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Projecting future forest microclimate using a land surface model

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E-mail: gabriel.hes@univ-tlse3.fr and edouard.davin@unibe.ch**Keywords:** Community Land Model, forest modeling, temperature buffering, tropical biodiversity, climate change, Land Surface Model, soil temperature**Abstract**

The forest understory experiences temperature variations that are dampened compared to adjacent open areas, allowing the development of a forest microclimate and associated ecological conditions. It is however unclear to what extent forests will maintain this buffering effect under increasing global warming. Providing reliable projections of future forest microclimates is therefore crucial to anticipate climate change impacts on forest biodiversity, and to identify corresponding conservation strategies. Recent empirical studies suggest that the buffering of air temperature extremes in forest understory compared to open land could increase with global warming, albeit at a slower rate than macroclimate temperatures. Here, we investigate the trend of this temperature buffering effect in a high-emission global warming scenario, using the process-based Land Surface Model CLM5.1. We find biome-dependant buffering trends with strongest values in tropical forests where buffering increases for every degree of global warming by 0.1 °C for maximum soil temperature, and by 0.2 °C for maximum canopy air temperature. In boreal regions, forest microclimate exhibits a strong seasonality and the effect of global warming is more uncertain. Thus, our results highlight the importance of tropical forest canopies in particular, in maintaining hospitable conditions for understory species while increasing their climate debt under global warming. Our research also illustrates the potential and limitations of Land Surface Models to simulate forest microclimate, and calls for further collaborations between Earth system modelers and ecologists to jointly question climate and biosphere dynamics.

1. Introduction

Forests play an important role in shaping climate conditions from local to global scales by mediating greenhouse gas, water and energy fluxes between the land and the atmosphere (Davin and Noblet-Ducoudré 2010, Teuling *et al* 2017, O'Connor *et al* 2021, Roebroek *et al* 2023). In addition, tree canopies create distinct climatic conditions in the forest understory, which are typically referred to as the forest microclimate (Geiger *et al* 2009). Tree canopies act as thermal insulators, which buffer understory temperature variations (von Arx *et al* 2012, De Frenne *et al*

2019, Meier *et al* 2019). Indeed understory temperatures (T_{in}) are usually cooler than outside-forest temperatures (T_{out}) under hot conditions and warmer under cold conditions, a phenomenon known as the 'buffering effect', and thus typically reduce the diurnal or seasonal amplitude of temperature variations (Geiger *et al* 2009, Von Arx *et al* 2013, Li *et al* 2015, De Frenne *et al* 2019). However, the opposite phenomenon -an amplification effect- is also observed in specific forest environments such as open canopy mountain forests (Vandewiele *et al* 2023). A common metric to assess both effects is the difference between within-forest temperature (forest

microclimate) and outside-forest temperature (macroclimate) defined as $\Delta T = T_{in} - T_{out}$ (De Frenne *et al* 2019). This temperature offset has been shown to correlate negatively with the mean macroclimate temperature across biomes and to increase in magnitude with decreasing latitudes (De Frenne *et al* 2019).

Temperature extremes strongly affect biodiversity by altering species' physiological performances (Huey *et al* 2009, Vasseur *et al* 2014). In that respect, forest microclimates act as climatic refugia (i.e. hospitable habitats) for forest understory species (Keppel *et al* 2012, Hannah *et al* 2014). Over the past decades it has been shown that such species respond more closely to microclimate variations than macroclimate warming (De Frenne *et al* 2013, Williamson *et al* 2020, Zellweger *et al* 2020, De Pauw *et al* 2022). To date, we observe that the discrepancy between forest microclimate and regional climate often delays species migration or local extinction compared to what we would expect from macroclimate warming (Lenoir *et al* 2017), a process often coined as 'climatic debt' (Bertrand *et al* 2016, Richard *et al* 2021). However, as stronger and more frequent extreme temperatures are expected with climate change (IPCC 2021), we do not know if forest understories can serve as long-term refugia for species, or, in other words, how large of a climatic debt can species accumulate before being locally extinct (De Frenne *et al* 2021, Richard *et al* 2021). To address this question, one must first understand how stable the temperature decoupling in forests (inside vs. outside) is under global climate warming (Lenoir *et al* 2017, De Frenne *et al* 2019, 2021, Lembrechts and Nijs 2020). In particular, determining the evolution of tropical forest microclimates is of utmost importance given that at least $\frac{2}{3}$ of global species lie within tropical forest ecosystems (Pillay *et al* 2022). In that respect, better understanding biotic and abiotic drivers of forest microclimate is key to anticipating the response of forest biodiversity to global warming (Frey *et al* 2016, De Frenne *et al* 2021).

The growing number of monitored forest sites allows assessment of the relations between species distribution, forest structure and microclimate, but such efforts are strongly limited by the scarcity of microclimate data (De Frenne *et al* 2021). Analysis of the available data shows that forest temperature offsets correlate negatively with outside forest mean temperature (i.e. forest buffering generally decreases with latitude) (De Frenne *et al* 2019). In particular, we observe that tropical forests are strongly buffered even for small macroclimate temperature increases (Senior *et al* 2018). However, we currently lack long time-series required to provide robust evidence of the impact of a substantial rise in global mean temperatures on forest microclimate (Lembrechts and Nijs 2020, Sanczuk *et al* 2023). It is yet unclear whether

the magnitude of the temperature offset will remain stable, increase, or decrease over time as macroclimate warms (De Frenne *et al* 2021). De Lombaerde *et al* (2022) addressed this question through a statistical modeling approach. Macroclimate was found to be a major driver of the observed temperature offset across multiple forest sites. The associated statistical relationship was used to generate future projections of temperature offsets globally. Assuming a constant forest cover, the warming rate of the forest understory was forecast to be slower than that over non-forested areas in a global warming scenario. Such a prediction is consistent with the observation of a higher buffering in warmer biomes. The question we address in this study is whether this space-for-time statistical approach is appropriate to forecast warming rates of understory microclimate across different biomes. For instance, in their study, De Lombaerde *et al* (2022) assume that the statistical relationship between macroclimate and microclimate is constant, irrespective of the warming conditions.

Because large-scale experimental warming of forest is neither feasible nor desirable, estimation of how forest microclimate will be altered in a changing climate requires the use of models encapsulating our mechanistic understanding of the physical controls over land-atmosphere interactions, energy budgets and gas exchange. Here we use a process-based global land surface model (CLM5.1) to produce estimates of the microclimatic buffering effect. CLM5.1 represents the forest understory microclimate via the differentiation between the forest soil temperature and the grassland soil temperature. Similarly, the canopy microclimate buffering is represented here as the differentiation between the forest canopy air temperature and the grassland canopy air temperature. Although several multilayer canopy models (CLMml (Bonan *et al* 2018), ORCHIDE-CAN (Ryder *et al* 2016), MuSICA (Ogée *et al* 2003), microclimC (Maclean and Klingses 2021)) represent forest energy and hydrology budgets more accurately (Bonan *et al* 2021), none is currently operational in a global modeling system and they have primarily been used in single point evaluations. CLM5.1 can be run globally with moderate resolution (0.5°) and in addition, was recently modified to include a biomass heat storage scheme, generating more realistic diurnal temperature variations in vegetated areas (Meier *et al* 2019). Land model simulations provide a globally consistent representation of forests while observational studies are confronted to data scarcity for specific biomes. For instance, the SoilTemp database (Lembrechts *et al* 2020) currently lacks data over the boreal region.

In this article, we investigate the global change of forest-induced soil and canopy air temperature over historical and future periods. Our main hypothesis is that there is an increasing buffering effect under

forest canopies with global warming, as suggested by De Lombaerde *et al* (2022). To test this hypothesis in the presence of large changes in temperature, we use the highest IPCC warming scenario (SSP5-8.5) and focus on minimum and maximum temperatures, assuming that a trend in the offset signal would be more visible in this scenario. We first analyse global historical and future patterns of simulated temperature offsets. We then provide estimations of future offsets per degree of global warming. Finally, we discuss the implications of future temperature buffering on forest biodiversity and the limits of using a single layer Land Surface Model such as CLM5.1 to represent within-forest temperature variations.

2. Methods

2.1. Model description

To investigate the effect of climate change on global forest temperatures, we used the land component CLM (Community Land Model) of the Community Earth System Model (CESM2, version 2.1.2, Danabasoglu *et al* (2020)), a state-of-the-art and open-source Earth System Model. CLM version 5.1 (CLM5.1) is an advanced land surface model, representing biophysical and biochemical processes related to the Earth surface (Lawrence *et al* 2019). The model accounts for surface heterogeneity through a nested-subgrid hierarchy. Each gridcell is divided into land units, representing different land uses, which are composed of one or multiple soil columns (Lawrence *et al* 2019). For vegetated land units, the soil column is subdivided into a maximum of 15 different plant functional type (PFT) patches and bare ground, which all share the same column state and compete for soil water. The state variables and fluxes for water and energy are defined at the column level as weighted averages over all PFTs within the column. Here we used a reduced complexity version of the model (satellite phenology, SP mode) that disables biogeochemical processes and instead is driven by a constant seasonal cycle for leaf and stem area indices (LAI and SAI) and by prescribed canopy height for each PFT patch. The SP mode allows a focus on biophysical processes that are not subject to biases introduced via uncertain simulation of plant growth and competition.

CLM5.1 features a homogeneous single layer representation of the plant canopy (also known as the 'big leaf approach') with no internal vertical structure (Bonan *et al* 2021). This simplified representation of the forest canopy is complemented by recent developments of the biomass heat storage scheme by Meier *et al* (2019) and Swenson *et al* (2019), resulting in more dampened diurnal land surface temperature profiles in forests that better represents observed diurnal temperature profiles. Based on PFT-specific

thermal properties, stems and leaves store and radiate heat, which generates a lag in air temperature variations. Finally, we used the novel surface roughness parametrization described in Meier *et al* (2022) for improved temperature and wind speed profiles above forested regions.

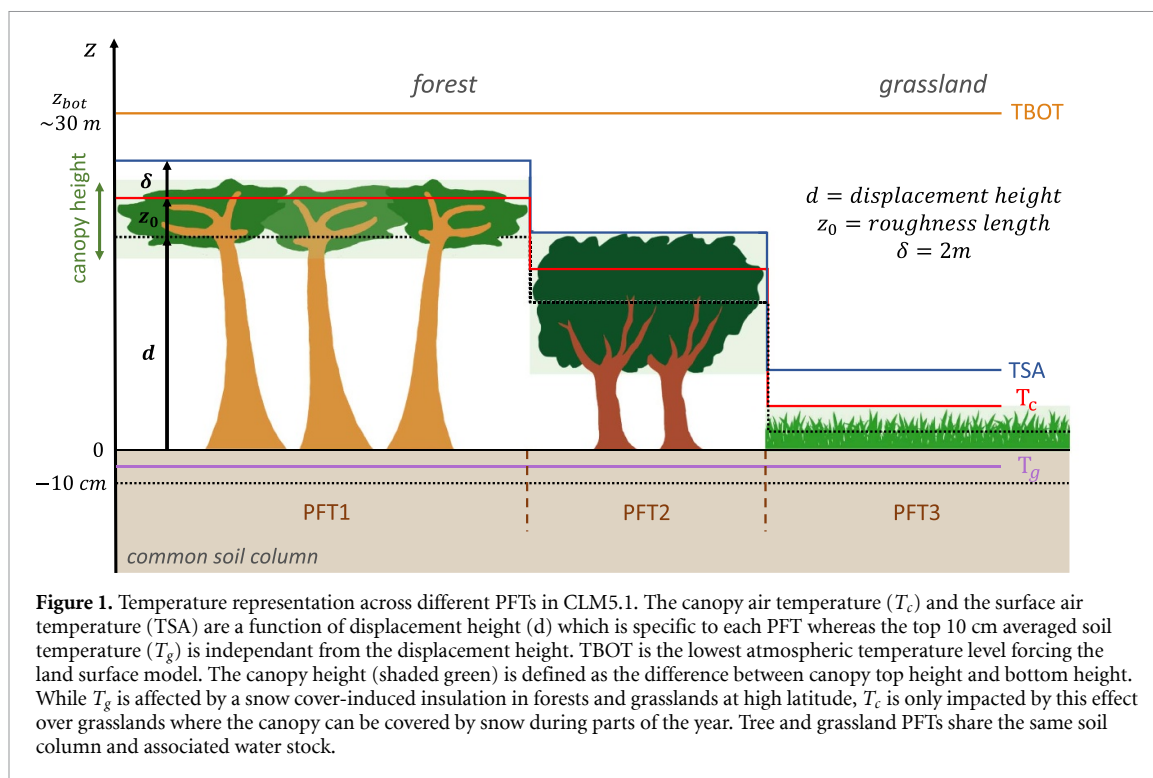
2.2. Experimental design

We performed two land-only simulations with CLM5.1 in SP mode at 0.5° by 0.5° resolution which together cover the 21st century (1995–2099). Before starting the first simulation, we ran the model over the 1990–1994 period (the spin-up phase) in order to reach a state of statistical equilibrium under the applied forcing. The spin-up of the full CLM5.1 model typically takes hundreds of years, but for SP mode this is not necessary because the carbon pools in this mode are prescribed and not prognostic, and thus only the physical state of the model (energy and water) needs to achieve an equilibrium. The first simulation is a historical run covering the 1995–2014 period for which we use the CLM5.1 default GSWPv3 reanalysis (Kim *et al* 2017) as atmospheric forcing. The second simulation is branched from the historical one and consists of a future SSP8.5 high emissions scenario over 2015–2099 (appendix figure B1). The required future atmospheric forcing was obtained by applying climate anomalies of a previous CESM simulation to the GSWPv3 2001–2013 climatology (Lawrence *et al* 2020). This offline approach allowed to focus on the warming effects on land (rather than the feedbacks on the atmosphere) and avoids running computationally-intensive coupled land-atmosphere simulations.

Both simulations employ the same static land-use map, omitting land use changes. The distribution of PFTs is based on MODIS satellite data (Lawrence and Chase 2007, Lawrence *et al* 2020). Monthly prescribed LAI and SAI are also derived from MODIS data (Myneni *et al* 2002) following the methods described in Lawrence and Chase (2007) and Zeng *et al* (2002), respectively. The fixed canopy top and bottom heights for trees are retrieved from the Geoscience Laser Altimeter System (GLAS) aboard the ICESat satellite and are PFT-independent (Simard *et al* 2011), while their counterparts for short vegetation are PFT-specific (Bonan *et al* 2002, Lawrence *et al* 2020). By ignoring future changes in land use or vegetation dynamics we can focus in this analysis on quantifying the evolution of forest microclimates under increasing warming independently of potential changes in future forest distribution and structure.

2.3. Model analysis and evaluation

In field studies, the forest microclimate is typically quantified using a wide range of temperature variables (soil temperature, 1 m air temperature or



canopy air temperature), depending on the technical feasibility and the research question. Given the single-layer canopy representation in CLM5.1, an estimated 2 m air temperature (TSA) was provided by interpolating between the atmospheric temperature given by the atmospheric forcing, and the surface temperature simulated in the model. Therefore, this 2 m air temperature diagnostic does not represent the shading or buffering effect of vegetation canopies, as also shown in Malyshev *et al* (2015). On the contrary, the soil temperature (T_g , °C) -the temperature average over the first 10 cm of soil- and the canopy air temperature (T_c , °C) -the surface air temperature defined within vegetation canopies- better represent the temperatures under and within the canopy respectively. T_g is calculated by solving the one-dimensional heat equation over the first soil layer and T_c is defined as a weighted mean of the atmospheric potential temperature, the vegetation temperature and the soil temperature (see appendix C.1). T_c depends on canopy height and represents the air temperature below the canopy top in forests and above canopy top over grasslands (figure 1). Therefore, T_c inside and outside forests is not at the same physical height. This has important implications at temperate and boreal latitudes where grass can be buried under snow during part of the year. In such cases, T_c is close to the temperature of the snow surface (Lawrence *et al* 2018). In contrast, the top soil layer T_g does not depend on canopy height and is thus simulated at the same physical height for every PFT. Therefore, it is subject to the same snow cover conditions over forest and grassland. In addition, most biodiversity inventories are

performed near the soil surface but future projections of soil temperature are typically lacking (Lembrechts *et al* 2022). For the boreal biome in particular, the soil environment below the snow cover is a key refuge for many organisms (Niittynen *et al* 2018, Kayes and Mallik 2020). Therefore, this paper focuses on T_g to provide the most relevant temperature projections for species dynamics. While T_g is probably the best CLM5.1 proxy for near soil air temperature within forests, additional factors such as heat capacity, soil moisture and snow cover dynamics can result in a decoupling of soil and air temperature (Lembrechts *et al* 2022). In this perspective, we also provide a complementary analysis for canopy temperature in the appendix C.1, to serve as a comparison in modelled responses and to discuss model development pathways for below canopy temperature representations.

Observations of forest microclimate are commonly expressed as a temperature difference between inside forest and outside forest environments that are subject to similar atmospheric forcings (De Frenne *et al* 2019, Lembrechts *et al* 2020). Likewise, we constructed a modeled offset temperature ΔT defined as the difference between the internal temperature of the forest tile minus the internal temperature of the grassland tile. For each grid cell, we computed the difference between the area-weighted average temperature in forests, and the area-weighted average temperature over grassland. Forest and grass areas were derived per the PFT distribution map. The 9 forest and 3 grassland PFTs are listed in appendix table A1.

Similar to De Lombaerde *et al* (2022), we defined the offset for minimum $\Delta T_{g,\min}$, maximum $\Delta T_{g,\max}$

and mean $\Delta T_{g,\text{mean}}$ soil temperature. $\Delta T_{g,\text{min}}$ (respectively, $\Delta T_{g,\text{mean}}$ and $\Delta T_{g,\text{max}}$) was derived by selecting the monthly minimum (respectively, mean and maximum) of the 6-hourly forest T_g and subtracting the simultaneous T_g over grassland within a grid cell. The distinction between forest, grassland and other vegetation types is possible through the subgrid tiling of CLM5.1, which allows to directly write the output per PFT, corresponding to the level at which the computations are conducted before being averaged at the grid cell level (Lawrence *et al* 2019). Thus, we considered all grid cells that contain both more than 0% of tree PFT cover (forest) and more than 0% of grass PFT cover (grassland). T_c offsets ($\Delta T_{c,\text{min}}$, $\Delta T_{c,\text{max}}$ and $\Delta T_{c,\text{mean}}$) were generated using the same method although we note that unlike for T_g , the differentiation between forest and grassland T_c also encapsulates a difference in height. In order to evaluate the effects of global warming on forest temperature offsets, we compared the historical climatology (1995–2014) to the future climatology (2080–2099). We focused our analysis on three major biomes (boreal, temperate and tropical) defined by the grid cells containing the corresponding tree PFTs (table A1). Geographically speaking, these categories roughly correspond to 3 latitudinal bands: between 50° and 70° for the boreal forest, 25° and 50° for the temperate forest and 0° and 25° for the tropical forest as shown on figure A1.

Finally, we constructed a global mean surface temperature anomaly relative to the 1850–1899 average (ΔGMT), based on HadCrut5 observational data (Morice *et al* 2021). This allows to assess soil and canopy air temperature offsets per degree of global warming.

3. Results

3.1. Annual mean offset patterns

The annual average soil temperature offset (negative when the soil surface is cooler below the forest canopy compared to below the grassland, and positive otherwise) over the historical period features a pronounced latitudinal gradient, but with opposite signs for maximum and minimum soil temperature (figure 2). $\Delta T_{g,\text{max}}$ increases with latitude, from very negative values (-6°C) in tropical regions to slightly positive values ($+0.5^\circ\text{C}$) in boreal regions indicating that tropical forests have a consistent buffering effect on forest understory maximum soil temperature whereas some boreal forests have a small amplifying effect. $\Delta T_{g,\text{mean}}$ shows a similar but smaller gradient (ranging between -2 and $+0.5^\circ\text{C}$) compared to $\Delta T_{g,\text{max}}$. $\Delta T_{c,\text{min}}$ was found to decrease with latitude, from slightly positive values ($+1^\circ\text{C}$) in the tropics to slightly negative values in the boreal region

(-2°C), implying a buffering of minimum T_g in tropical forests and an amplification in boreal forests. Overall, canopy air temperature offsets show similar latitudinal gradients during the historical period (figure E1), albeit with larger magnitudes, especially in boreal regions for which minimum T_c and maximum T_c are strongly amplified. However, T_c and T_g offsets differ in the tropics and subtropics where only maxima are dampened for T_c (figures E1(g) and (i)) whereas both extremes are systematically buffered for T_g (figures 2(g) and (i)).

Next, we compared the future period (2080–2099) forced by the high emission SSP5-8.5 scenario ($\sim+4.4\text{ K}$ in global mean surface temperature, (IPCC 2021)), to the historical one (period 1850–1900). We calculated the difference between the temperature offset in the future relative to the historical baseline (figures 2(d)–(f)).

We found a significant negative difference for maximum T_g in the tropics, suggesting that under high warming levels, tropical forests increase their buffering capacity for high temperature extremes by an average of 0.55°C . This result is underlined by the shift in the distribution of tropical $\Delta T_{g,\text{max}}$ figure 2(g). In addition, a small shift of $\Delta T_{g,\text{min}}$ towards a zero-centered distribution in boreal forests implies less amplification of minimum soil temperatures during the future period.

Future changes for canopy air temperature offsets (figure E1) are qualitatively similar but with even larger amplitudes than for soil temperatures. In tropical forests, we observe that $\Delta T_{c,\text{max}}$ is buffered by 1.16°C in the warmer period compared to the past period. We also found an increase of buffering capacities for boreal forests, decreasing the warming effect by $\sim 0.43^\circ\text{C}$ (figure E1(g)). Finally, $\Delta T_{c,\text{min}}$ remains very small in the future over the tropical and temperate regions, but increases significantly at boreal latitudes ($\sim 0.34^\circ\text{C}$ change) while staying overall negative. Most boreal regions show a pronounced increase in temperature offset (over 3°C on the Labrador peninsula), with forests further increasing minimum temperatures in the future, while Eastern Siberia shows negative values (figure E1(f)). This pattern reveals a contrasting biophysical response across the boreal region at higher warming levels with an overall reduced minimum T_c amplification (the minimum temperature offset is less negative in the future) opposed to an increased minimum T_c amplification over Siberia. However, this boreal dipole pattern is not visible on the $\Delta T_{g,\text{min}}$ future map (figure 2(f)), suggesting that it results from tree-atmosphere interactions rather than soil-tree interactions.

3.2. Seasonal offset patterns

Exploring the seasonality of temperature offsets, we find that annual patterns mask out a seasonal

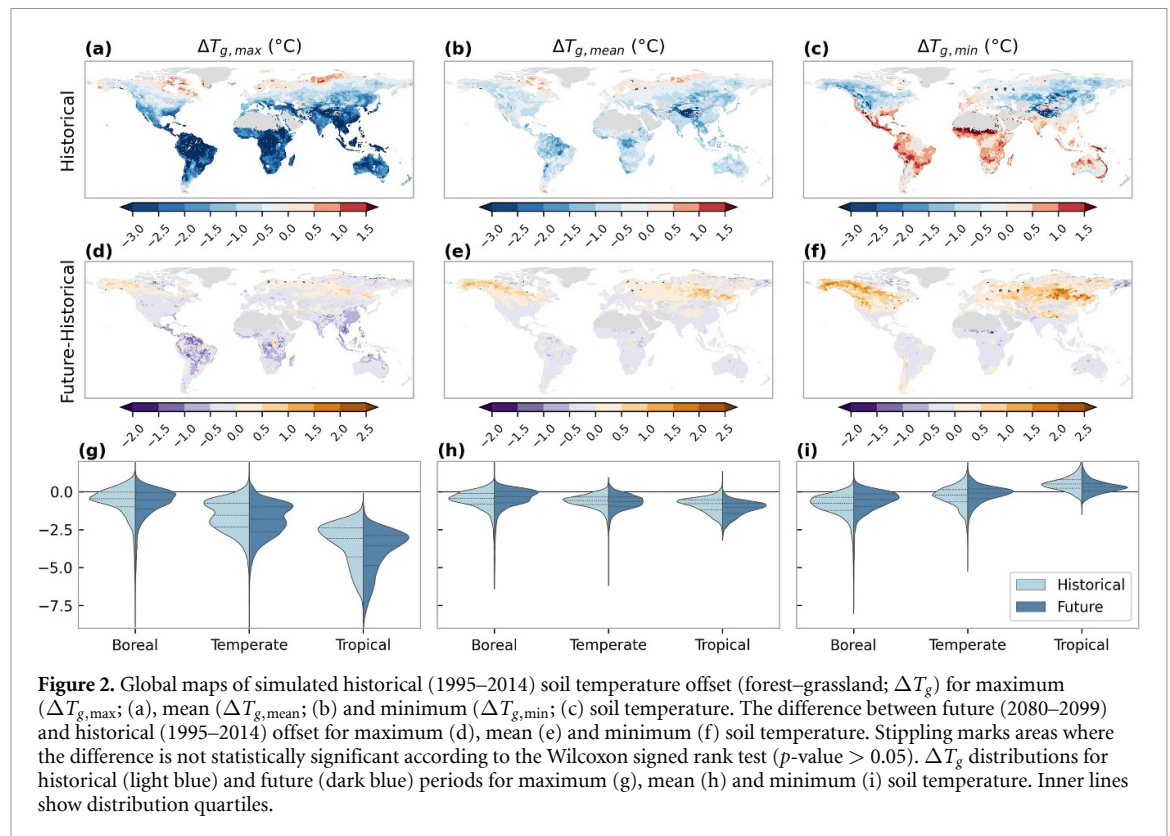


Figure 2. Global maps of simulated historical (1995–2014) soil temperature offset (forest–grassland; ΔT_g) for maximum ($\Delta T_{g,max}$; (a), mean ($\Delta T_{g,mean}$; (b) and minimum ($\Delta T_{g,min}$; (c) soil temperature. The difference between future (2080–2099) and historical (1995–2014) offset for maximum (d), mean (e) and minimum (f) soil temperature. Stippling marks areas where the difference is not statistically significant according to the Wilcoxon signed rank test (p -value > 0.05). ΔT_g distributions for historical (light blue) and future (dark blue) periods for maximum (g), mean (h) and minimum (i) soil temperature. Inner lines show distribution quartiles.

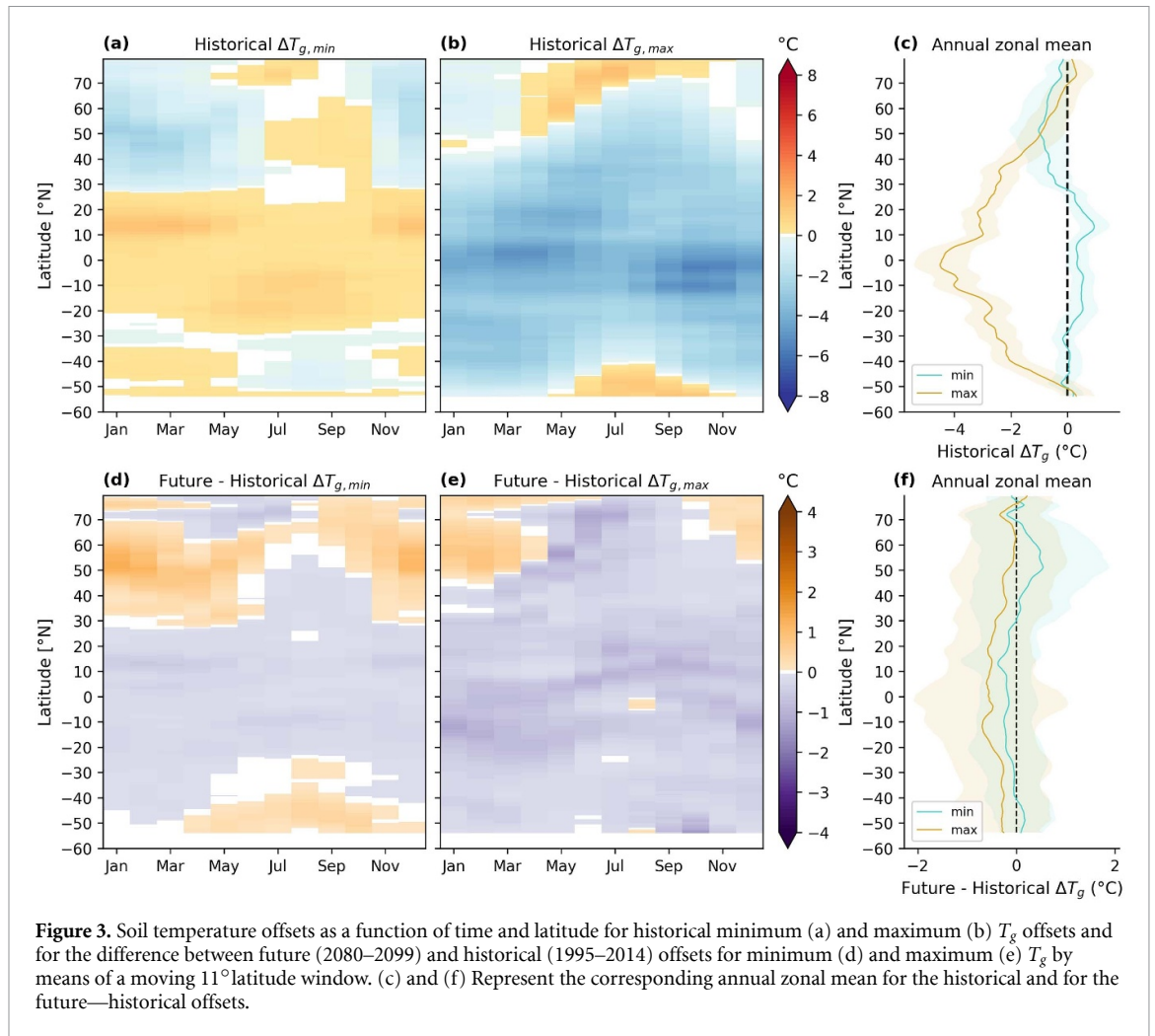
complexity that is subject to change with global warming.

The tropical region displays a strong year-round buffering of maximum T_g (negative $\Delta T_{g,max}$, figure 3) and, to a larger extent, of maximum T_c (figure E2). $\Delta T_{g,min}$ is slightly positive (whereas $\Delta T_{c,min}$ is null) indicating a buffering of the highest and lowest soil temperatures. The negative difference between future and historical $\Delta T_{g,max}$ in the tropical region suggests a stronger yearly buffering of soil temperatures at the end of the century (figure 3(e)). A similar increase in future buffering capacity is observed for maximum canopy air temperature (figure E2(e)). In the boreal region, our simulations suggest a clear seasonal pattern of the T_g offset (figure 3) even more pronounced for the T_c offset (figure E2). $\Delta T_{g,min}$ is negative most of the year (boreal autumn, winter and spring) revealing an amplification process. $\Delta T_{g,max}$ is positive during boreal spring and summer, highlighting an amplification effect of forest on soil temperatures, while winter soil temperatures are slightly buffered. The amplitude of the canopy air temperature offsets (figure 3) are much larger than for soil temperature and feature a slightly different pattern. $\Delta T_{c,min}$ is also negative during boreal winter and spring and slightly positive the rest of the year. Conversely, $\Delta T_{c,max}$ is positive over the same period, extending until July for the northernmost latitudes, while it is moderately negative during the remaining months. $\Delta T_{c,max}$ is particularly latitude-dependent with peak values

reached later in the year for higher latitudes. These results suggest that there is a strong amplification of minimum and maximum boreal forest T_c in winter and spring over the historical period.

The comparison with future T_g offsets in the boreal region (figures 3(d) and (e)) indicates less T_g amplification relative to the historical period, with increasing $\Delta T_{g,min}$ in winter and less positive $\Delta T_{g,max}$ in early spring. However, given the considerable spread both for $\Delta T_{g,min}$ and $\Delta T_{g,max}$ at high latitudes of the northern hemisphere, the signal in this region is not very robust (figure 3(f)). While $\Delta T_{g,max}$ is similar over both periods, $\Delta T_{g,min}$ is higher and much closer to zero during the future period, suggesting that there is little difference between soil temperatures inside forests compared to outside (figure 3(f)).

Figure 4 shows that the buffering of maximum T_g and T_c in tropical forests over the historical period further increases in the future global warming scenario (panel (f)). This figure also underlines the contrasted seasonal cycles for T_g and T_c offsets in temperate and boreal forests: (i) minimum T_c is strongly amplified in winter and spring compared to minimum T_g , (ii) maximum T_c is amplified whereas maximum T_g is unchanged during spring in boreal forests and (iii) maximum T_g is strongly buffered whereas maximum T_c is slightly amplified during spring in temperate forests. Finally, in addition to future shifts in temperature offsets discussed in figure 3, we



observe a lag effect for future maximum temperature offsets as compared to the historical period, especially in boreal and temperate forests for both variables. In other words, the increase of maximum T_g and T_c buffering occurs earlier in the year under higher global warming.

3.3. Biome-dependent trends in temperature offsets per global warming level

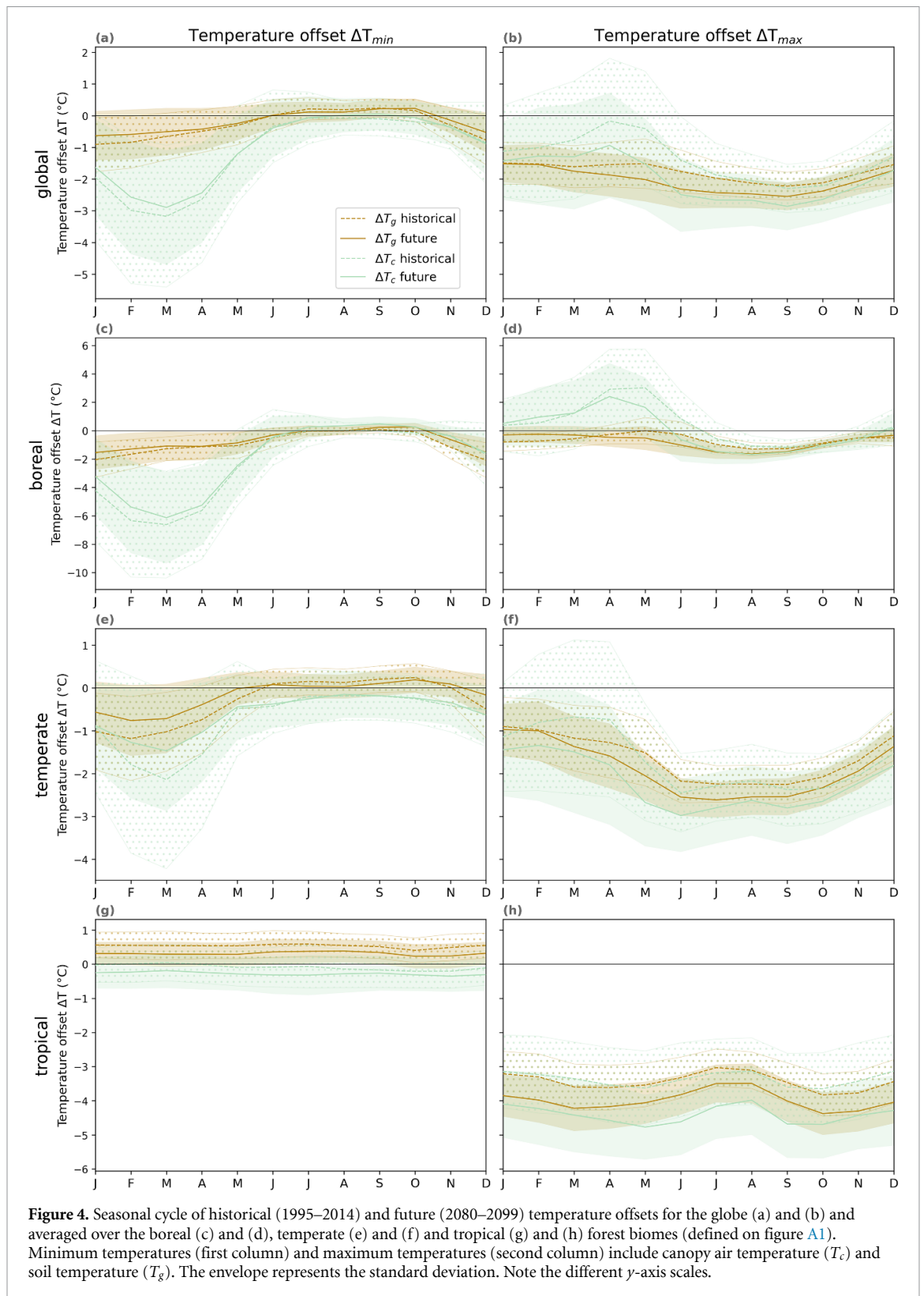
Our transient simulation, spanning the 21st century, highlights biome-dependent offset trends per degree of warming (figure 5). The relationship between offsets and macroclimate is found to be linear across the century for all biomes. The long-term multi-year averages highlight a significant increase of maximum T_g offset with global warming for all biomes, thereby confirming our hypothesis of stronger maximum temperature buffering in forest understories with increasing global temperatures. The magnitude of the maximum T_g buffering increase is biome-dependent and particularly strong for tropical forests with an offset trend of $-0.1^\circ\text{C}/^\circ\text{C}$. In contrast, minimum T_g offset trends vary across biomes with negative values for tropical forests and positive values

for temperate and boreal forests. Similar results are observed for maximum T_c offset. The maximum T_c offset trend is particularly strong in tropical forests with $-0.2^\circ\text{C}/^\circ\text{C}$ (figure E3).

4. Discussion

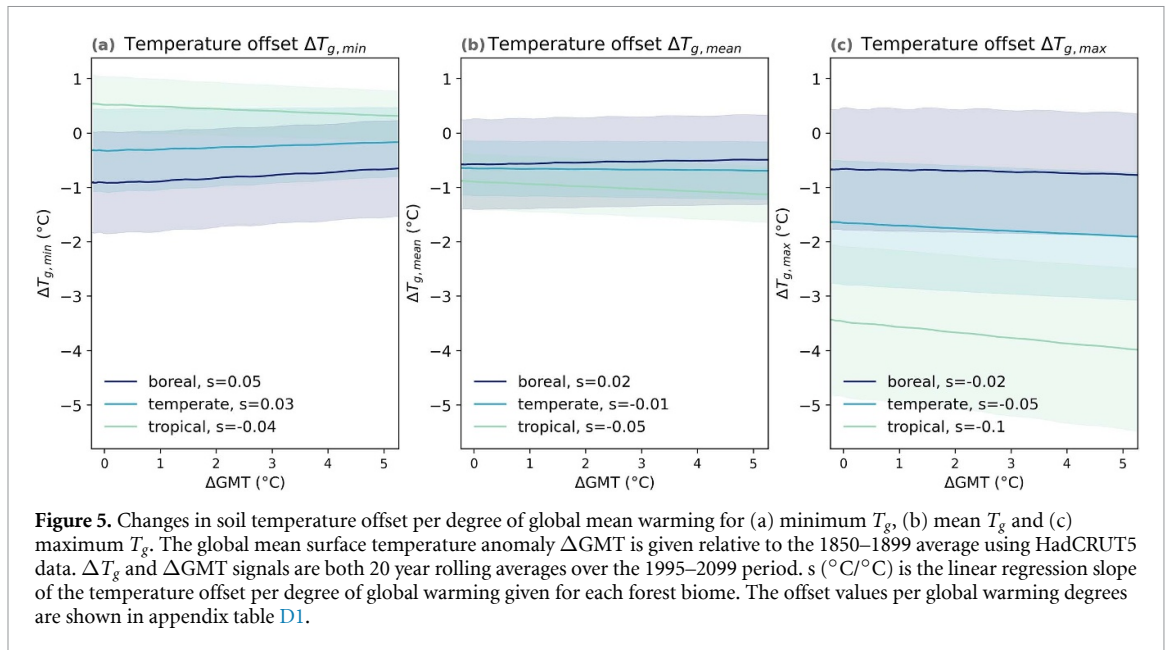
4.1. Contrasting seasonal and latitudinal effects of forests on temperature offsets

In this study, we provide a first process-based modeling assessment of global forest effects on microclimate under a warmer climate, complementing observation-based approaches (Haesen *et al* 2021, De Lombaerde *et al* 2022, Lembrechts *et al* 2022). For historical annual averages (figures 2 and E1), we find a positive latitudinal gradient for mean and maximum T_g and T_c offsets with a strong forest buffering effect in the tropics, consistent with observations by De Lombaerde *et al* (2022) and Lembrechts *et al* (2022). In the boreal region, the simulated $\Delta T_{g,\text{max}}$ matches the results of De Lombaerde *et al* (2022), with near-zero values, while our simulated $\Delta T_{c,\text{max}}$ indicates a strong amplification in the canopy. The main difference between our results and empirical



findings is observed for the minimum temperature offset: De Lombaerde *et al* (2022) show buffered minimum temperatures in boreal forests whereas we find that minimum T_c and T_g are amplified compared to grassland. The strong amplification of minimum canopy temperature at high latitudes can be explained

by multiple effects. First, the presence of snow over grassland canopy may buffer grassland T_c compared to forest T_c . Additionally, CLM5.1 is known to have 5 to 10 m taller canopies than observed for the boreal needleleaf evergreen PFT (Lawrence *et al* 2019) which accounts for the majority of tree PFTs at high



latitudes. Higher canopies generate stronger turbulence that decreases canopy temperature over these forested areas.

The simulated latitudinal gradients are the result of strong seasonal signals. Given the low incoming solar radiation during winter in boreal regions, variations in surface albedo between plant types have a negligible impact on T_g . Therefore, it is more likely that the simulated amplification of minimum T_g is caused by differences in surface roughness. The roughness length for tree PFTs is larger compared to grassland, and this is especially the case for snow-covered grassland (Meier *et al* 2021). Hence, over the forest canopy there is more turbulent mixing and therefore higher sensible heat fluxes compared to grassland, which decreases forest soil temperature. Given our definition of the canopy temperature offset, the negative wintertime $\Delta T_{c,min}$ at high latitudes significantly differs from seasonal observations by Lembrechts *et al* (2022) for which it is positive.

As daytime lengthens through spring, the surface albedo difference between forested and grassland areas increases in importance in the radiative budget. Qu and Hall (2014) have shown that there is a strong correlation between the surface albedo feedback and local surface warming, and that the largest correlations are located at increasing latitudes across springtime. Interestingly, the maximum T_c offset signal (figure E1(b)) follows such a pattern as spring progresses into summer, suggesting that the surface albedo feedback could be an important driver of forest canopy air temperature amplification in the boreal region. Several other studies (Lee *et al* 2011, Duveiller *et al* 2018) focusing on surface temperature also support a local forest-induced warming at

boreal latitudes. Given these complex effects at high latitude, future studies should pay attention to these PFT-dependant snow effects on canopy temperature. In particular, the effect of snow on grassland canopy could undermine the use of T_c for analyzing buffering effects in colder regions. We note here that the interactions between snow and vegetation are also held responsible for the low predictive quality of microclimate temperatures in Lembrechts *et al* (2022).

The simulated future increase of maximum T_g and T_c offsets in tropical forests corresponds to the expected response under global warming, assuming a space-for-time substitution and no constraint on water availability (De Frenne *et al* 2021). The simulated maximum T_g offset trend of -0.1 °C/°C is consistent with De Lombaerde *et al* (2022) but the values (ranging from ~ -3.5 °C to ~ -4 °C across the century (appendix table D1)) indicate less buffering than what they observe (~ -5 °C to ~ -5.5 °C). The observed linear relationship between forest microclimate and global mean temperature (figure 5) is linked to the model assumption of static forests in terms of distribution and surface. Finally, the seasonal shifts in future temperature offsets for temperate and boreal biomes could be attributed to the simulated snow dynamics featuring a shorter snow cover duration and a smaller fraction of snow-covered ground in the future (figure 4). We expect that simulating phenological responses to the changing climate could generate an extra lag effect on the temperature offset.

4.2. Potential impacts of maximum temperature offset trends on tropical biodiversity

Our results show that future maximum T_g offset change and the associated buffering is stronger in

tropical forests than for higher latitudes (figure 5). Given such a latitude-dependant trend, we identify two important implications for tropical biodiversity. First, tropical forests could become a key refugia for thermosensitive understory species relative to other forested biomes in a warmer world. Second, such species might accumulate an even higher climatic debt than their temperate forest counterparts for which microclimate buffering already explains part of species' persistence rather than migration (Bertrand *et al* 2016). Given that tropical species have narrower thermal niches than temperate ones (Deutsch *et al* 2008, Tewksbury *et al* 2008, Jirinec *et al* 2022), we expect that they may fare worse if tropical forests were to lose their buffering effect. In fact, despite continuous canopy buffering, Trew *et al* (2023) shows that understory microclimate has already undergone significant changes over the 1990–2019 period. Additionally, tropical canopies are currently approaching their critical temperature threshold and will most likely exceed it within RCP 8.5 (Doughty *et al* 2023), causing tree death and subsequent temperature buffering failure. The maximum T_g offset trends for boreal and temperate forests are relatively smaller than for tropical forests and the implications of such temperature changes on biodiversity would require more in depth research.

4.3. Perspectives and direction for modeling future forest microclimates

Forest microclimates are at the core of a large set of interdisciplinary research questions, including species' response to global changes, forest regeneration dynamics or adaptation strategies to preserve biodiversity (De Frenne *et al* 2021, Kemppinen *et al* 2023). Land surface models such as CLM5.1, linking atmosphere dynamics to land-based ecological processes within Earth System Models, are crucial tools to address these complex questions. Nevertheless, our study is subject to limitations, and addressing these through model development would improve the representation of microclimates in forests. The first important limitation here is the assumption that forests are static, i.e. their composition and structure have no temporal dynamics aside from seasonal fluctuations in LAI. This assumption allows to isolate and focus on the biophysical response of present-day forest cover in future climates, without considering additional uncertainties related to forest dynamics under a changing climate and the choice of land-use scenarios. In particular, the responses of LAI seasonality and phenological dates, which strongly modulate the surface energy balance, to global temperature increase is not covered by the SP mode (Park and Jeong 2021). How would our results change with an evolving phenology and physiology? This question is

out of the scope of this study and would require the use of a simulation with prognostic carbon pools, which would provide an additional layer of complexity by representing climate vegetation feedbacks, and make the simulations subject to existing biases in the processes simulating LAI, which can be substantial. In these circumstances, we could either expect less temperature buffering due to lower LAI or more temperature buffering induced by longer growing seasons and more CO₂-induced LAI.

Additionally, the present analysis limits describing temperature offset to temperature extremes, hereby omitting humidity extremes, which are acknowledged to have a considerable influence on forest biodiversity (Aguirre-Gutiérrez *et al* 2020). Studying the effects of droughts and the combination of humidity and temperature extremes are important avenues future research could explore.

The other limitations are inherent in the structure of CLM5.1 model itself. For instance, there is no horizontal transport of energy between tiles in the land model, the energy budget being solved for the vertical land-atmosphere interactions only. Although the new roughness parametrization used here allows for a more accurate vertical energy budget, the single layer canopy representation of forests in CLM5.1 does not consider vertical heterogeneity within the canopy of a given PFT (Bonan *et al* 2021). We describe some candidate future modeling pathway in appendix C.1. Finally, the 0.5° by 0.5° resolution used here (despite being high for LSM simulations) is too coarse to account for local structural heterogeneity of forest stands such as canopy gaps, vegetation density, proximity to rivers and topography known to be drivers of forest microclimates (Bramer *et al* 2018, De Frenne *et al* 2021, Malle *et al* 2021).

5. Conclusion

This study assesses the effect of global warming on forest understory temperatures in CLM5.1. We find a strong buffering of maximum soil and canopy air temperatures in tropical forests. This buffering effect increases with global warming suggesting that tropical species could accumulate a larger climatic debt than at higher latitudes, provided that trees remain alive. On the contrary, boreal forests slightly amplify maximum canopy air temperature compared to grassland areas. In boreal forests, our results also highlight a strong seasonal cycle with a large temperature amplification during boreal winter calling for more seasonal-based studies focusing on snow cover effects on microclimate. We encourage future LSM developments to include and assess additional microclimate drivers such as canopy structure, forest fragmentation or forest regeneration, which were out

of the scope of this study. To this end, the ecology, biogeography and climate science communities should determine collectively the most promising approaches for forest microclimate modeling and for use in evidence-based decision-making.

Data availability statement

The data that support the findings of this study are openly available at the following URL/DOI: <https://doi.org/10.5281/zenodo.8308545>.

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Appendix A. PFT based biomes

Table A1. PFT categorization of forest and grassland environments based on Lawrence *et al* (2020) documentation.

Environment	PFT	PFT index
Forest	Needleleaf evergreen tree—temperate	1
	Needleleaf evergreen tree—boreal	2
	Needleleaf deciduous tree—boreal	3
	Broadleaf evergreen tree—tropical	4
	Broadleaf evergreen tree—temperate	5
	Broadleaf deciduous tree—tropical	6
	Broadleaf deciduous tree—temperate	7
	Broadleaf deciduous tree—boreal	8
Grassland	C ₃ arctic grass	12
	C ₃ grass	13
	C ₄ grass	14

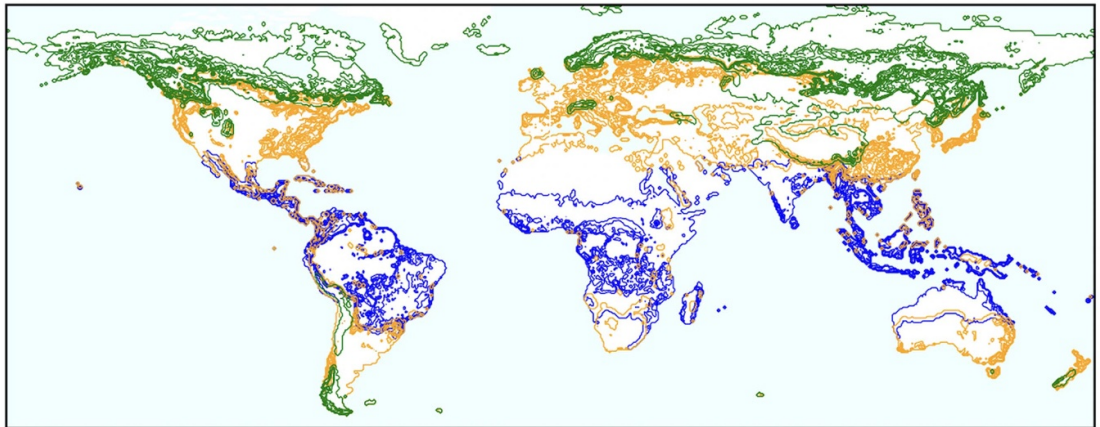


Figure A1. Map of PFT-based forest biomes used in this study. The boreal biome (red) is made of PFT indices 2, 3 and 8, the temperate biome (green) corresponds to PFT indices 1, 5 and 7 and the tropical biome (blue) encompasses PFT indices 4 and 6 (see table A1).

Appendix B. Model simulation experiment design

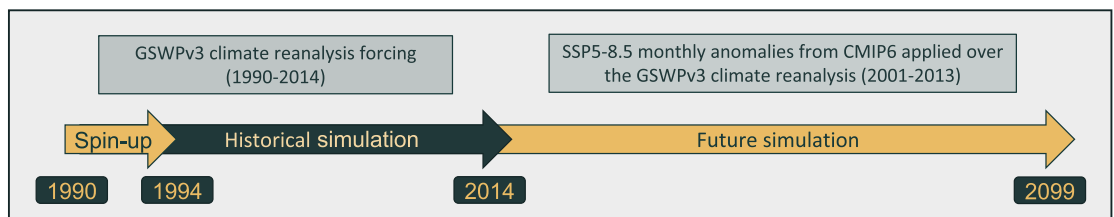


Figure B1. Model simulation experiment design. The spin-up phase allows CLM5.1 to reach a state of statistical equilibrium under the applied forcing. Following this, two land-only simulations are run using two different settings. The historical period follows the HIST_DATM%GSWP3v1_CLM51%SP_SICE_SOCN_MOSART_CISM2%NOEVOLVE_SWAV 'compset' (in CESM specific notation) applying the GSWPv3 reanalysis forcing. The future simulation follows the SSP585_DATM%GSWP3v1_CLM51%SP_SICE_SOCN_MOSART_SGLC_SWAV 'compset' integrating future SSP5-8.5 climate anomalies.

Appendix C. Reflections on canopy representation

C.1. Canopy air temperature in CLM5.1

In Clm5.1, T_c ('TAF' in CESM specific notation) is defined as a weighted mean of the atmospheric potential temperature (Θ_{atm} , °C), the vegetation temperature (T_v , °C) and the soil temperature (T_g , °C):

$$T_c = \frac{c_a^h \Theta_{\text{atm}} + c_g^h T_g + c_v^h T_v}{c_a^h + c_g^h + c_v^h} \quad (\text{C1})$$

where c_a^h , c_g^h and c_v^h are the sensible heat conductances from the canopy air to the atmosphere, from the ground to the canopy air, and from the leaf surface to the canopy air, respectively (ms^{-1}). Θ_{atm} is provided by the atmospheric temperature forcing TBOT at height z_{bot} following $\Theta_{\text{atm}} = \text{TBOT} + \Gamma_d z_{\text{bot}}$ where $\Gamma_d = 0.0098 \text{ Km}^{-1}$ is the negative of the dry adiabatic lapse rate. T_v represents the leaf temperature derived from the total sensible heat flux balance.

C.2. Potential pathways for improved canopy representations

Here we propose model development pathways to enhance the representation of forest microclimate processes. First, adopting a multilayer canopy representation was proven to improve the radiative balance across the canopy in CLM for point-based sites (Bonan *et al* 2021). Incorporating such vertical canopy descriptions at global scales could help to diagnose the relative role of roughness and hydrology processes for different forest biomes. On the

other hand, rerunning the present simulation with a dynamical vegetation would allow to question the evolution of forest microclimates considering ecological shifts, regeneration dynamics for different land-use scenarios and potential forest tipping points. CLM-FATES model is well suited for such experimentations because it adds a size- and age-structured representation of vegetation dynamics within the land surface model structure (Fisher *et al* 2018). In particular, the integration of a multilayer scheme in CLM-FATES would allow finer evaluation of age- and species-related effects on forest microclimates at multiple storey heights. Alternatively, building an understory air temperature variable as a weighted mean of T_g and T_c could prove a promising path for future CLM developments, provided that such a weighted mean can be easily constructed for all forest biomes and atmospheric stability condition. Ultimately, modeling forest microclimates requires a variety of spatial scales in order to account for the different driving processes. For instance, investigating the effect of forest fragmentation on forest microclimate would require a meter-scale resolution and the implementation of a distance metric to represent proximity to forest edges. Determining the level of complexity (including the number of processes and the degree of spatial resolution) needed to represent forest microclimate dynamics would be very valuable in order to optimize the balance between realism and energy consumption at the heart of our modeling issues. It might be that regional models are more appropriate than global models to provide forest microclimate projections.

Appendix D. Future temperature offsets per degree of global warming

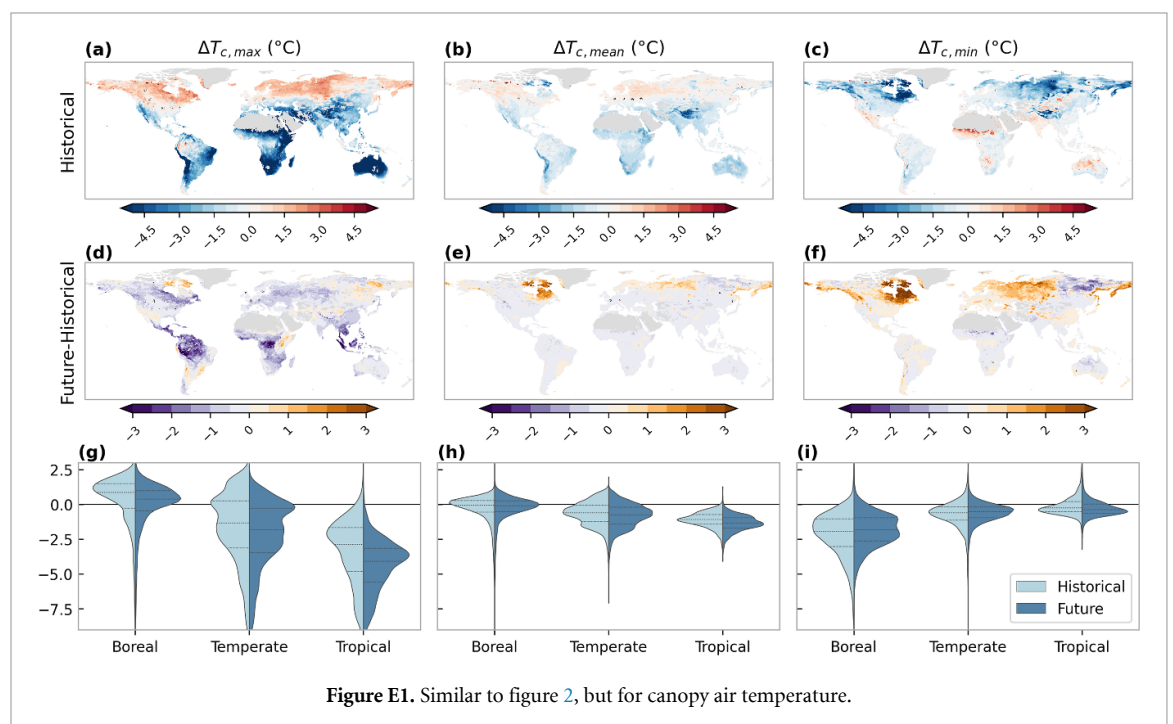
Table D1. Soil temperature minimum, mean and maximum offsets ($^{\circ}\text{C}$) per degree of global mean warming ΔGMT ($^{\circ}\text{C}$). Standard deviation is given in brackets.

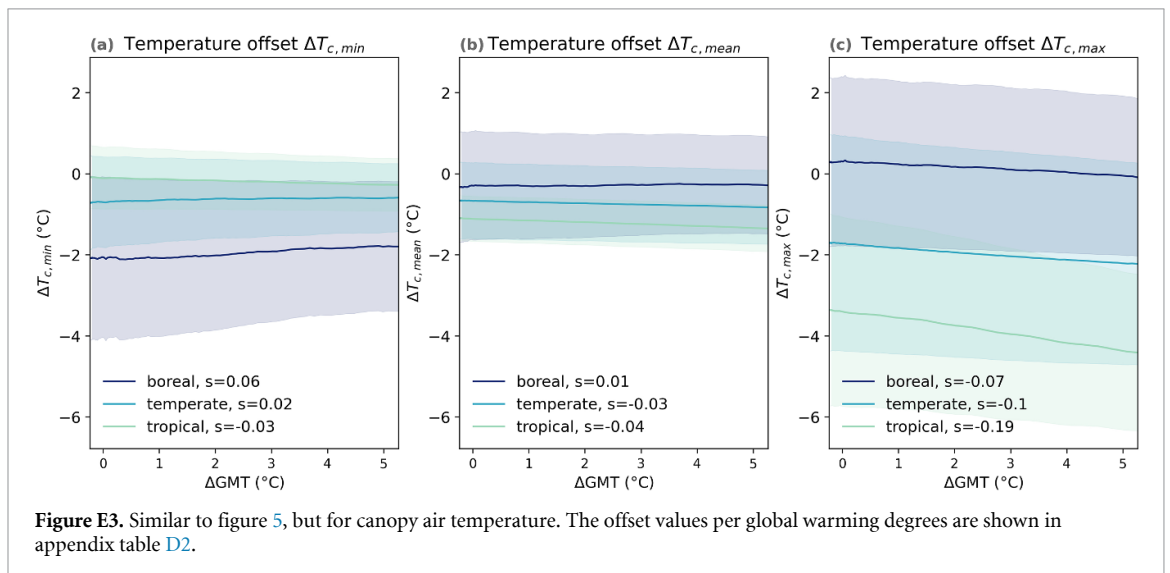
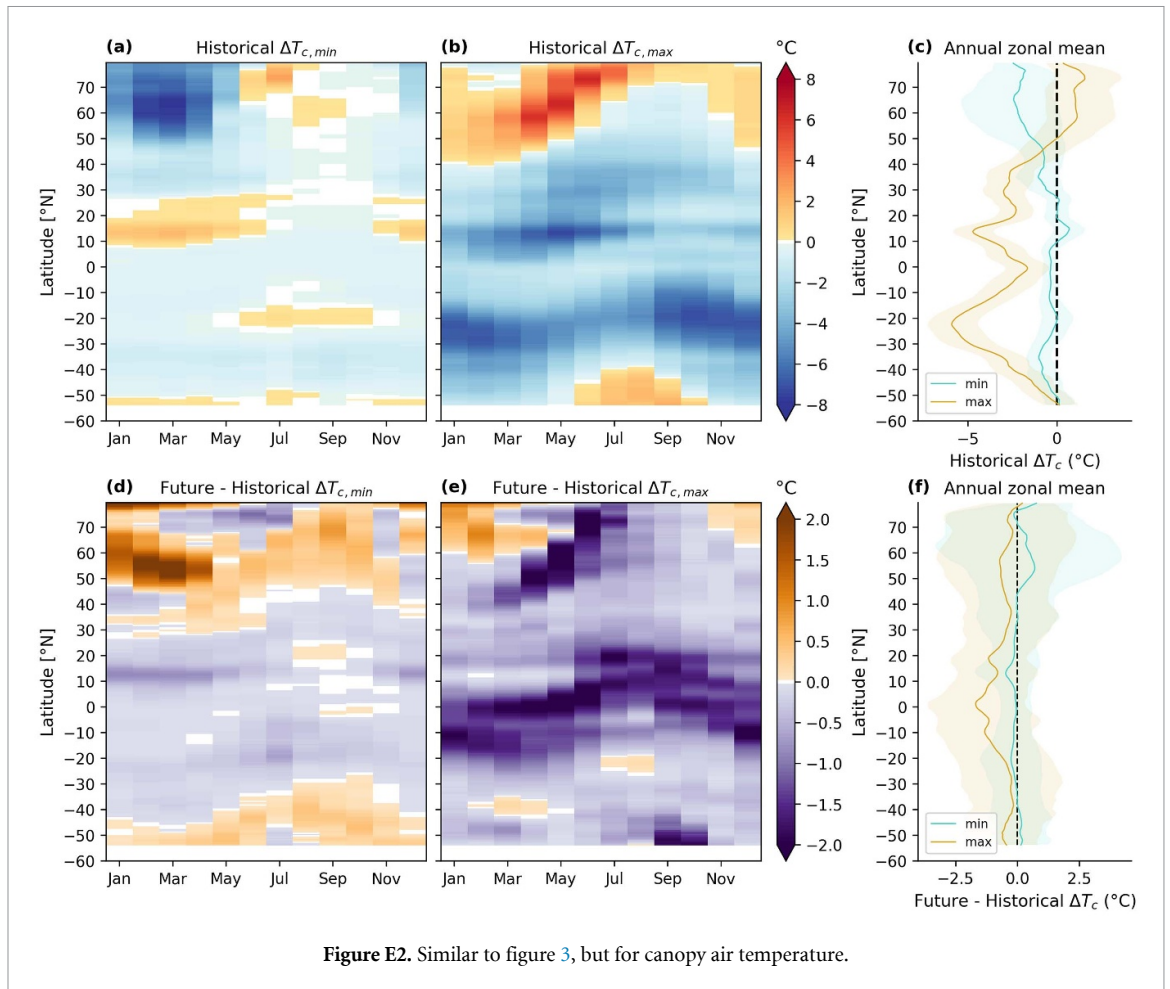
ΔGMT		0	1	2	3	4	5
Boreal	$\Delta T_{g,\min}$	-0.91 [0.94]	-0.89 [0.92]	-0.83 [0.91]	-0.78 [0.89]	-0.73 [0.89]	-0.67 [0.89]
	$\Delta T_{g,\text{mean}}$	-0.57 [0.84]	-0.57 [0.83]	-0.54 [0.82]	-0.53 [0.82]	-0.51 [0.82]	-0.49 [0.83]
	$\Delta T_{g,\max}$	-0.66 [1.12]	-0.68 [1.12]	-0.7 [1.13]	-0.72 [1.13]	-0.74 [1.13]	-0.76 [1.14]
Temperate	$\Delta T_{g,\min}$	-0.33 [0.76]	-0.3 [0.74]	-0.27 [0.71]	-0.24 [0.69]	-0.21 [0.67]	-0.18 [0.64]
	$\Delta T_{g,\text{mean}}$	-0.65 [0.5]	-0.66 [0.5]	-0.67 [0.51]	-0.67 [0.52]	-0.68 [0.52]	-0.69 [0.53]
	$\Delta T_{g,\max}$	-1.65 [1.13]	-1.7 [1.15]	-1.75 [1.16]	-1.8 [1.16]	-1.85 [1.17]	-1.9 [1.17]
Tropical	$\Delta T_{g,\min}$	0.52 [0.52]	0.49 [0.5]	0.44 [0.49]	0.4 [0.47]	0.36 [0.46]	0.32 [0.46]
	$\Delta T_{g,\text{mean}}$	-0.9 [0.51]	-0.94 [0.51]	-0.99 [0.51]	-1.03 [0.51]	-1.08 [0.51]	-1.12 [0.51]
	$\Delta T_{g,\max}$	-3.47 [1.38]	-3.57 [1.4]	-3.67 [1.42]	-3.77 [1.44]	-3.87 [1.47]	-3.96 [1.48]

Table D2. Canopy air temperature minimum, mean and maximum offsets ($^{\circ}\text{C}$) per degree of global mean warming ΔGMT ($^{\circ}\text{C}$). Standard deviation is given in brackets.


ΔGMT		0	1	2	3	4	5
Boreal	$\Delta T_{c,\min}$	-2.06 [1.98]	-2.08 [1.94]	-2.02 [1.84]	-1.92 [1.74]	-1.85 [1.65]	-1.8 [1.61]
	$\Delta T_{c,\text{mean}}$	-0.3 [1.34]	-0.31 [1.31]	-0.3 [1.27]	-0.28 [1.24]	-0.27 [1.22]	-0.28 [1.21]
	$\Delta T_{c,\max}$	0.3 [2.09]	0.23 [2.05]	0.17 [2.03]	0.1 [2.02]	0.04 [1.99]	-0.05 [1.97]
Temperate	$\Delta T_{c,\min}$	-0.71 [1.12]	-0.66 [1.04]	-0.62 [0.97]	-0.61 [0.93]	-0.6 [0.89]	-0.61 [0.85]
	$\Delta T_{c,\text{mean}}$	-0.67 [0.94]	-0.7 [0.94]	-0.73 [0.93]	-0.76 [0.93]	-0.79 [0.93]	-0.83 [0.91]
	$\Delta T_{c,\max}$	-1.72 [2.65]	-1.83 [2.62]	-1.94 [2.59]	-2.04 [2.57]	-2.12 [2.55]	-2.21 [2.5]
Tropical	$\Delta T_{c,\min}$	-0.11 [0.76]	-0.13 [0.74]	-0.17 [0.71]	-0.2 [0.7]	-0.24 [0.67]	-0.28 [0.65]
	$\Delta T_{c,\text{mean}}$	-1.12 [0.56]	-1.16 [0.56]	-1.2 [0.56]	-1.24 [0.57]	-1.29 [0.57]	-1.34 [0.58]
	$\Delta T_{c,\max}$	-3.4 [2.33]	-3.56 [2.26]	-3.75 [2.19]	-3.96 [2.12]	-4.18 [2.03]	-4.37 [1.95]

Appendix E. Complementary figures for canopy air temperature (T_c)






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