CAM5	43.7	1.26	1.46	**0.074**	**0.05**	**0.31**	-1.60	0.30	**8.52**	**0.52**	1.65	**0.21**	**4.51**	**0.22**
CanESM2	4.1	1.38	3.23	0.143	-0.05	**0.39**	**-0.15**	0.71	0.82	**0.84**	**0.84**	0.34	7.14	0.36
GFDL-ESM2G	10.0	0.86	**1.07**	**0.083**	**0.05**	0.55	-1.45	0.57	1.15	0.88	1.34	0.24	5.99	0.35
GFDL-ESM2M	10.0	0.87	**1.20**	**0.097**	**0.04**	0.47	-1.66	**0.48**	0.95	0.86	1.46	**0.16**	6.15	**0.32**
GISS-E2-R	3.5	1.75	**0.31**	**0.042**	0.18	**0.15**	**-0.21**	0.66	**2.81**	**0.55**	**0.80**	0.30	**4.21**	**0.18**
HadCM3	3.0	0.63	**1.17**	**0.053**	0.00	**0.34**	2.06	**0.41**	0.70	0.93	2.99	**0.17**	6.26	0.42
HadGEM2-CC	3.0	0.64	**0.80**	**0.038**	0.02	**0.31**	1.72	**0.46**	0.80	1.01	2.66	**0.15**	6.29	0.47
HadGEM2-ES	3.0	0.64	**1.11**	**0.048**	0.00	**0.34**	1.83	**0.44**	0.89	0.99	2.65	**0.15**	6.23	0.44
INM-CM4	15.0	0.00	**0.36**	0.144	**0.13**	**0.29**	**-0.99**	0.29	**4.04**	**0.42**	**1.00**	0.25	**3.72**	**0.24**
IPSL-CM5A-LR	5.5	2.67	1.57	0.171	0.22	0.77	2.07	0.53	-0.14	1.05	2.91	**0.23**	7.17	0.50
IPSL-CM5A-MR	5.5	2.69	1.70	0.183	0.23	0.80	1.83	0.54	-0.23	1.05	2.56	0.24	7.19	0.50
MIROC-ESM	14.0	1.59	1.99	0.160	**0.08**	0.88	**-0.07**	**0.44**	**3.94**	**0.59**	**0.91**	0.24	**5.29**	**0.27**
MIROC5	14.0	1.00	**1.03**	**0.089**	**0.07**	0.75	**-0.07**	**0.47**	**1.47**	**0.85**	**0.92**	**0.13**	**5.80**	0.35
MPI-ESM-LR	9.6	3.22	1.77	0.172	0.19	0.70	1.18	0.62	-0.01	1.03	1.88	0.28	6.92	0.49
MRI-CGCM3	10.0	1.00	**1.11**	0.108	**0.10**	0.75	**-0.55**	0.33	**6.29**	**0.56**	**0.83**	**0.22**	**3.43**	**0.19**
NorESM1-M	42.1	0.85	1.32	**0.089**	**0.07**	**0.37**	**-1.02**	**0.40**	**7.51**	**0.50**	**1.21**	**0.15**	**3.12**	**0.22**
Observations	—	—	0.32	0.03	0.09	0.13	-0.53	0.40	6.17	0.57	–	–	–	–

TABLE 3. Summary statistics of model-data comparisons. For each column where an observational constrain was available, we assigned a qualitative threshold of the 50% of models that were closest to observations, and note these with a boldface type.

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11 Multi-model predictions of permafrost area under climate change scenarios: (a,d) RCP2.6, (b,e) RCP4.5, (c,f) RCP8.5. (a-c) Total permafrost extent for each model in CMIP5 analysis. A small gap is shown between historical experiment and RCP scenarios. (d-f) Box-whisker diagrams of fraction of permafrost at start of RCP scenarios remaining during 21st century.

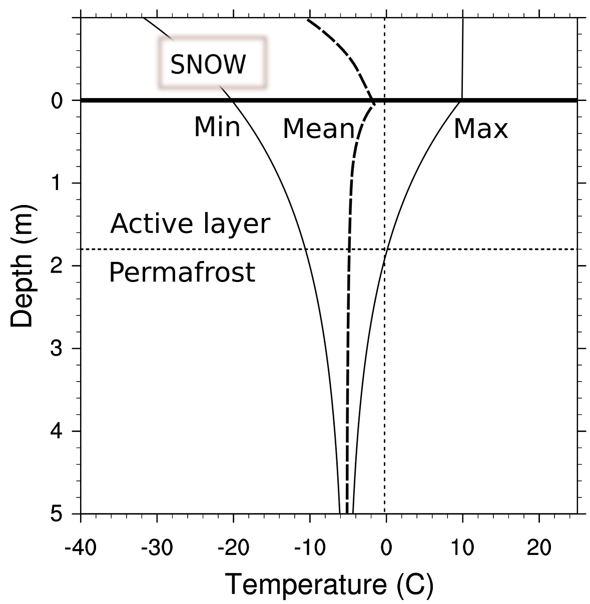


FIG. 1. Simplified schematic of permafrost thermal dyamics: Solid lines show vertical profiles of minimum and maximum annual temperatures across air-soil interface; dashed line shows mean annual temperature profile. Snow acts to insulate soils during winter, leading to steep gradient in wintertime and mean temperatures.

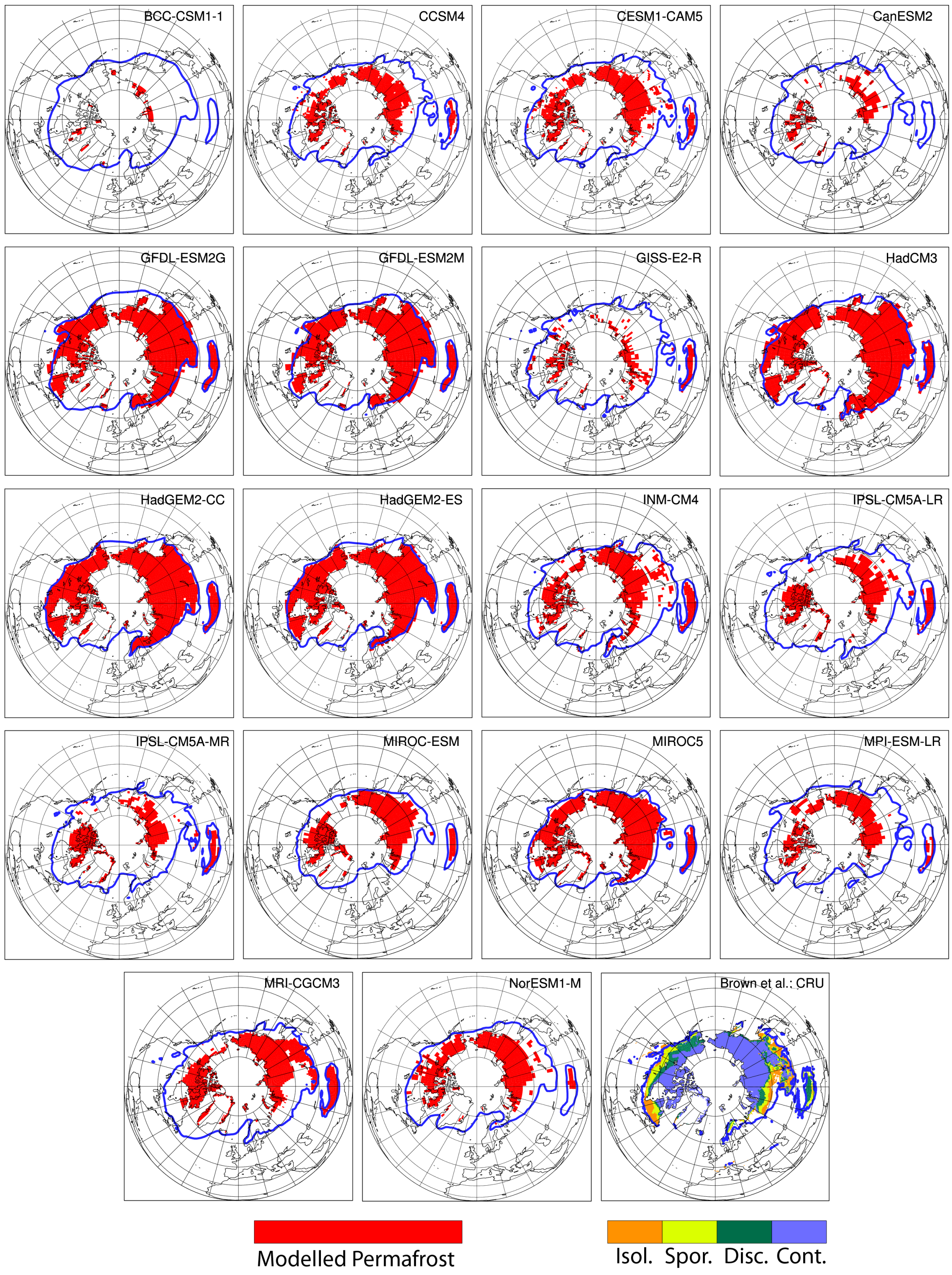


FIG. 2. Extent of permafrost for each of the CMIP5 models under the current climate (using years 2005-2015 from the RCP4.5 scenario.), as well as for the map of Brown et al. (1998). Also shown in thick blue line is the zero-degree isotherm in MAAT for each of the models over the same time period, and from the CRU dataset in the observations panel.

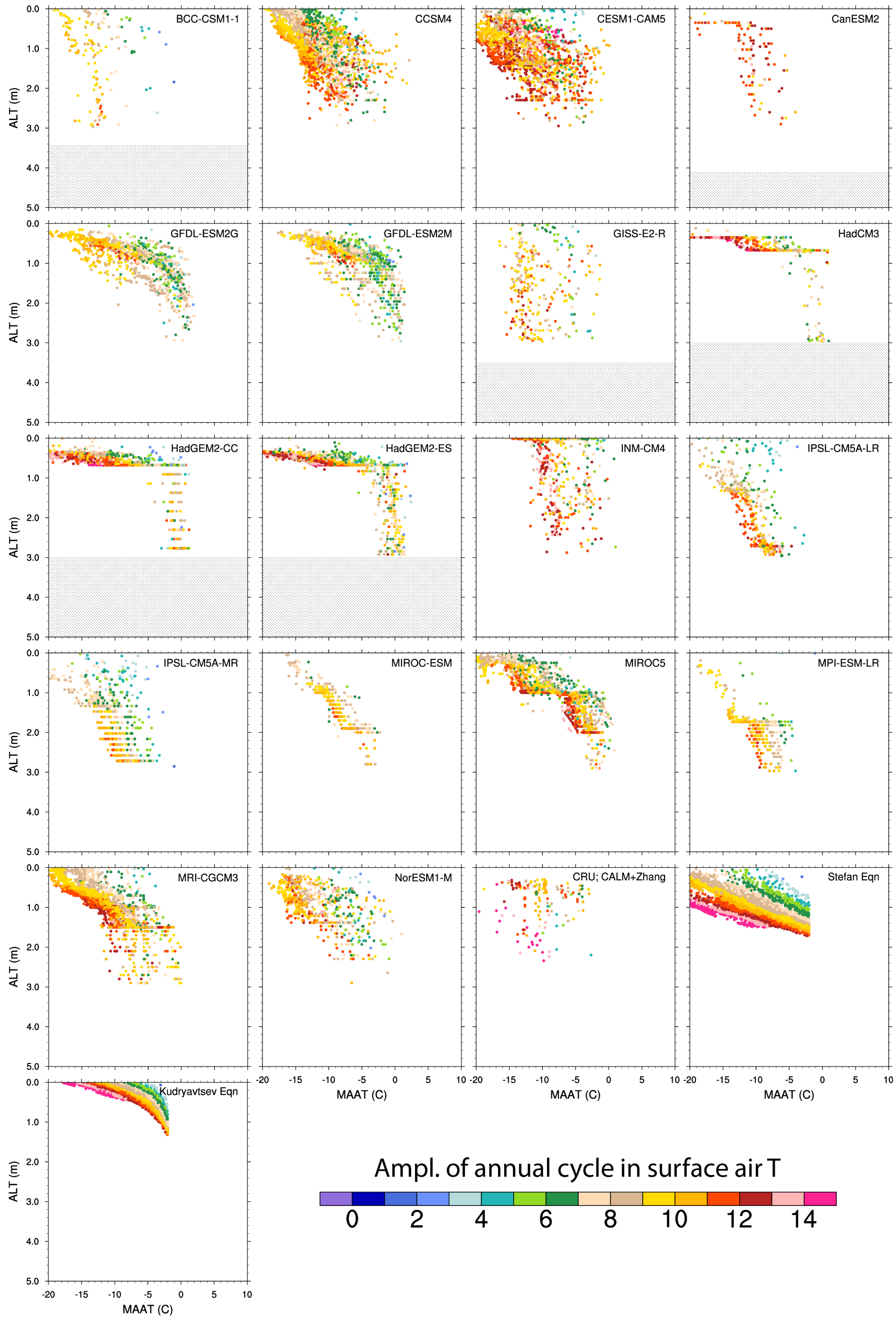


FIG. 3. ALT plotted against MAAT for each of the CMIP5 models and observations. Dots are colored based on the amplitude (1/2 total seasonal range) of the annual temperature cycle. The grey area at the base of some panels indicates depth beyond that model's vertical domain. Also shown are high latitude (lat > 55°N) data from the CALM and Zhang et al. (2006) datasets, with atmospheric climatology defined by the corresponding gridcells of the CRU dataset. Last 2 panels show two analytic solutions to the heat equation with cyclic upper boundary and phase change (Stefan and Kudryavstev equations), given a single set of soil thermal properties and CRU climatologies.

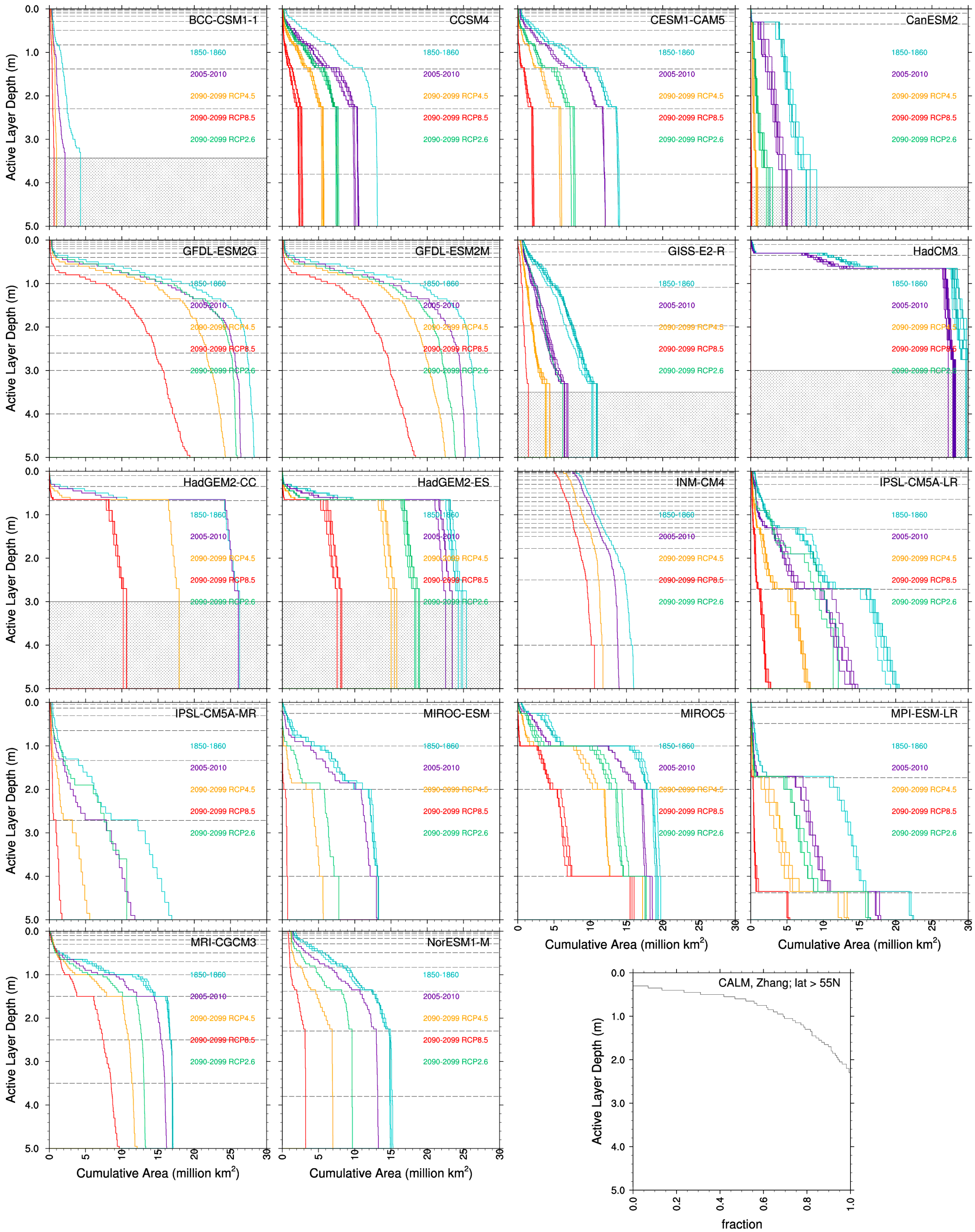


FIG. 4. Cumulative distribution of ALT over the entire modelled permafrost domain in each of the models for historical, current, and future time periods. Also included are observations of current ALT for sites in the CALM and Zhang et al. (2006) datasets north of 55°N, though a direct comparison cannot be made from the observations to the models since the observations have different spatial coverage than the models. Each curve corresponds to the total area (on the horizontal axis) with an active layer less than or equal to the depth (on the vertical axis). Multiple curves for a given time period correspond to individual ensemble members. Model level interfaces are shown as dashed horizontal lines. The grey area at the base of some panels indicates depth beyond that model's vertical domain.

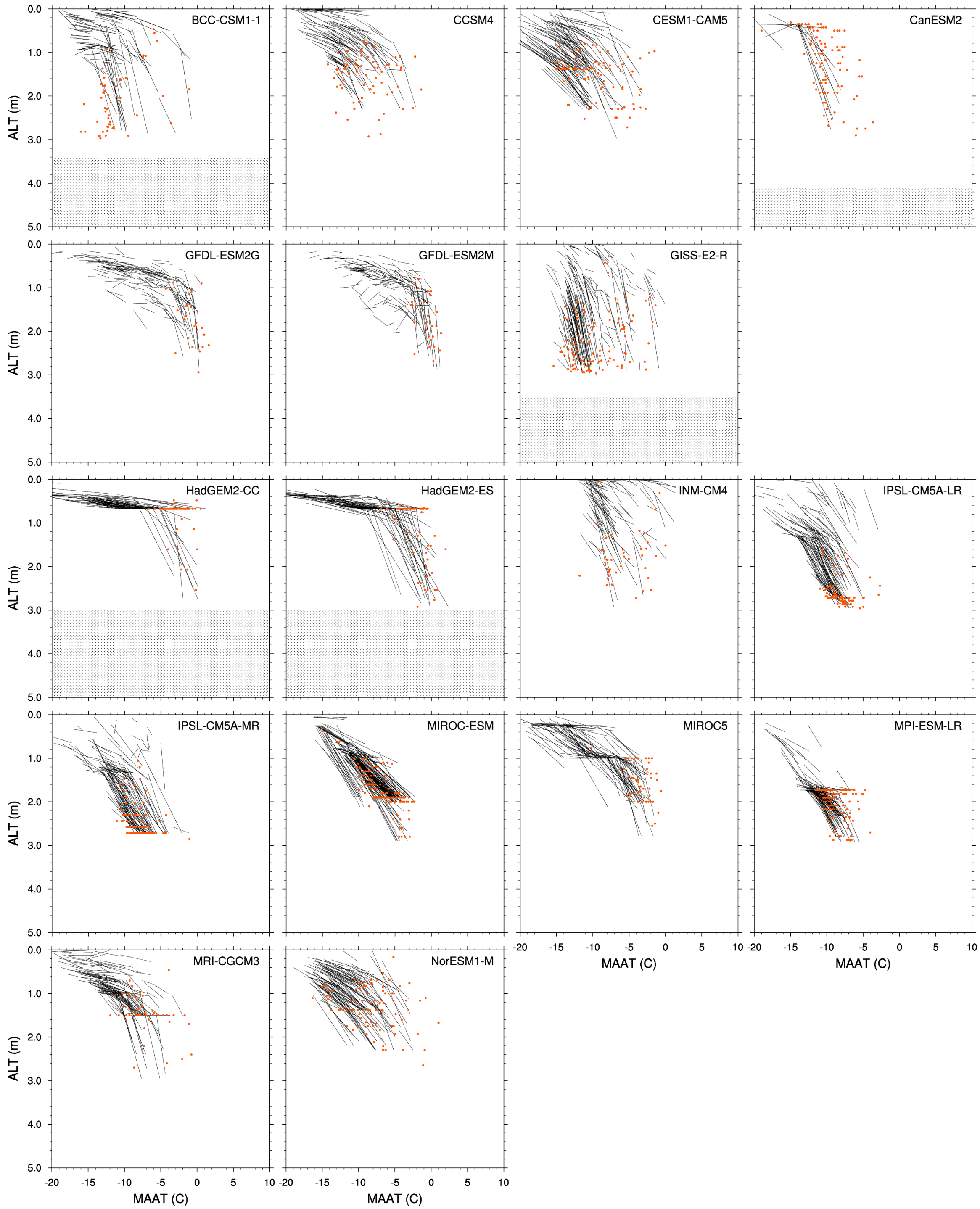


FIG. 5. Time trajectories of modelled active layer change in response to changing climate: the ends of each line correspond to the current (2005-2010) and future (2090-2099) ALT and MAAT for a collection of initially permafrost gridcells within each model, for the scenario RCP4.5. Red dots correspond to initially-permafrost gridcells that did had lost near-surface permafrost by the end of the 21st century. The grey area at the base of some panels indicates depth beyond that model's vertical domain. HadCM3 simulations extend only through the mid-21st century and are thus omitted here.

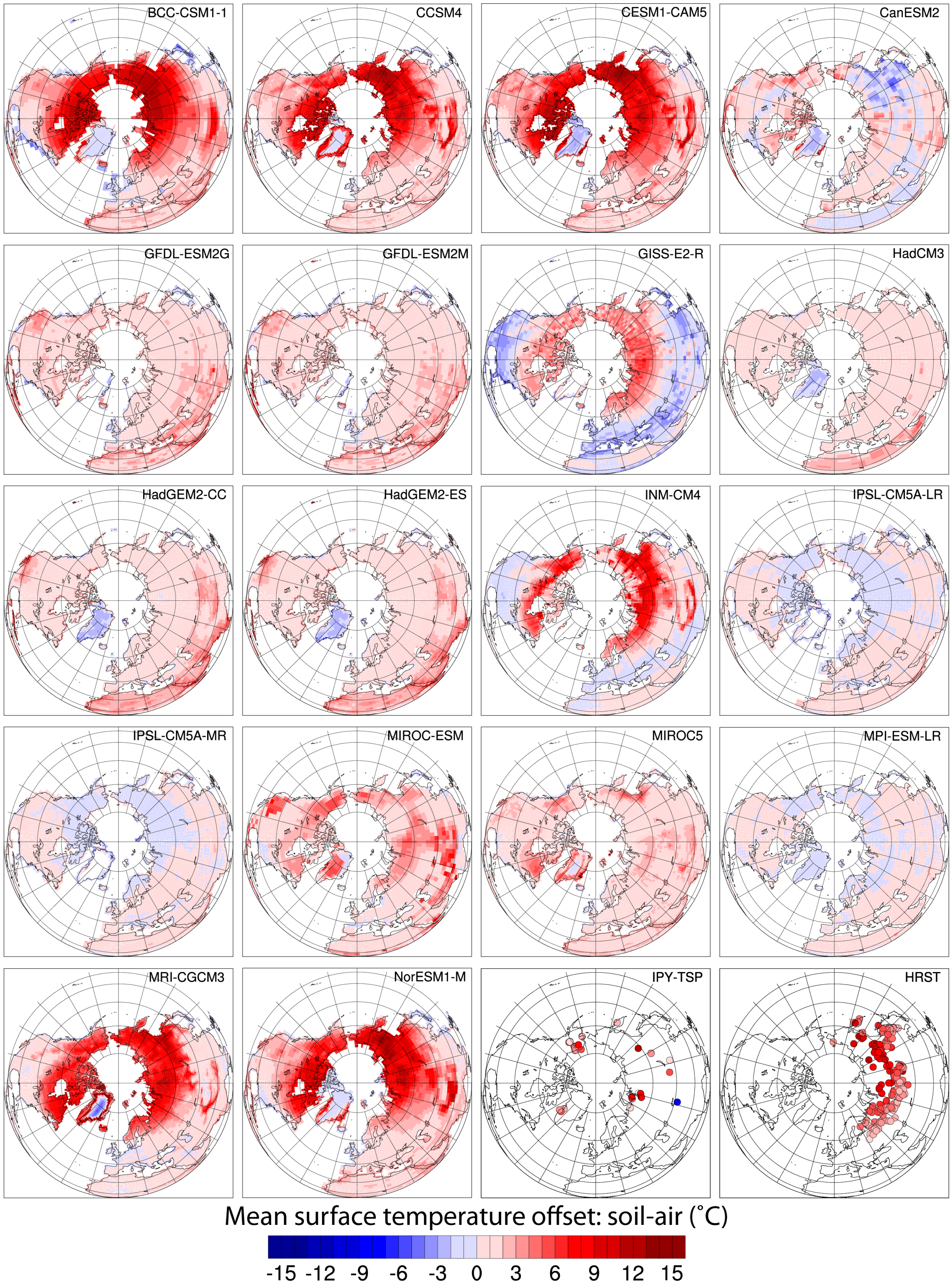


FIG. 6. Maps of $\Delta\bar{T}_{air-soil}$, the difference in mean annual temperatures between uppermost soil and air. Warm colors indicate soil is warmer than air temperature. A major source of the high-latitude temperature offset is from snow insulation effects.

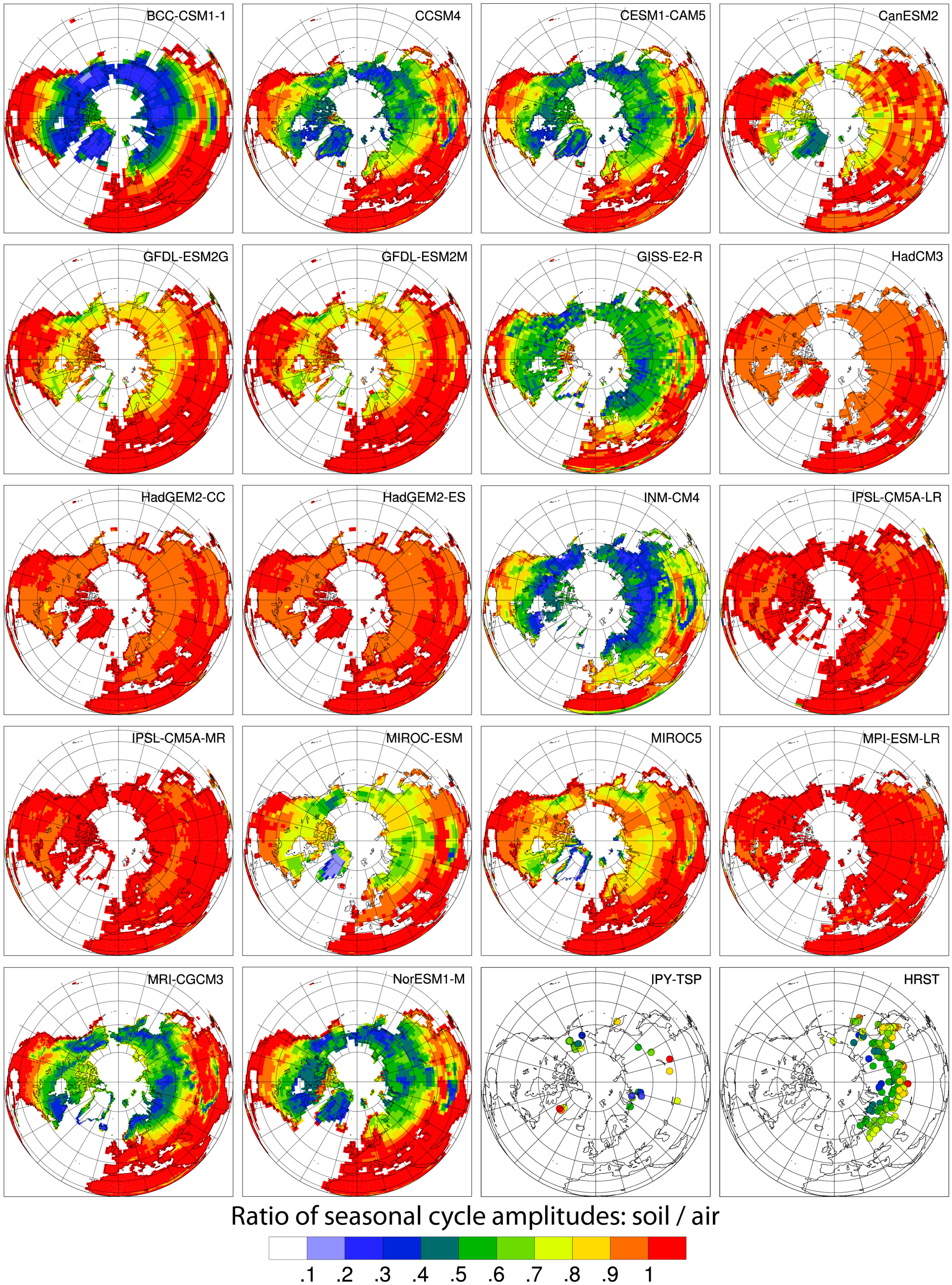
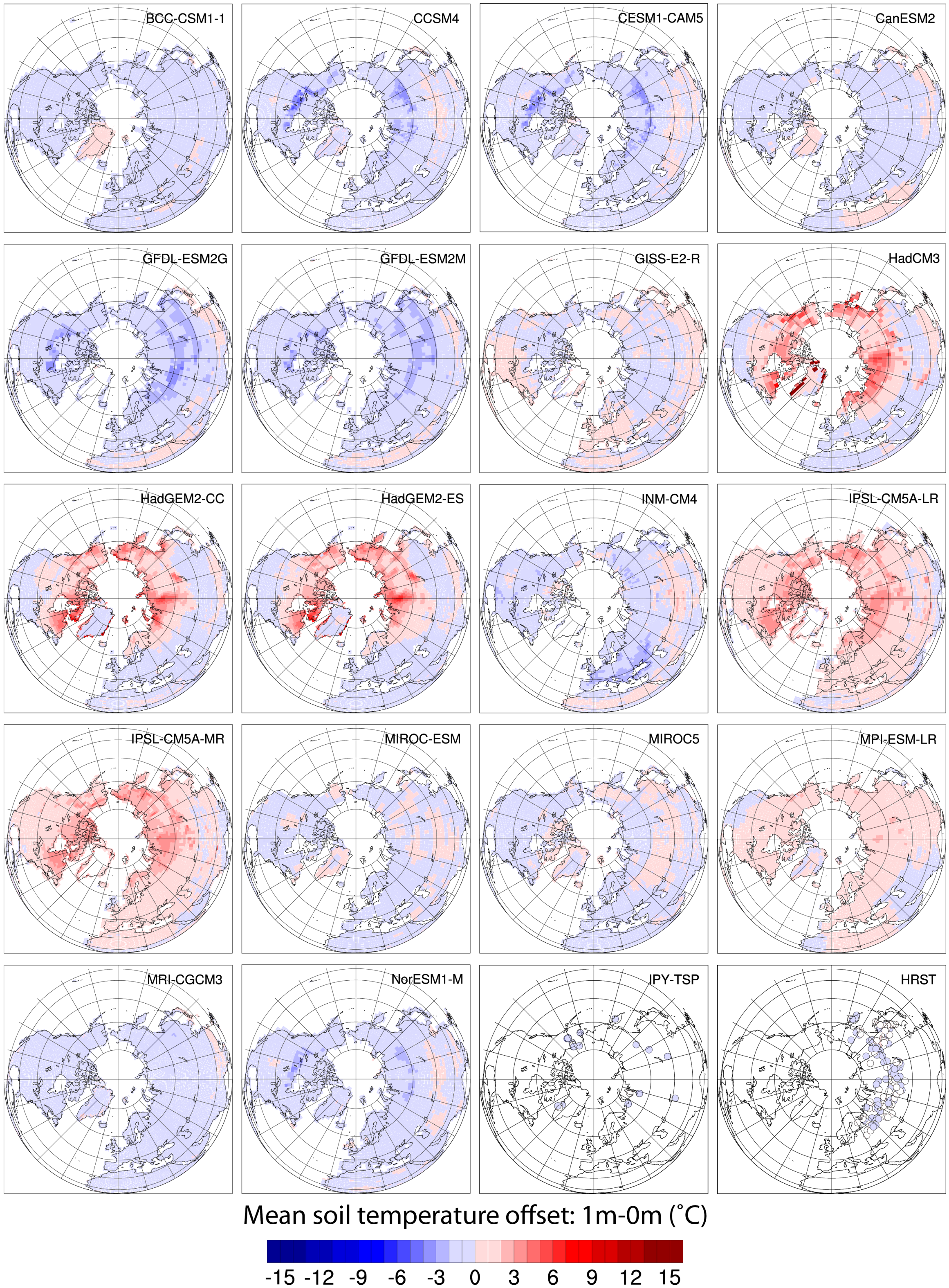


FIG. 7. Maps of $\alpha_{air-soil}$, the ratio in amplitude of seasonal temperature cycle between air and surface soil. As with temperature offset, a major source of high-latitude seasonal cycle attenuation is from snow insulation effects, and spatial patterns follow those in fig. 6 closely.



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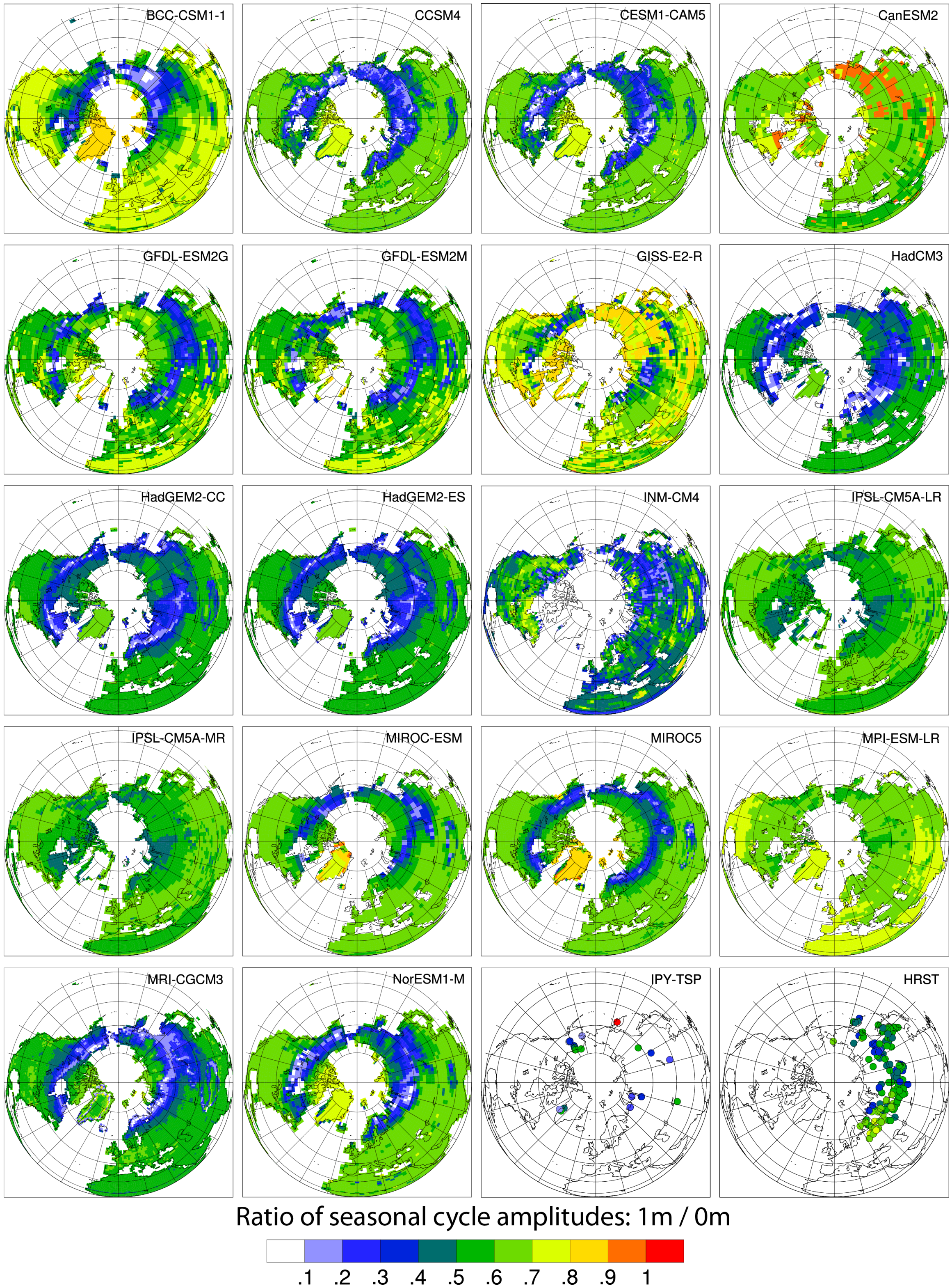

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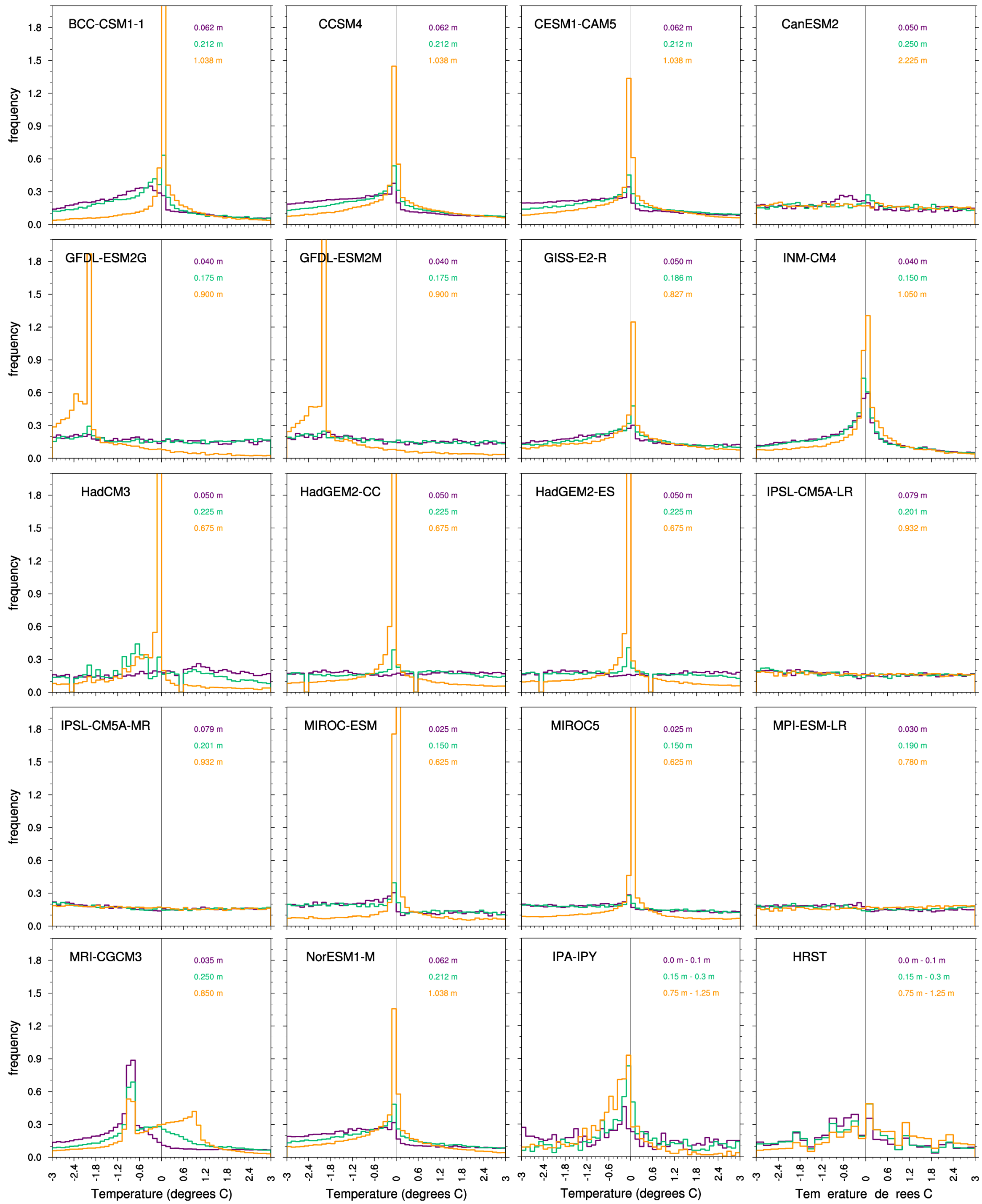

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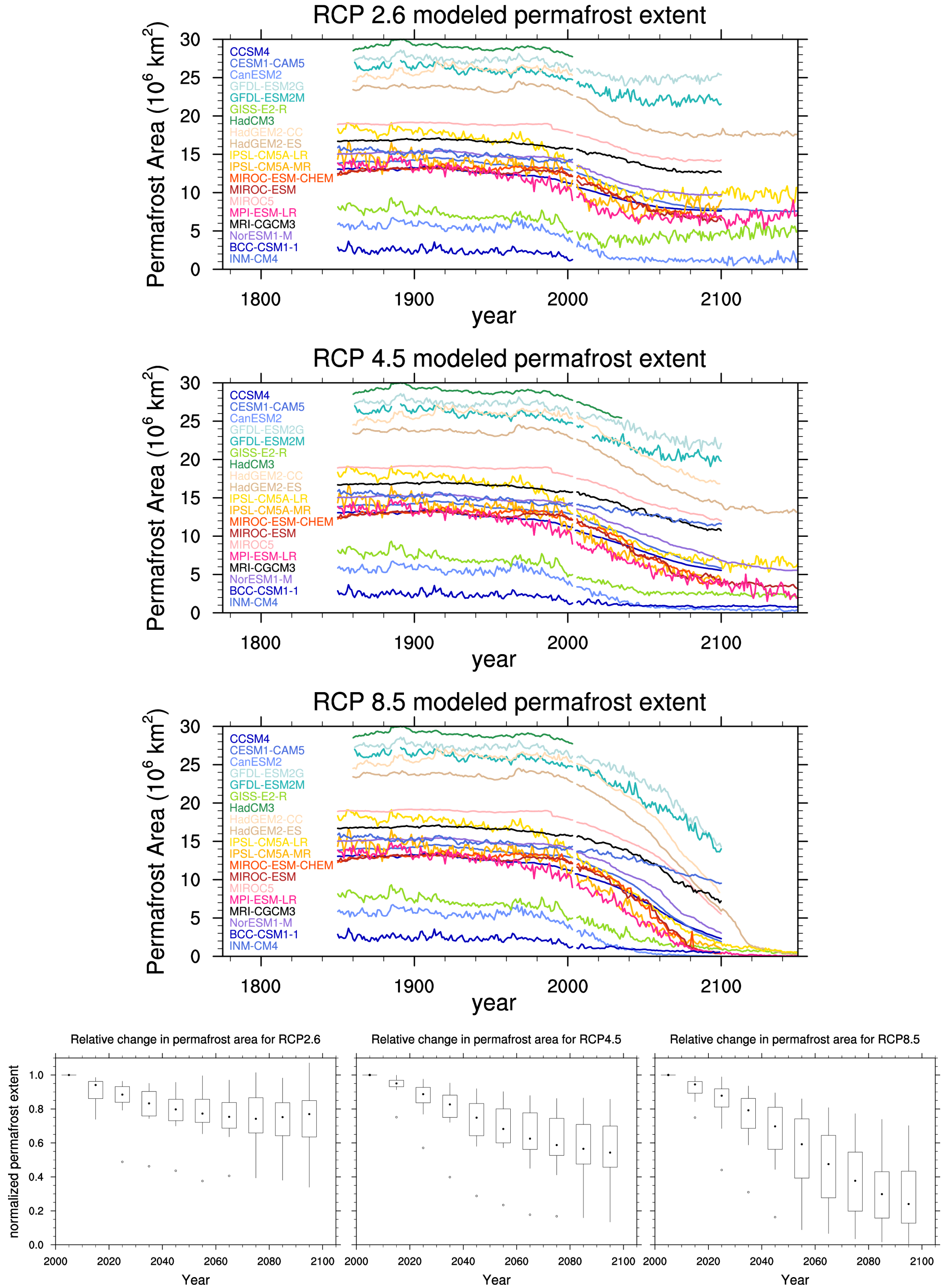

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