Differing Responses of the Diurnal Cycle of Land Surface and Air Temperatures to Deforestation^①

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(Manuscript received 2 January 2019, in final form 20 July 2019)

ABSTRACT

Recent studies have shown the impacts of historical land-use land-cover changes (i.e., deforestation) on hot temperature extremes; contradictory temperature responses have been found between studies using observations and climate models. However, different characterizations of surface temperature are sometimes used in the assessments: land surface skin temperature T_s is more commonly used in observation-based studies while near-surface air temperature T_{2m} is more often used in model-based studies. The inconsistent use of temperature variables is not inconsequential, and the relationship between deforestation and various temperature changes can be entangled, which complicates comparisons between observations and model simulations. In this study, the responses in the diurnal cycle of summertime T_s and T_{2m} to deforestation are investigated using the Community Earth System Model. For the daily maximum, opposite responses are found in T_s and T_{2m} . Due to decreased surface roughness after deforestation, the heat at the land surface cannot be efficiently dissipated into the air, leading to a warmer surface but cooler air. For the daily minimum, strong warming is found in T_{2m} , which exceeds daytime cooling and leads to overall warming in daily mean temperatures. After comparing several climate models, we find that the models agree in daytime land surface (T_s) warming, but different turbulent transfer characteristics produce discrepancies in T_{2m} . Our work highlights the need to investigate the diurnal cycles of temperature responses carefully in land-cover change studies. Furthermore, consistent consideration of temperature variables should be applied in future comparisons involving observations and climate models.

1. Introduction

Historical human land-use activities have affected over 50% of the land surface, in which over 25% of forests have been permanently cleared, and over 30% of the land is occupied by agriculture (Hurt et al. 2011). These land-use land-cover changes (LULCCs) have substantial impacts on global climate by altering the carbon inventory and carbon fluxes and thereby modifying the global atmospheric CO_2 concentration (Pongratz et al. 2010; Houghton et al. 2012). Meanwhile,

regional climate can be thoroughly affected through biogeophysical feedbacks, such as modifying albedo, surface roughness, and evapotranspiration efficiency, which may predominate over biogeochemical feedbacks especially over the middle and high northern latitudes (Claussen et al. 2001; Brovkin et al. 2006; Mahmood et al. 2014). At low latitudes, tropical deforestation generally leads to an annual mean warming due to decreased evapotranspiration, while temperate and boreal deforestation leads to an overall cooling driven mainly by increased surface albedo (Lawrence and Chase 2010; Davin and de Noblet-Ducoudré 2010; Pitman et al. 2012). Besides the annual mean temperature, temperature extremes have drawn increasing attention due to their devastating impacts on human societies and ecosystems. Recent studies have suggested a large degree of impact of historical LULCC (i.e., deforestation) on hot

^① Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-19-0002.s1>.

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temperature extremes (Lejeune et al. 2018; Li et al. 2018; Findell et al. 2017; Chen and Dirmeyer 2019; Stoy 2018).

To understand the mechanism of temperature changes, land surface temperature (or skin temperature, T_s) is commonly used in observational studies. The T_s value is readily retrieved from satellite observations and is usually used to estimate the local and regional impacts of LULCC on surface temperature (Li et al. 2015; Bright et al. 2017; Schultz et al. 2017), especially in areas that are not easily accessible to researchers. Local warming by deforestation on summer daytime T_s has been found in most of these studies, implying that deforestation can intensify hot extremes (Stoy 2018). Based on the net land surface energy balance, the change in T_s may be directly attributed to different components of land surface change (such as albedo, Bowen ratio, and surface roughness; Lee et al. 2011; Chen and Dirmeyer 2016).

However, temperature extremes discussed in model-based studies are mainly derived from the near-surface air temperature (called T_{2m} hereafter), a common meteorological variable measured at weather stations. Note that T_{2m} is not a model prognostic variable, but rather is diagnosed empirically based on assumptions about the vertical air temperature profile near the surface. Moreover, the definition of T_{2m} can be different across observations and models. The term T_{2m} is referred to as the temperature at 2 m above ground (Pulliainen et al. 1997), but it is calculated in some models [such as the Community Land Model (CLM)] 2 m above the apparent sink for sensible heat (i.e., 2 m above the canopy top for vegetated surfaces; Oleson et al. 2010) unless otherwise noted. If T_{2m} is calculated at 2 m above ground, for the forest land-cover uncertainties of temperature profiles within forest canopies should be considered (Flerchinger et al. 2015). Not surprisingly, disagreements in T_{2m} responses to deforestation are often found among climate models (Li et al. 2018; Lejeune et al. 2017; Chen and Dirmeyer 2019). The majority of models in phase 5 of the Coupled Model Intercomparison Project (CMIP5) reveal cooling effects of deforestation on summer daily maximum temperature (Lejeune et al. 2017), which are contradictory to the deforestation-induced warming in T_s -based observations.

Given the inconsistency between the two temperature variables being used, the relationship between deforestation and temperature can be misconstrued, which also casts doubt on the comparisons between observations and model simulations. Therefore, this paper attempts to address the following questions: 1) Are there consistent changes between T_s and T_{2m} in response to historical LULCC in the climate models? 2) How does

the diurnal cycle of the temperature change? To answer these two questions, we investigate the diurnal cycle of the deforestation-induced changes in T_s and T_{2m} using simulations with the Community Earth System Model, which are compared with similar simulations from other models in CMIP5. Section 2 describes the experiment design and datasets used in our analysis. Results are presented in section 3. Discussions and conclusions are given in section 4.

2. Methodology

a. LULCC experiments in CESM

The LULCC experiments are conducted with the Community Earth System Model (CESM). CESM is a coupled Earth system model composed of separate climate system components for atmosphere, ocean, land, sea ice, and land ice (Versteinsten et al. 2013). Due to the focus on land-atmosphere interactions from the biogeophysical climate perspective, only the Community Atmosphere Model (CAM4.0; Neale et al. 2010) and Community Land Model (CLM4; Oleson et al. 2010) with prescribed satellite-based vegetation physiology are used in our simulations (with the component set F_2000).

Two sets of historical LULCC experiments are carried out at two spatial resolutions. In the high-resolution ($0.47^\circ \times 0.63^\circ$) experiment, the preindustrial (1850) and present (2000) land-cover conditions are generated based on the standard land-cover datasets provided in CESM (Lawrence et al. 2012). However, we only keep the dominant plant functional type (PFT) within each grid cell, so that there is a single explicit land cover assigned for each grid with no subgrid tiles. Changes in tree cover from 1850 to 2000 are shown in Fig. 1, which suggest that both deforestation (decreases in tree PFTs and increases in grass and crop PFTs) and afforestation (increases in tree PFTs and decreases in crop PFTs) have happened during the historical period. Because of the explicit configuration of the land cover, all the land-cover changes (e.g., deforestation or afforestation) occur over 100% of grid cell area, and thereby the signal of LULCC can be maximized. In the low-resolution ($1.9^\circ \times 2.5^\circ$) experiment, the land-cover conditions are aggregated from the high-resolution land surface data with subgrid PFT configuration. To ensure the two resolutions have the identical aggregate LULCC conditions at $1.9^\circ \times 2.5^\circ$ resolution, the percentage of each PFT for each single grid is calculated based on the PFT of the 16 high-resolution grids that fall within each low-resolution grid. Changes in tree cover for the low-resolution experiment are also shown in Fig. 1. The

FIG. 1. Percentage changes in tree PFTs from 1850 to 2000 prescribed in high-resolution and low-resolution LULCC experiments in three regions (NA: North America; EU: Europe; AS: Asia). The global maps of the tree PFT changes are shown in Fig. S1.

subgrid land-cover configuration in the low-resolution experiments is commonly used in current LULCC studies (including CMIP5). The comparison between high-resolution and low-resolution results would confirm or refute the null hypothesis that responses are invariant with resolution (LULCC responses are effectively linear) and if the findings in the high-resolution simulation may be comparable to other climate models.

All experiments are climatology simulations with cyclic circa year 2000 forcing, having identical prescribed forcings [e.g., sea surface temperature (SST), sea ice, CO_2 concentration, and aerosol] representative of the late twentieth century. There is a fixed CO_2 concentration of 367.0 ppm. The prescribed SSTs are a monthly mean climatology calculated from 1982 to 2001 (Hurrell et al. 2008), which is derived from a merged product based on the monthly mean Hadley Centre Sea Ice and SST dataset, version 1 (HadISST1), and version 2 of the NOAA weekly Optimum Interpolation SST (OISST2) analysis. The land initial conditions for each experiment are generated through 40-yr offline CLM simulations, and then used for the coupled simulations. We run all simulations for 60 years saving 3-hourly output.

In this study, T_s is the skin temperature (or radiative surface temperature), which is calculated based on the outgoing longwave radiation. Over the vegetated land, it is assumed to be at the canopy top, which is consistent

with the satellite-derived land surface temperature (Jin and Dickinson 2010). The term T_{2m} is the air temperature 2 m above the apparent sink for sensible heat (i.e., the canopy height; Oleson et al. 2010). We identify the timing of summer daily maximum (minimum) T_s , and the corresponding surface fluxes and T_{2m} when T_s reaches the daily maximum (minimum). We admit that the timing of daily maximum (or minimum) T_{2m} and T_s can be different. However, air temperature changes calculated based on the identified T_{2m} (corresponding to the maximum or minimum T_s) are the same as that from daily maximum (or minimum) T_{2m} (not shown). To keep consistent timing among all the variables, the identified T_{2m} when T_s reaches the daily maximum (minimum) are used in our analysis.

Additionally, historical deforestation has mainly occurred in the Northern Hemisphere (Fig. S1 in the online supplemental material); and our goal is to investigate how the hot temperature extremes are associated with deforestation. Therefore, our analysis is focused on the summertime temperate and boreal forest responses in the Northern Hemisphere.

b. CMIP5 land-use experiments

To determine whether the results reflect biases from the choice of model, we also examine the simulations from five CMIP5 models, which provide the necessary

TABLE 1. Information about the CMIP5 simulations. For the historicalMisc simulations, LU is land-use forcing, noLU is all forcings except land use. The LULCC-induced changes (LU – Control) in temperature and surface fluxes can be estimated based on the difference between the control run and LU run. Spatial resolution refers to the approximate size of grid cells in the global model grid.

Model	Land surface model	Spatial resolution (lat × lon)	Temporal resolution	Control run	LU run
GFDL-ESM2M	GFDL- LM3	2° × 2.5°	3-hourly	piControl	historicalMisc_LU
IPSL-CM5A-LR	ORCHIDEE	1.875° × 3.75°	3-hourly	historicalMisc_noLU	Historical
CanESM2	CTEM	2.8° × 2.8°	Monthly	piControl	historicalMisc_LU
CCSM4	CLM4	0.9° × 1.25°	Monthly	piControl	historicalMisc_LU
CESM-CAM5	CLM4	0.9° × 1.25°	Monthly	piControl	historicalMisc_LU

piControl, historical, and land-use related historicalMisc simulations for sensitivity calculations (Table 1). The piControl simulations are preindustrial control runs with all the forcings fixed at the year 1850 values. The historical simulations are forced by observed atmospheric composition changes (reflecting both anthropogenic and natural sources) including time-evolving land cover. The historicalMisc simulations are forced by various combinations of anthropogenic or natural agents (Taylor et al. 2012). For instance, there are simulations with land-use forcing only (LU) and simulations with all forcing agents but without considering land use (noLU). The LU simulations only consider the time-evolving land cover from 1850 to 2005 with other forcings fixed at their 1850 values. The noLU simulations consider all other forcings but with land cover fixed at its 1850 values. The signals of historical LULCC can be identified based on the difference between the LU and piControl simulations. Because IPSL-CM5A-LR does not have a LU simulation available, the impacts of LULCC are estimated based on the difference between the historical and noLU simulations. The former (LU minus piControl) is close to the configuration in our own LULCC experiments in CESM, described in section 2a. For each experiment of each CMIP5 model, the last 60 years of simulation (e.g., the period of 1941–2000 for the historical and historicalMisc simulations) from the first ensemble run are used in our analysis.

c. Temperature decomposition

Based on the surface energy balance, the LULCC-induced T_s change can be divided into different components that are associated with the changes in surface radiation and fluxes (Juang et al. 2007; Luyssaert et al. 2014; Vanden Broucke et al. 2015; Chen and Dirmeyer 2016). In this decomposition, the T_s change is attributed to a linear combination of changes in the terms for surface albedo, incoming shortwave and longwave radiation, and latent, sensible, and ground heat fluxes. This approach was first proposed by Juang et al. (2007) and

then further modified by Luyssaert et al. (2014). Equation (1) shows the surface energy balance. By reordering Eq. (1) and performing a first-order derivative, the discrete changes of ΔT_s can be expressed as Eq. (2):

$$R_n = (1 - \alpha_s)SW_{in} + LW_{in} - \varepsilon\sigma T_s^4 \\ = H + LE + G, \quad (1)$$

$$\Delta T_{s_reconstructed} = \frac{1}{4\sigma T_s^3}[-SW_{in}\Delta\alpha_s + (1 - \alpha_s)\Delta SW_{in} \\ + \Delta LW_{in} - \Delta LE - \Delta H - \Delta G], \quad (2)$$

where R_n is net surface radiation, SW_{in} is incoming shortwave radiation, LW_{in} is incoming longwave radiation, ε is surface emissivity, which is assumed to be 1 in CLM (Oleson et al. 2010), σ is the Stephan–Boltzmann constant, α_s is surface albedo, H is sensible heat flux, LE is latent heat flux, and G is ground heat flux; also, Δ indicates the LULCC-induced changes from 1850 to 2000, and $\Delta T_{s_reconstructed}$ is the reconstructed T_s change based on a first-order temperature decomposition.

3. Results

a. Temperature responses

To demonstrate the contrast between the deforested and afforested areas, the diurnal cycle of temperature changes is shown in Fig. 2. After tree loss, there is a significant warming at the land surface with the strongest warming during the daytime hours. However, a daytime cooling and strong nighttime warming are found in T_{2m} , leading to a slight warming in daily mean temperature. Both the daily mean T_s and T_{2m} exhibit warming effects from deforestation even though their magnitudes are different. Without investigating the diurnal cycle of temperature changes, which are closely related to the hot extremes, the discrepancies between the daily maximum T_s and T_{2m} cannot be adequately revealed.

FIG. 2. Changes in the diurnal cycle of (top) T_s and (bottom) T_{2m} over three regions of deforestation and two regions of afforestation in the Northern Hemisphere. The deforested (afforested) areas are defined as the grid cells with complete tree loss (gain) in the high-resolution simulations (shown as Fig. 1). Global maps of the temperature changes are shown in Figs. S2 and S3.

Over the afforested areas in North America and Europe, which are the only two regions with significant tree gain from 1850 to 2000 according to Fig. 1, changes in temperatures are opposite those for the deforested areas. An overall cooling is found in daily mean T_s and T_{2m} . During the daytime, afforestation results in cooling

of the land surface but warming of the air, further demonstrating the discrepancies in the daily maximum T_s and T_{2m} . Even though other types of LULCC (such as the replacement of grassland with cropland) also occur during the historical period, neither the temperature changes nor the contrasts between T_s and T_{2m} are

FIG. 3. Changes in daily (top) maximum and (bottom) minimum T_s averaged over the deforested and afforested areas in the Northern Hemisphere. T_s (red) is the T_s change directly calculated from the deforestation experiments; $T_{s_reconstructed}$ (blue) is the reconstructed T_s change based on surface energy balance, which is the sum of the contributions from different components (including surface albedo, incoming shortwave and longwave radiation, latent heat flux, sensible heat flux, and ground heat flux). The difference between the simulated and reconstructed T_s changes is shown as the gray bar.

as evident as in the regions with tree loss or gain (not shown).

b. Mechanism of T_s changes

To understand the mechanism of daytime T_s changes in response to tree cover changes, we analyze the contribution of different surface features on T_s based on the temperature decomposition averaged over all deforested (or afforested) areas (Fig. 3). The reconstructed T_s agrees well with the simulated T_s , indicating that the temperature decomposition can well represent the changes in T_s through the surface energy budget. Deforestation leads to an overall warming by +1.85 K (+2.19 K based on the temperature decomposition) in daily maximum T_s , while afforestation leads to an overall cooling by -1.86 K (-2.17 K based on the temperature decomposition).

Generally, the cooling effects of increased albedo are well balanced by warming effects of incoming radiation, in which both incoming shortwave and longwave radiation increase due to the cloud-radiative effects; and nonradiative effects play a more important role in the T_s change. Among the nonradiative factors, decreased sensible heat flux caused by reduced surface roughness, and less efficient turbulent heat exchange with the atmosphere can cause a strong surface warming (about +7.41 K) over the deforested areas. There is a smaller warming associated with decreased latent heat flux, which is the result of both roughness changes and contrasting water conduction in different PFTs. Most of the

remaining heat is conducted into the soil by increased ground heat flux, which is typically calculated after other heat flux terms by land surface models. Afforestation shows the same pattern but with opposite signs. Therefore, surface roughness, which strongly influences surface turbulent fluxes, is the major contributor of deforestation-induced T_s warming.

It should be noted that the relatively small warming from decreased latent heat flux does not necessarily indicate that it has little contribution to the surface temperature change. The change in latent heat flux exhibits large spatial variability (Fig. S4), which is determined by the moisture conditions and the type of land-cover change. For instance, over the eastern United States and eastern China, which is in a humid climate with relatively high soil moisture, deforestation (from needleleaf evergreen temperate tree to crop) leads to increased soil evaporation and increased canopy transpiration, which exceeds the decreased canopy interception evaporation, resulting in the overall increased latent heat flux. However, over the central United States and eastern Europe, where soil moisture is lower, the same type of land-cover change does not lead to decreased latent heat flux mainly because the limited soil moisture may reduce the crop transpiration. In India and southeastern Asia, where deforestation leads to a decrease in broadleaf deciduous tropical tree and an increase in crops, there is decreased canopy evapotranspiration (evaporation and transpiration), which exceeds the increased soil evaporation, therefore resulting

FIG. 4. Changes in daytime sensible heat flux (H), aerodynamic resistance (r_a), and temperature gradient ($T_{\text{gradient}} = T_s - T_a$) over the deforested and afforested areas: r_a and T_{gradient} are shown as percentage changes so that their relative contributions to the changes in H can be compared.

in the overall decreased latent heat flux. Because of the inconsistent changes of latent heat flux, the value of its spatial average across the deforested areas can be relatively small compared with the change in sensible heat flux, which by contrast shows a consistent decrease over all the deforested areas.

We also examine the decomposition of daily minimum T_s (Fig. 3b). Deforestation leads to a slight warming ($+0.86$ K) at the surface during the night, which is mainly associated with increased heat flux from the subsurface soil compensating the reduction in sensible heat exchange. It should be noted that sensible heat flux becomes negative during the night (Fig. S5), indicating the heat is transferred from the lower atmosphere to the land. Deforestation reduces surface roughness and increases aerodynamic resistance, so the heat transfer is less efficient over the open land than forest, leading to less of a warming effect on the land surface. Again, the opposite effect is seen for afforestation.

c. Relationship between the changes in T_s and T_{2m}

Changes in daily maximum T_s and T_{2m} have different characteristics over the deforested and afforested areas. Here we explore the possible reasons for the differences. During the daytime, the land surface absorbs shortwave

radiation and warms the air above it mainly through sensible heat transfer, which is determined by the aerodynamic resistance and the temperature gradient between the land surface and air [shown as Eq. (3), in which H is sensible heat flux, ρ is air density, C_p is specific heat of air at constant pressure, T_s is land surface temperature, T_a is air temperature, and r_a is aerodynamic resistance]:

$$H = \rho C_p \frac{T_s - T_a}{r_a}. \quad (3)$$

Figure 4 shows the changes in daytime sensible heat flux, aerodynamic resistance, and temperature gradient over the deforested and afforested areas. Aerodynamic resistance here refers to the resistance to heat transfer. The roughness of the forest canopy gives a low aerodynamic resistance and efficient turbulent transfer away from the surface. For short vegetation like grass, the aerodynamic resistance is relatively high as the surface is smoother. Deforestation increases the aerodynamic resistance, which hinders the efficiency of heat transfer to the air during the day, increasing the temperature gradient between the land and lower atmosphere. Despite the competing effects of the increased aerodynamic

resistance and increased temperature gradient on sensible heat flux [according to Eq. (3)], there is a greater increase in aerodynamic resistance, resulting in decreased sensible heat flux and cooler air. On the contrary, although afforestation cools the surface, the decreased aerodynamic resistance enhances the heat transfer to the atmosphere during the daytime, warming the air. Figure 5 shows that there is a clear linear relationship between changes in the product of sensible heat flux and aerodynamic resistance versus the changes in vertical temperature gradient, further demonstrating that the disagreement between T_s and T_{2m} is mainly attributed to the changes in the aerodynamic resistance (i.e., surface roughness) and the resulting sensible heat flux.

d. Resolution and model dependency

Based on the high-resolution simulations in CESM discussed above, we have found opposite temperature responses to deforestation in the land surface and near-surface air. However, the results may be influenced by the model resolution and the choice of model. We first check the resolution dependency of the LULCC signals in CESM, which might be amplified by the extreme LULCC prescribed in the high-resolution experiments. Figure 6 shows the changes in daily maximum T_s and T_{2m} in the low-resolution simulations compared with the aggregated high-resolution simulations. Compared with the high-resolution deforestation experiments (shown as Fig. 2), the average temperature changes across the deforested area (with at least 15% of the grid area with tree loss) in the low-resolution experiment is relatively small: about +0.6 K warming at the land surface and -0.4 K cooling at the near-surface atmosphere. However, being consistent with area average vegetation changes of the high-resolution experiment, the low-resolution experiment also reveals increased T_s and decreased T_{2m} after deforestation. The magnitudes of the temperature changes are slightly different, which may be associated with the subgrid configuration of land-cover data, the passing of land-cover area-weighted mean surface fluxes to the atmosphere in each grid cell of the low-resolution runs, and model biases related to the spatial resolution (not shown). However, the overall agreement between the aggregated high-resolution experiments and low-resolution experiments further demonstrates the discrepancy between T_s and T_{2m} , and also indicates that the findings from the high-resolution experiment are comparable to the low-resolution experiment in CESM, and the CMIP5 simulations. Resolution changes within the range of $\sim 0.5^\circ$ to 2° do not seem to be a factor in temperature responses to LULCC.

Figure 7 shows the changes in temperature and sensible heat flux among five CMIP5 models. Even though

FIG. 5. Relationship between the changes in the product of sensible heat flux (H) multiplied by aerodynamic resistance (r_a) and the changes in temperature gradient ($T_s - T_a$). Each dot represents a deforested grid cell in the high-resolution experiment.

there is a harmonized set of land-use scenarios used in CMIP5, changes of specific land-cover types can still be very different due to different interpretations of the land-use scenarios among land surface models (Brovkin et al. 2013; Lejeune et al. 2017). Therefore, deforested areas in Fig. 7 are defined based on the changes in LAI, albedo and aerodynamic resistance. We assume that there is tree loss within a grid cell if the LAI decrease is smaller than $-0.1 \text{ m}^2 \text{ m}^{-2}$, the albedo increase is greater than 0.1%, and the change of aerodynamic resistance is greater than zero. All the changes are calculated as the difference between the control and LU experiments shown in Table 1. The five CMIP5 models consistently exhibit a warmer land surface (increased T_s) in the deforested areas. Two models (GFDL-ESM2M and CanESM2) show a general daytime warming in both the land surface and near-surface air. No significant changes are found in daily maximum T_{2m} in IPSL-CM5A-LR, but there is extensive cooling in daily maximum T_{2m} in CCSM4 and CESM-CAM5.

It should be noted that the CMIP5 simulations in CCSM4 or CESM-CAM5 are different in two ways from the LULCC experiments in this study, so their temperature changes might be different if compared with Fig. 6. First, the CMIP5 simulations use historical transient LULCC, while our simulations have fixed land-cover conditions prescribed at the preindustrial times and present, respectively. Second, the CMIP5 simulations are fully coupled with an ocean model and

T_s (aggregated high-res)

T_s (low-res)

T_{2m} (aggregated high-res)

T_{2m} (low-res)

FIG. 6. Changes in T_s and T_{2m} over the deforested areas based on aggregated high-resolution (from $0.47^\circ \times 0.63^\circ$ to $1.9^\circ \times 2.5^\circ$) and low-resolution simulations. The deforested areas here are defined as the grid cell in the low-resolution simulations have at least 15% of the area with tree loss, which accounts for at least 3 high-resolution grid cells having complete tree loss within the low-resolution grid cell. Global maps of the temperature changes are shown in Fig. S6.

FIG. 7. Changes in (left) T_s , (middle) T_{2m} , and (right) sensible heat flux due to the historical deforestation calculated based on five CMIP5 models. Only GFDL-ESM2M and IPSL-CM5A-LR have subdaily output available, which allow for the calculation of the daily maximum T_s and corresponding daytime T_{2m} and H . For CanESM2, CCSM4, and CESM1-CAM5, T_s and H are the changes in their daily mean, and T_{max} is daily maximum 2-m air temperature.

have carbon–nitrogen biogeochemistry included in CLM, while our simulations only have the land and atmosphere components activated and prescribed satellite-based vegetation physiology. Despite these differences, opposite changes in T_s and T_{2m} are still evident in CCSM4 or CESM-CAM5.

To understand the responses of T_s and T_{2m} , we also examine the changes in sensible heat flux in the CMIP5 experiments. The GFDL-ESM2M model shows increased daytime sensible heat flux, indicating that more heat can be transferred from the land surface to the air, leading to warmer air over the deforested areas. The IPSL-CM5A-LR model exhibits decreased daytime sensible heat flux in most of the regions, suggesting that the heat transfer from the land surface to the air is less efficient over the deforested areas and cannot effectively sustain warming air. For CanESM2, CCSM4, and CESM-CAM5, because there is no subdaily output available, the identified changes in sensible heat flux may not fully represent the summer daytime situation. However, we see consistent decreased sensible heat flux

in most regions of North America and Europe in CCSM4, CESM-CAM5, and CanESM2.

4. Discussion and conclusions

In this study, we find that the deforestation-induced daily maximum temperature changes can be different between the land surface and near-surface air in the climate models. These discrepancies between T_s and T_{2m} suggest that the relationship between the deforestation and hot extremes can be swayed by the choice of temperature variables. For most global observation-based studies, T_s has been favored because the daytime and nighttime T_s can be obtained through the satellite retrievals at the global scale (Li et al. 2015; Bright et al. 2017; Schultz et al. 2017; Chen and Dirmeyer 2019; Alkama and Cescatti 2016; Forzieri et al. 2017), and they all suggest a daytime warming at the land surface after deforestation during summer. There are also a few studies using paired flux-tower sites to explore deforestation-related T_{2m} changes (Lee et al. 2011; Zhang et al. 2014;

Vanden Broucke et al. 2015). Unlike T_s , the response of T_{2m} is more regionally varying (Snyder et al. 2004). Tropical deforestation can cause a significant daytime warming in T_{2m} (Lee et al. 2011; Zhang et al. 2014), but no significant difference is found between the open land and temperate (or boreal) forest (Lee et al. 2011; Zhang et al. 2014; Vanden Broucke et al. 2015). Climate models provide both temperature variables, although T_{2m} is a derived quantity and there are variations among models in the calculation of both variables. In other words, the definition of T_{2m} among the models is still unclear. Therefore, care should be taken to ensure consistency when comparing to observational studies. When examining the possible impacts of deforestation on temperature, the use of temperature variables should be carefully considered and specified precisely.

The CESM deforestation experiments and CMIP5 land-use simulations consistently show increased daily maximum T_s after deforestation, which agrees with the observed warmer surface in the open land than forest. However, discrepancies exist in daily maximum T_{2m} : warmer in GFDL-ESM2M and CanESM2, cooler in CCSM4 and CESM1-CAM5, and no evident change in IPSL-CM5A-LR. We acknowledge that CCSM4 and CESM1-CAM5 in CMIP5 and the model used in our deforestation experiments are variations of the same model (but with different versions and configurations). However, our results highlight the similar issue that the response of T_s and T_{2m} to deforestation can be very different in the sign or magnitude within the CMIP5 models. Lejeune et al. (2018) also found the discrepancies in the reconstructed T_{2m} change among the CMIP5 models, even though signs of temperature changes for some models are different than our study (warmer in IPSL-CM5A-LR and cooler in GFDL-ESM2M in their estimation). Furthermore, the responses of hot extremes to deforestation are diverse: cooler in HadGEM2-ES (Christidis et al. 2013), CESM (Chen and Dirmeyer 2019), GFDL-ESM2M, and GFDL-ESM2G (Li et al. 2018), warmer in CanESM2 (Li et al. 2018), and warmer and drier in GFDL-ESM2G (Findell et al. 2017). Even though the definitions of the hot extremes may be different among these studies, all the hot extreme indices are calculated based on and closely related to the summertime daily maximum temperature used in this study. Therefore, our findings have implications for future hot extreme assessment.

The discrepancies in T_{2m} changes among the climate models reveal two issues. First, we find that the changes in T_{2m} are closely related to the surface roughness and sensible heat fluxes. This indicates the uncertainties among the climate models about roughness formulation and the partitioning of available energy between the

latent and sensible heat fluxes, as reported in previously studies (de Noblet-Ducoudré et al. 2012; Lejeune et al. 2017; Li et al. 2018), especially regarding the parameterization of sensible heat flux (Best et al. 2015). In this study, four out of five CMIP5 models show decreased sensible heat flux, which agrees with the observations in previous studies (Vanden Broucke et al. 2015; Chen and Dirmeyer 2019). Only GFDL-ESM2M shows increased sensible heat flux over the deforested areas, implying that the increased daily maximum T_{2m} in GFDL-ESM2M should be interpreted with caution.

Second, current observations of deforestation-related T_{2m} have very poor spatial coverage and are temporally limited (Lejeune et al. 2017). Meanwhile, the observational studies usually estimate the deforestation-related signals based on the difference between the adjacent forest and open land units (also known as the space-for-time analogy; Lee et al. 2011; Li et al. 2015; Chen and Dirmeyer 2016; Bright et al. 2017; Schultz et al. 2017). However, the space-for-time analogy cannot be directly compared with the coupled LULCC experiments, because the assumption of identical atmospheric background states between the open land and forest cannot be satisfied in model sensitivity studies. Especially for T_{2m} , it is dominantly influenced by the large-scale atmospheric background state (Chen and Dirmeyer 2016). Therefore, it is relatively difficult to verify the simulated T_{2m} responses with the observations. The variable T_s can be more consistently applied for use in future comparisons involving observations and climate models. However, because T_{2m} is more commonly used for climate assessment and more related to human activities, future deforestation/climate studies should also pay attention to the uncertainties in the relationship between T_s and T_{2m} . Inconsistencies in the definition of T_{2m} discussed in the introduction can also contribute to uncertainties.

For the nighttime temperature, CESM shows warming effects from deforestation in both the land surface and near-surface air. However, observation-based studies agree on a significant nighttime cooling over the deforested areas (Lee et al. 2011; Zhang et al. 2014; Vanden Broucke et al. 2015; Schultz et al. 2017), which is mainly attributed to the decreased turbulence after tree loss, and less heat aloft being brought to the surface (Lee et al. 2011). The CESM deforestation experiments can capture the cooling effects that are derived from the decreased nighttime sensible heat flux. However, the cooling effects are compensated and exceeded by the warming effects of nighttime heat flux upward through the soil, resulting in overall warming at the surface. Previous studies have suggested that the changes in ground heat flux are greatly overestimated

in climate models (Vanden Broucke et al. 2015; Chen and Dirmeyer 2019), mistakenly resulting in a surface warming. Among the CMIP5 models, only CanESM2 exhibits a nighttime cooling (not shown). Because ground heat flux is calculated as the residual of the surface energy budget (Oleson et al. 2010), this bias may be an aggregation of model uncertainties in the parameterization of sensible heat flux and the nocturnal boundary layer.

Last, considering the asymmetric temperature responses to deforestation during the daytime and nighttime, more work should be focused on understanding the diurnal cycle of deforestation-induced climate changes. This has implications for other diurnally driven processes, such as atmospheric boundary layer growth and the triggering of convection. Besides the availability of more spatially extensive observational datasets of surface energy budget quantities, it is also important to have subdaily output in future modeling studies for model evaluation and associated LULCC experiments.

Acknowledgments. This study was supported by the National Science Foundation (AGS-1419445). We would like to acknowledge high-performance computing support from Cheyenne (doi:10.5065/D6RX99HX) provided by NCAR's Computational and Information Systems Laboratory, sponsored by the National Science Foundation. The Community Earth System Model is freely available at <http://www.cesm.ucar.edu/models/cesm1.2/>. We also acknowledge the World Climate Research Programme's Working Group on Coupled Modelling, which is responsible for CMIP, and we thank the climate modelling groups for producing and making available their model output. For CMIP the U.S. Department of Energy's Program for Climate Model Diagnosis and Intercomparison provides coordinating support and led development of software infrastructure in partnership with the Global Organization for Earth System Science Portals. We also thank the reviewers for their constructive and thoughtful comments, which helped us to improve this manuscript.

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