Sensitivities of Land Cover–Precipitation Feedback to Convective Triggering

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ABSTRACT

The land surface state can be an important factor in the triggering of precipitation, whose depiction in Earth system models (ESMs) crucially relies on the representation of convective initiation. However, the sensitivity of land-cover change–precipitation feedbacks to different parameterized triggering criteria in ESMs has not been examined. In this study, a new triggering mechanism based on the heated condensation framework (HCF) is implemented in the Community Earth System Model (CESM). A set of land-cover change experiments with different convective triggering conditions are performed to evaluate the influence of convective triggering on land–atmosphere coupling strength and the response of summer afternoon precipitation to land-cover change over North America. Compared with the default parameterization, which depends on a CAPE threshold, the HCF trigger shows an improvement in the diurnal timing of summer precipitation but larger dry biases over much of the study area. With the HCF trigger, CESM exhibits weakened coupling strength between soil moisture and surface turbulent fluxes over the Great Plains. The surface temperature deficit, as an additional triggering criterion in HCF, is not significantly coupled with surface fluxes over the central Great Plains despite strong latent heat–CAPE coupling. In contrast to the CAPE-trigger simulations, which indicate increased precipitation over the Great Plains after agricultural expansion, the HCF-trigger simulations show significantly increased afternoon precipitation only over the northern plains, which is mainly associated with more frequent deep convection. The discrepancies suggest caveats when investigating the impacts of land-cover change on precipitation, because the magnitude and spatial patterns of precipitation change can be greatly affected by the treatment of convection in ESMs.

1. Introduction

The interaction between the land surface and atmosphere is an important element of the climate system. Land-use/land-cover changes (LULCCs) can modify the surface energy and water fluxes and can therefore affect climate at regional and broader scales. Within the land–atmosphere coupling “process chain” (Santanello et al. 2011), the land surface affects precipitation by influencing planetary boundary layer (PBL) development thereby triggering and maintaining convection (Carleton et al.

2001; Courault et al. 2007; Adegoke et al. 2007; Dirmeyer et al. 2014; Ek and Holtslag 2004; Findell and Eltahir 2003; Margulis and Entekhabi 2001; Pielke 2001; Santanello et al. 2013; Tawfik and Dirmeyer 2014). Many model-based studies have demonstrated the impacts of land-cover change on precipitation (e.g., Bagley et al. 2014; Chen and Dirmeyer 2017; Douglas et al. 2009; Lawrence and Chase 2010; Paeth et al. 2009). However, climate models largely manifest land cover–precipitation feedbacks via the convective parameterization, in which convective triggering, a threshold to activate the convective parameterization, is a critical but often arbitrary factor in the correct simulation of convection.

Because convection cannot be properly resolved at typical Earth system model (ESM) scales of 50–100-km-wide grids, convection must be parameterized to stabilize the large-scale environment. Recently, there have

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been several studies outlining how sensitive parameterized convection is to the convective triggering mechanism. Convective triggering in models is usually defined as a condition or set of conditions that need to be achieved in order for the remaining parameterization to be activated. [Suhas and Zhang \(2014\)](#) performed a broad assessment of some of the more common triggering mechanisms, applying them to the U.S. Department of Energy's Atmosphere Radiation Measurement (ARM) field data. They found that the triggering mechanisms with the best skill were those that included a large-scale forcing component in addition to the contribution from local instability generation. This was similar to prior work that modified convective closure and the convective trigger in the Community Climate System Model, version 3 ([Zhang and Mu 2005](#); [Wang et al. 2015](#)).

To develop a more physically based convective triggering mechanism with contribution from large-scale and local land surface forcings, [Tawfik and Dirmeyer \(2014\)](#) introduced the heated condensation framework (HCF), which quantifies how primed the atmosphere is to initiation of moist free convection. Instead of lifting a hypothetical unmixed parcel through a fixed atmospheric profile, the HCF constructs a hypothetical mixed boundary layer by incrementally adding heat at the surface. [Bombardi et al. \(2015, 2016\)](#) implemented the HCF into the NCEP CFSv2 and found that the HCF trigger improves the representation of the seasonal precipitation cycle over the Indian subcontinent ([Bombardi et al. 2015](#)) and better captures the frequency of convection ([Bombardi et al. 2016](#)). The HCF trigger was added in the Community Earth System Model (CESM) as an additional criterion for deep convective initiation ([Tawfik et al. 2017](#)), and it was shown to delay diurnal convective onset time by several hours and to alleviate the convective overactivity issue in CESM. With such improvements in the representation of convective precipitation, it is worthwhile to examine how land cover–precipitation feedback responds to the new convective triggering mechanism, especially as it takes account of both the large-scale atmosphere background and local land surface forcing.

Furthermore, uncertainties remain in our understanding of soil moisture–precipitation feedback. Strong positive relationships between soil moisture and precipitation in model-based studies (e.g., [Koster et al. 2004](#); [Guo et al. 2006](#)) are not well supported by recent observation-based studies (e.g., [Wei et al. 2010](#); [Phillips and Klein 2014](#); [Tuttle and Salvucci 2016](#); [Taylor et al. 2012](#)). As an end of the land–atmosphere coupling process chain, convective initiation and its parameterization in ESMs can play an important role in determining the sign and strength of soil moisture–precipitation feedback.

Therefore, it is also worthwhile to investigate the changes in land–atmosphere coupling strength in response to the implementation of the HCF trigger. This is particularly relevant to studies of the impact of land-use change, in which land–atmosphere coupling is the mechanism by which climate is affected.

In this paper, the sensitivity of land cover–precipitation feedback is investigated through a set of land-cover change experiments with different triggering conditions for deep convection in CESM. As a follow-up to [Chen and Dirmeyer \(2017\)](#), this study is mainly focused on the relationship between morning land surface conditions and afternoon precipitation over North America in summer, during which the sensitivity of convection to the land surface is expected to be the greatest ([Dirmeyer et al. 2013](#); [Taylor et al. 2012](#)). [Chen and Dirmeyer \(2017\)](#) demonstrated a strong positive land–precipitation relationship over the Great Plains in CESM. We further explore the evolution of this relationship over the Great Plains with a more physically based convective triggering in CESM and investigate how the convective triggering scheme influences precipitation responses to land-use and land-cover change. [Section 2](#) describes the experimental design and metrics used to investigate land–atmosphere coupling strength. [Section 3](#) presents results from both of the convective triggering schemes. [Section 4](#) includes discussion and conclusions.

2. Methodology

a. Model description

We use CESM, version 1.2.2, in this study. CESM is a coupled Earth system model composed of separate climate system components for atmosphere, ocean, land, sea ice, and land ice ([CESM Software Engineering Group 2013](#); [Hurrell et al. 2013](#)). Because of the focus on land–atmosphere interactions from the physical climate perspective, only the Community Atmosphere Model, version 4.0 (CAM4.0; [Neale et al. 2010](#)), and Community Land Model, version 4.5 (CLM4.5; [Oleson et al. 2013](#)), are activated in this study. In CLM, satellite-based vegetation phenology is used and carbon–nitrogen cycling is not included. The component set is defined as 2000_CAM4_CLM45%SP_CICE%PRES_DOCN%DOM_RTM_SGLC_SWAV. For all experiments, identical prescribed SST and sea ice cover climatologies with a fixed CO₂ concentration of 367.0 ppm are used. The prescribed SSTs are a monthly mean climatology averaged over 1982–2001 ([Hurrell et al. 2008](#)), which is derived from a merged product based on the monthly

TABLE 1. Land-cover change experiments using CESM.

Name	Convective triggering	Land cover
CTRL1850	$\text{CAPE} > 70 \text{ J kg}^{-1}$	PFTs in 1850
CTRL2000	$\text{CAPE} > 70 \text{ J kg}^{-1}$	PFTs in 2000
HCF1850	$\text{CAPE} > 70 \text{ J kg}^{-1}; \theta_{\text{def}} = 0 \text{ K}$	PFTs in 1850
HCF2000	$\text{CAPE} > 70 \text{ J kg}^{-1}; \theta_{\text{def}} = 0 \text{ K}$	PFTs in 2000

mean Hadley Centre Sea Ice and Sea Surface Temperature dataset, version 1 (HadISST1), and the weekly NOAA Optimum Interpolation (OI) Sea Surface Temperature, version 2.

b. Experimental design

The land-cover change and convective triggering sensitivity experiments are listed in Table 1. Two land-cover scenarios are used, corresponding to the pre-industrial (1850) and present conditions (2000). Details of the datasets for land cover in 1850 and 2000 can be found in Lawrence et al. (2012). Figure 1 shows the changes to major plant functional types (PFTs; trees, grass, and crops) from 1850 to 2000. Table 2 shows the changes in JJA surface properties over the two subregions with significant precipitation changes (described later). Both of the regions show decreased leaf area index (LAI) and increased surface albedo due to the deforestation. Over the northern plains, most of the land-cover changes have occurred over grassland rather than forests, so the changes in LAI and albedo are not as strong as in the lower Mississippi River basin, where the forests have been replaced with grassland and crop fields. The aerodynamic resistance has increased over the lower Mississippi River basin after deforestation (with decreased surface roughness), while it is slightly decreased over the northern plains.

Also, two convective triggering criteria are used. In CAM4.0, deep convection is parameterized using the Zhang–McFarlane (ZM) scheme (Zhang and

McFarlane 1995), with modifications to include convective momentum transports (Richter and Rasch 2008) and a dilute plume calculation of convective available potential energy (CAPE; Neale et al. 2013). By default in ZM, deep convection is triggered when CAPE exceeds 70 J kg^{-1} . Tawfik et al. (2017) introduced a process-based diagnostic based on HCF as an additional requirement for triggering in CAM. Within HCF, the buoyant mixing potential temperature θ_{BM} is calculated from the vertical profiles of temperature and humidity. It quantifies how well conditioned the atmosphere is to initiating moist convection. The surface temperature deficit θ_{def} is the difference between θ_{BM} and near-surface potential temperature $\theta_{2\text{m}}$, indicating how much $\theta_{2\text{m}}$ needs to increase by surface heating to reach a profile that becomes saturated at the top of a mixed layer (when $\theta_{\text{def}} = 0 \text{ K}$), which evolves through the day potentially triggering convection. Previous studies have demonstrated that HCF can well capture convective initiation over the central plains as well as initiation caused by land–sea breeze circulations (Tawfik and Dirmeyer 2014; Tawfik et al. 2015a,b). Therefore, the second convective trigger includes both the CAPE and θ_{def} thresholds. In other words, deep convection is triggered only when both triggering conditions ($\text{CAPE} > 70 \text{ J kg}^{-1}$ and $\theta_{\text{def}} = 0 \text{ K}$) are met. Details of the HCF implementation can be found in Tawfik et al. (2017). However, unlike in Tawfik et al. (2017), this study simply implements the HCF trigger into CAM without considering subgrid land surface heterogeneity information, meaning that gridcell average $\theta_{2\text{m}}$ is used to calculate θ_{def} . Hereafter, control (CTRL) experiments indicate simulations with the default convective triggering, and HCF experiments use the trigger requiring both CAPE and θ_{def} thresholds to be satisfied.

All simulations are run at a horizontal resolution of $0.9^\circ \times 1.25^\circ$ with hourly output. To maintain consistency with Chen and Dirmeyer (2017), 45-yr simulations

FIG. 1. Changes of each grid (%) in three major PFTs: (a) tree, (b) grass, and (c) crop, from 1850 to 2000. Dashed boxes are used to identify two subregions for diurnal cycle analysis.

TABLE 2. Changes in JJA LAI, surface albedo, and aerodynamic resistance over the two subregions with significant precipitation changes.

Subregions	LAI	Albedo	Aerodynamic resistance (s m^{-1})
Northern Great Plains	-0.04	0.12	-9.51
Lower Mississippi River basin	-0.35	0.24	18.16

(discarding the first 5 years for spinup) are conducted, with the default initial condition provided by the model. An extra 40 years of simulation has been carried out to test the robustness of the signals. Results from the extra 40-yr simulations exhibit consistent patterns in precipitation changes over the regions of concern compared with the first 40-yr simulations (see Fig. S1 in the supplemental material). To be comparable with the CTRL experiments from [Chen and Dirmeyer \(2017\)](#), we concentrate on the first 40 years of simulation. Furthermore, because the sensitivity of convection to land surface states over the United States is expected to be the greatest during afternoons in the summer, only boreal summer (JJA) is considered in this study, and local afternoon (1300–1800) and local morning (0900–1200) states and fluxes are considered for each grid cell with averages taken over those periods of the day. Statistical significance for the differences is evaluated using the Student's t test on yearly averaged morning/afternoon values (sample size of 40).

c. Precipitation observations

Two observational datasets with high spatial and temporal resolution are used to validate the CESM simulations. The Climate Prediction Center morphing technique (CMORPH) dataset provides 3-hourly precipitation estimates with 0.25° spatial resolution that are derived from passive microwave satellites aboard low-Earth-orbiting space craft and geostationary satellite infrared data from December 2002 to present ([Joyce et al. 2004](#)). Because of the possible biases in CMORPH suggested by previous studies ([Janowiak et al. 2007](#); [Zeweldi and Gebremichael 2009](#)), Multi-Source Weighted-Ensemble Precipitation (MSWEP) is also used. The MSWEP dataset merges the highest-quality precipitation sources available (including two gauge-based observational analyses, three satellite products, and two reanalysis datasets) and provides gridded precipitation “observations” over land for the period 1979–2015 with a 3-hourly temporal and 0.25° spatial resolution ([Beck et al. 2017](#)).

To be comparable with the climatological CESM simulations, both CMORPH and MSWEP are regridded to

$0.9^\circ \times 1.25^\circ$ resolution with bilinear interpolation. Note that the time span of CMORPH does not cover the earlier period (1982–2001) of the monthly mean climatology of prescribed SSTs used in the CESM simulations. The available precipitation estimates for the period (2003–15) are used for model validation.

d. Land–atmosphere coupling metrics

A two-legged coupling metric ([Dirmeyer 2011](#); [Guo et al. 2006](#); [Wei and Dirmeyer 2012](#)) is used to assess land–atmosphere interactions. As a follow-up to [Chen and Dirmeyer \(2017\)](#), this study is focused on the degree of influence of morning land conditions on afternoon atmospheric conditions. For the terrestrial leg, the coupling is focused on the relationship between morning soil moisture and morning surface fluxes. For instance,

$$CS_L = \text{corr}(\text{SM}, \text{LH})\sigma_{\text{LH}} \quad (1)$$

is the coupling index between morning top 10-cm soil moisture (SM) and morning surface latent heat flux (LH), where corr indicates the temporal Pearson's product-moment correlation coefficient and σ is the temporal (day to day) standard deviation.

For the atmospheric leg, the coupling is focused on the relationship between morning surface fluxes and subsequent afternoon atmospheric variables (such as CAPE and θ_{def}). For instance,

$$CS_A = \text{corr}(\text{LH}, \text{CAPE})\sigma_{\text{CAPE}} \quad (2)$$

is the coupling index between the morning LH and afternoon CAPE.

Considering the focus of this study is on convective triggering, it is also worthwhile to examine the coupling index between morning surface fluxes and the probability of the afternoon convective triggering criterion being satisfied ($\text{CAPE} > 70 \text{ J kg}^{-1}$ and $\theta_{\text{def}} = 0 \text{ K}$):

$$CS_A = \text{corr}(\text{LH}_{\text{bin}}, \Gamma_{\text{CAPE}})\sigma_{\Gamma_{\text{CAPE}}}, \quad (3)$$

in which the distribution of morning LH is split into 10 bins with an equal number of days in each bin (LH_{bin}); Γ_{CAPE} is the probability of afternoon CAPE satisfying the criterion above for each LH bin.

3. Results

a. Afternoon precipitation simulations with different triggers

Figure 2 shows the afternoon precipitation from both observations and CESM simulations over North America during summer. Generally, discrepancies exist

FIG. 2. Average afternoon precipitation (mm) in observations (a) CMORPH and (b) MSWEP and in CSM simulations (c) CTRL2000 and (d) HCF2000 during JJA.

both between observational datasets and between observations and simulations. There is greater afternoon rainfall over the southeastern United States and northern Mexico from both of the observations. However, CMORPH exhibits an erroneous rainfall maximum over the eastern side of the Rocky Mountains and the adjacent plains, which is not present in MSWEP and CESM. Lee et al. (2007) also found greater rainfall amplitudes over the central United States in CMORPH, which can be attributed to a positive bias of the infrared satellite estimates over land (Janowiak et al. 2007), especially over dry but convectively active regions where there is significant evaporation of the rainfall below the cloud base (McCollum et al. 2002).

Compared to CMORPH, CESM underestimates the afternoon precipitation over a large area of North America, especially over the southeastern and central United States and northern Mexico (Figs. 3a,b). With the HCF trigger, the underestimation bias is larger over the central plains and western foothills of the Sierra Madre Occidental in Mexico, while the slight increase in afternoon precipitation over Florida compared with the default triggering represents a reduced bias (Fig. 3f).

The comparison between MSWEP and CESM also indicates an underestimation of afternoon precipitation in the model (Figs. 3c,d), but a better agreement over the central United States than CMORPH, which can be taken as encouraging. Additionally, we examine the simulated daily total precipitation (not shown), which exhibits similar biases to the afternoon precipitation. The underestimation over the central United States is greater in the HCF simulation, however, which shows an improvement of the overestimation of daily precipitation over the Rocky Mountains and the southeastern United States.

Because the temporal resolutions are different between the observations and model simulations, it is difficult to compare the diurnal cycles of precipitation precisely. The focus of this study is on afternoon precipitation, so we calculate it as a percentage of daily total precipitation during summer (Fig. 4). The two observational products show good agreement in the spatial pattern of afternoon precipitation as a proportion of total daily precipitation. In the summer, the maximum proportion of afternoon rainfall is seen over the southeastern United States, the North American monsoon

FIG. 3. Comparison between observations and CESM simulations in afternoon precipitation (mm) during JJA. (a),(b) The difference between CESM simulations and CMORPH; (c),(d) the difference between CESM simulations and MSWEP; (e) the difference between the two observational analyses; and (f) the difference between the two convective triggering simulations of CESM.

regime, and over the Rocky Mountains (Figs. 4a,b). The smallest contribution of afternoon rainfall to daily total precipitation is found over the Great Plains, where there is a nocturnal maximum in rainfall (Dai et al. 1999; Janowiak et al. 2007), as well as across the West Coast, which might be influenced by the nocturnal or early morning maximum in rainfall over the ocean (Janowiak et al. 2005). CESM with the default triggering (Fig. 4c) does not well capture the observed spatial pattern of the

percentage of afternoon rainfall (the spatial correlation with MSWEP $r = 0.47$, $p < 0.01$), and in general does not show as much spatial structure as seen in the observations. The HCF-trigger simulation exhibits a clear improvement (the spatial correlation with MSWEP $r = 0.60$, $p < 0.01$), especially over the Rockies, the southeastern United States, and the Great Plains, though it also produces an overestimation of afternoon rainfall percentage over the U.S. Gulf Coast (Fig. 4d).

FIG. 4. As in Fig. 2, but for afternoon precipitation (%) as a percentage of daily total precipitation.

Overall, compared to the two observational datasets, CESM simulations with both triggering mechanisms show a drier bias in afternoon precipitation over the United States, but the HCF trigger shows some improvement in the diurnal cycle of summer precipitation.

b. Land–atmosphere coupling strength with different triggers

The terrestrial leg of land–atmosphere coupling between soil moisture and surface fluxes is examined first. The coupling strength in CESM with the default trigger (CTRL2000) has been assessed in [Chen and Dirmeyer \(2017\)](#), so only the HCF-trigger results (HCF2000) and their comparison with the default trigger (HCF2000 – CTRL2000) are presented in Fig. 5. Generally, significant coupling strength between soil moisture and surface fluxes is found over the Great Plains, suggesting soil moisture is a limiting factor for latent fluxes in this region for the HCF2000 simulation. Compared with the default trigger, the HCF trigger decreases the coupling strength between soil moisture and surface fluxes over the Great Plains (Figs. 5b,d). On the other hand, there is strengthened coupling over the southeastern United States, especially

for the coupling strength between soil moisture and latent heat flux.

Decomposition of the coupling index into its standard deviation and correlation components (not shown) suggests that the increase over the southeastern United States is mainly attributable to the increased correlation coefficient between soil moisture and latent flux, while the decreased coupling strength over the Great Plains is mainly attributable to decreased variability (standard deviation) of morning latent heat flux. The changes in the terrestrial leg of coupling strength can be explained by the dry bias in the HCF simulations. Over the Great Plains, which is a moisture-limited regime with strong coupling between soil moisture and latent heat flux, further decreased precipitation may have little impact on the sensitivity of latent heat flux to soil moisture (the correlation), but decreases latent flux variability substantially. Over the southeastern United States, especially on the western side of the Appalachian Mountains, moisture is not a strong limit for surface latent heat flux in the CTRL case but is on the margin of the transition region to significant coupling strength to the west. The decreased precipitation over this region increases the sensitivity of

FIG. 5. The terrestrial leg of land-atmosphere coupling index (W m^{-2}) (a),(c) between soil moisture and surface fluxes in HCF2000 and (b),(d) its difference with CTRL2000. Hatching indicates significance at the 95% confidence level. Meridional bands are an artifact of the hourly model output.

latent heat flux to soil moisture, increasing the correlation component and thus the coupling strength.

Chen and Dirmeyer (2017) showed that there was strong coupling in CESM between morning surface fluxes and afternoon CAPE over the Great Plains, which could explain the large afternoon rainfall response to land-cover change over this region. Here, the relationship between latent heat flux and CAPE is examined in the HCF-trigger simulation (Fig. 6). Compared with the default trigger, the HCF trigger also exhibits significant and stronger coupling between latent heat flux and CAPE over the western United States as well as over Canada and New England. The increased coupling strength is mainly attributed to greater variability in CAPE, especially over the eastern United States (not shown). We also find diminished coupling strength over the Gulf Coast, which is mainly attributed to the decreased correlation coefficient between latent heat flux and CAPE over this region (not shown).

With the HCF trigger, CAPE is not the only property that drives convective triggering. The potential temperature deficit also determines convective initiation. The relationship between surface latent heat flux and θ_{def} is shown in Fig. 7a. Latent heat flux is not significantly coupled with θ_{def} over the central plains, and there is a negative coupling index between latent heat flux and θ_{def} over the western United States and northern Mexico.

To explain this lack of coupling strength, we examined the individual contributors to the θ_{def} term in more detail. Recall that within the HCF formulation, θ_{def} is the difference between buoyant mixing potential temperature and near-surface potential temperature. Therefore, the sensitivity of θ_{def} to latent heat flux can be decomposed as

$$\frac{d(\theta_{\text{def}})}{d(\text{LH})} = \frac{d(\theta_{\text{BM}} - \theta_{2\text{m}})}{d(\text{LH})} = \frac{d(\theta_{\text{BM}})}{d(\text{LH})} - \frac{d(\theta_{2\text{m}})}{d(\text{LH})}. \quad (4)$$

The difference between the sensitivities of θ_{BM} and $\theta_{2\text{m}}$ to surface latent heat flux can be considered as the

FIG. 6. The atmospheric leg of land–atmosphere coupling index (J kg^{-1}) (a) between latent heat flux and CAPE and (b) its difference with CTRL2000. Hatching indicates significance at the 95% confidence level.

sensitivity of θ_{def} to the latent heat flux. Figure 7a shows the coupling index between surface latent heat flux and θ_{def} , which is mathematically equivalent to $\sigma(\text{LH})[d(\theta_{\text{def}})/d(\text{LH})]$, and shows good agreement with the difference between LH– θ_{BM} and LH– $\theta_{2\text{m}}$ coupling indices (Fig. 7b), indicating the robustness of our decomposition approach.

Over the western United States, northern Mexico, and the central plains, which are moisture-limited regions, there are significant negative indices between the latent heat flux and θ_{BM} (Fig. 7c), indicating that an increase in latent heat flux could moisten the atmospheric column and thus lower the buoyant condensation level (BCL) and decrease θ_{BM} , which makes intuitive sense. On the other hand, the negative relationship could also arise from a lower θ_{BM} favoring more precipitation, thereby increasing latent heat flux in moisture-limited regimes. Either one of these processes is plausible, and it should be noted that the correlation-based coupling metrics cannot discriminate between cause and effect, but rather indicate levels and directions of connectedness. Over parts of the eastern United States and Canada, there are positive coupling indices between the latent heat flux and θ_{BM} , which can be attributed to the effects of large-scale forcings in this largely atmosphere-controlled land–atmosphere coupling region.

For the second term of the decomposition, the sensitivity of $\theta_{2\text{m}}$ shows a very similar spatial pattern to θ_{BM} (Fig. 7d). However, over the central plains, the response of $\theta_{2\text{m}}$ to surface latent flux is greater than the response of θ_{BM} . In other words, with the subsequent input of latent flux, the preference of the decreased θ_{BM} to lower θ_{def} and convective initiation is undermined by a larger decrease in $\theta_{2\text{m}}$, which can even lead to an increase in

θ_{def} (Fig. 7b). This can explain the absence of coupling between latent heat flux and θ_{def} (Fig. 7a).

In a previous study, we showed that impacts of land-cover change on afternoon precipitation manifest through changes in rainfall frequency rather than intensity (e.g., Chen and Dirmeyer 2017). Therefore, the frequency of the convective triggering criteria being satisfied is also investigated in the land–atmosphere coupling framework. Figure 8 shows the relationship between morning latent heat flux and the probability of afternoon CAPE or θ_{def} being satisfied to trigger convection [calculated based on Eq. (3)] in the CTRL and HCF simulations. The relationship between latent heat flux and frequency of the CAPE criterion being satisfied shows very similar spatial patterns between the two simulations. There is strong positive coupling over the Great Plains and the western United States, but the strength is relatively weak in HCF.

For the coupling index between latent heat flux and the probability of the θ_{def} criterion being satisfied (Fig. 8c), no connection is found over the central plains, which agrees with the relationship between latent heat flux and the magnitude of θ_{def} in Fig. 7. However, there is positive relationship over the northern plains, the western United States, the Gulf Coast, and parts of the southeast (the lower Mississippi River basin), implying that anomalous moisture input from the land surface may be associated with more deep convective activity being triggered. Because the HCF triggering requires both the CAPE and θ_{def} criteria to be satisfied at the same time, Fig. 8d shows their combined sensitivity to morning latent heat flux, which exhibits the same pattern as the coupling index for the probability of the θ_{def} criterion alone being satisfied, indicating that

FIG. 7. (a) The coupling index (K) between latent heat flux and θ_{def} and the decomposed coupling index (K) within the moisture pathway of the HCF based on Eq. (1), including (b) the difference between (c) the LH- θ_{BM} coupling and (d) LH- $\theta_{2\text{m}}$ coupling index. Note that the LH- $\theta_{2\text{m}}$ coupling index is multiplied by -1 in (d) so that (b) equals the sum of (c) and (d). Hatching indicates significance at the 95% confidence level.

the convective triggering is more regulated by the LH- $\Gamma_{\theta_{\text{def}}}$ relationship.

c. Response to land-cover change with HCF triggering

Figure 9a shows the change in total afternoon precipitation due to land-cover change from 1850 to 2000 in HCF simulations. Results with the extra 40 years of simulation (Fig. S1) exhibit consistent patterns in precipitation changes, indicating the robustness of the signals. Also consistent with Chen and Dirmeyer (2017), the changes in total precipitation are mostly attributed to changes to the convective component (not shown). Afternoon precipitation significantly increases over the northern Great Plains, where forests and grasslands have been replaced with crops (see Fig. 1). Over parts of the southeastern United States (the lower Mississippi River basin), there is significantly decreased afternoon precipitation when forests are replaced with grass or

crops. Figure 9b shows the difference of precipitation change between the default and HCF triggering. Generally, both simulations exhibit similar precipitation change over the northern Great Plains. However, the HCF triggering does not show a significant increase in afternoon precipitation over the central plains. The decreased afternoon precipitation is not found in the default triggering over the southeast.

To explore the mechanisms behind the disagreement in precipitation responses, changes in morning surface latent and sensible heat fluxes are examined (Fig. 10). Significant increases in latent heat flux associated with land-cover change are only found over the northern Great Plains (Fig. 10a). This is very different than in the CTRL simulations (Chen and Dirmeyer 2017), in which there is extensively increased latent heat flux over the entire Great Plains. In addition, the spatial pattern of sensible heat flux change shows good agreement with the land-cover change pattern (Fig. 10b). Deforestation

FIG. 8. The coupling index between latent heat flux and the probability of the CAPE criterion being satisfied in (a) CTRL2000 and (b) HCF2000, (c) the coupling index between latent heat flux and the probability of the θ_{def} criterion being satisfied in HCF2000, and (d) the coupling index between latent heat flux and the probability of both θ_{def} and CAPE criteria being satisfied in HCF2000. Hatching indicates significance at the 95% confidence level.

leads to decreased sensible heat flux, while reforestation over the eastern United States leads to increased sensible heat flux. The change in sensible heat flux is mainly influenced by changes in surface roughness directly resulting from deforestation and reforestation. Roughness changes appear to have much less effect on latent heat flux here, demonstrated as well by the weak coupling index in Fig. 5. Latent heat flux in CLM is modulated by many more factors than sensible heat flux, including multiple biophysical and biochemical properties of the vegetation that help maintain stasis, particularly in humid regions. Over the Great Plains, the smaller change in latent heat flux in the HCF simulations than the CTRL simulations is associated with the limited land–atmosphere coupling strength, which also weakens the impacts on sensible heat flux. However, the changes in latent and sensible fluxes should be interpreted with caution because of some potential issues

with evapotranspiration estimation in CLM that might be associated with land-use changes, such as excessive soil evaporation when the canopy is sparse and low contribution of transpiration to the total evapotranspiration (cf. Swenson and Lawrence 2014).

Figure 11 shows the land-cover change induced impact on CAPE, which exhibits very similar spatial patterns between the two sets of simulations. Land-cover changes lead to increased afternoon CAPE over the Great Plains and the western United States, but decreased CAPE over the eastern United States. Compared with the default trigger, the HCF simulations exhibit greater changes in CAPE over the northern plains and the eastern United States.

Figure 12 shows the change in the number of days with afternoon precipitation greater than 1 mm, days with afternoon CAPE greater than 70 J kg^{-1} , and days with θ_{def} achieving 0 K. Overall, both the changes in the frequency of

FIG. 9. (a) The change in afternoon total precipitation (mm) due to land-cover change from 1850 to 2000 based on the HCF-trigger simulations. Hatching indicates significance at the 95% confidence level. (b) The difference of afternoon precipitation change (mm) between the default and HCF triggering. The horizontal (vertical) hatching indicates significance at the 95% confidence level in HCF-trigger (default trigger) simulations. Boxes identify the two subregions for diurnal cycle analysis.

CAPE and θ_{def} show a consistent spatial pattern with the change in afternoon precipitation frequency, which also agrees well with the change in total afternoon precipitation (Fig. 9a). However, it should be noted that the frequency of θ_{def} shows more consistency with afternoon precipitation (with spatial correlation $r = 0.68$, $p < 0.01$) than CAPE (with spatial correlation $r = 0.61$, $p < 0.01$), indicating the θ_{def} criterion is more strongly related to convective triggering. In the HCF simulations, the θ_{def} criterion considerably delays the time of convective initiation, therefore allowing accumulation of CAPE, which tends to intensify precipitation events (Tawfik et al. 2017). In most cases, the CAPE criterion has been satisfied when θ_{def} reaches zero (Fig. S2), indicating that θ_{def} is a stricter criterion than $\text{CAPE} > 70 \text{ J kg}^{-1}$ within the HCF convective trigger.

Based on the changes in surface fluxes and atmospheric conditions, the disagreement in precipitation responses between the CTRL and HCF simulations can be explained. In the CTRL simulations, the significant widespread increase in precipitation over the crop area of the Great Plains is associated with strong positive land-atmosphere coupling. Over the central plains, where there is no precipitation increase in the HCF simulations, latent heat flux shows a weak positive relationship with θ_{def} (Figs. 7a,b, not significant) because of the competing effects of latent heat on $\theta_{2\text{m}}$ and θ_{BM} . In other words, higher latent heat flux may correspond to higher θ_{def} , which would impede convection initiation. Moreover, no relationship between latent heat flux and the probability of the θ_{def} criterion being satisfied

FIG. 10. The change in surface (a) latent and (b) sensible heat fluxes (W m^{-2}) due to land-cover change from 1850 to 2000 based on the HCF-trigger simulations. Hatching indicates significance at the 95% confidence level.

FIG. 11. The change in CAPE due to land-cover change from 1850 to 2000 based on the (a) HCF-trigger simulations and (b) default-trigger simulations. Hatching indicates significance at the 95% confidence level.

further explains the lack of precipitation change over the central plains in the HCF simulations.

Over the southeast (especially the lower Mississippi River basin), even though no significant relationship is found between latent heat and the magnitude of CAPE or θ_{def} , there is positive coupling between the LH and the probability of the θ_{def} criterion being satisfied (Fig. 8). Over the lower Mississippi River basin, the deforestation leads to decreased latent heat flux (Fig. 10), which may decrease the frequency of the θ_{def} criterion being satisfied and decrease afternoon precipitation frequency. We also see slightly decreased CAPE over the lower Mississippi River basin in the CTRL simulation (Fig. 11b), which is mainly associated with the decreased sensible heat flux after deforestation (Chen and Dirmeyer 2017). However, the local impacts on CAPE are not enough to modify the afternoon precipitation in the CTRL simulations. In the HCF simulations, θ_{def} acts as a stricter criterion than CAPE for convective triggering. As shown in Fig. S2, the θ_{def}

criterion allows considerable accumulation of CAPE, which is above 500 J kg^{-1} over the southeast when the θ_{def} criterion is satisfied. Therefore, the impacts on CAPE can be amplified over the regions with strong LH- $\Gamma_{\theta_{\text{def}}}$ relationship and further influence the afternoon precipitation. This can explain the greater increase in CAPE over the northern plains and decrease in the lower Mississippi River basin in HCF than CTRL (Fig. 11).

Finally, the diurnal cycle of precipitation with the two convective triggers is investigated over two subregions that exhibit significant precipitation responses to land-cover changes (Fig. 13). The boundaries of the two subregions are shown in Fig. 1. The first subregion is located in the northern Great Plains, where there is a significant increase in precipitation after trees and grass are replaced with crops. The second subregion is located in the lower Mississippi River basin, where precipitation significantly decreases after trees are replaced with grass and crops. The HCF simulation shows better agreement

FIG. 12. The change in frequency (days) of (a) afternoon precipitation greater than 1 mm, (b) afternoon CAPE greater than 70 J kg^{-1} , and (c) θ_{def} being equal to 0 K. Hatching indicates significance at the 95% confidence level estimated from 3000 bootstrap samples.

FIG. 13. Diurnal cycle of (a),(d) precipitation; (b),(e) CAPE; and (c),(f) θ_{def} in CTRL and HCF simulations (as well as from the two observations for precipitation) in the two subregions: (top) the northern Great Plains (42°–52°N, 110°–98°W) and (bottom) the lower Mississippi River basin (32°–37°N, 92°–86°W).

in the diurnal cycle of precipitation, especially over the Mississippi River basin (Fig. 13d). Uncertainties exist in the observed precipitation magnitude over the northern plains, as described previously, but there is a robust nocturnal maximum in both CMORPH and MSWEP (Fig. 13a). Compared to the CTRL simulations, the HCF simulations delay the precipitation maximum by about 4 h, although the timing is still not entirely consistent with the observations. Over both the regions, CAPE builds up and its maximum is delayed by a couple of hours in the HCF simulations because of the restriction of θ_{def} on convective initiation (Figs. 13b,e), which reaches its minimum in the late afternoon (Figs. 13c,f). With respect to land-cover change, it affects precipitation and CAPE throughout the day over the northern Great Plains. Over the Mississippi River basin, a slight decrease in precipitation can be found around noon in the CTRL simulations, but greater change is found in the afternoon in HCF simulations due to the delayed precipitation maximum.

Additionally, the mechanism within the land-atmosphere coupling process chain in terms of impacts of land-cover change on precipitation over the two

subregions is shown in Fig. 14. Over the northern Great Plains with a positive feedback in the CTRL simulations (Chen and Dirmeyer 2017), land-cover change leads to significantly higher morning latent heat flux ($+23.93 \text{ W m}^{-2}$) and lower sensible heat flux (-24.13 W m^{-2}) and higher afternoon CAPE ($+17.20 \text{ J kg}^{-1}$) and precipitation ($+0.10 \text{ mm}$), which in turn increases the soil moisture and latent heat flux. In the HCF simulations, this positive feedback is limited by the θ_{def} criterion, and we see a relatively small change in the surface fluxes and afternoon precipitation. Over the lower Mississippi River basin, no significant changes are found in morning latent heat flux and afternoon precipitation in the CTRL simulations. However, the increased θ_{def} ($+0.37 \text{ K}$) in the HCF simulations significantly reduces the afternoon precipitation, which in turn decreases the soil moisture and latent heat flux.

4. Discussion

This study has examined the sensitivities of land-atmosphere coupling strength to convective triggering parameterizations in CESM and its influence on

FIG. 14. Change in the components of the land–atmosphere coupling process chain over the two subregions due to land-cover change. Asterisks next to the numbers indicate significant changes at the 95% confidence level. Red asterisks between the bars indicate that the difference between the CTRL and HCF simulations are significant at the 95% confidence level. The θ_{def} is not included for the CTRL simulations because the HCF-based trigger is not used there.

precipitation feedback to land-cover change. Implementing a more physically based convective triggering (HCF) into CESM improves the model's simulation of the diurnal cycle of precipitation. The HCF convection trigger, which is a stricter criterion for convection, also acts to change land–atmosphere coupling strength in both atmospheric and terrestrial legs. Because of the shift in land–atmosphere coupling strength with the HCF trigger, the impacts of land-cover change on afternoon precipitation are also affected.

With the default triggering mechanism, [Chen and Dirmeyer \(2017\)](#) found a strong positive soil moisture–precipitation relationship over the Great Plains and a significant widespread increase in precipitation as a result of land-cover change over this region. The HCF trigger, on the other hand, exhibits weaker land–atmosphere coupling over the central plains with weakened coupling between soil moisture and surface latent heat flux (terrestrial coupling leg) and negligible coupling between latent heat flux and θ_{def} (atmospheric

coupling leg). Because coupling strength is low, implying that convection and precipitation are not strongly dependent on land fluxes, the impact of land-cover change on afternoon precipitation, even when surface fluxes are modified because of the land-cover change, is also low.

Previous studies have documented land–atmosphere coupling in climate models and concluded that their land–atmosphere coupling strength may be excessive (e.g., [Ferguson et al. 2012](#); [Phillips and Klein 2014](#); [Song et al. 2016](#)). Based on ARM measurements in the southern Great Plains, [Phillips and Klein \(2014\)](#) suggest that large-scale atmospheric forcings dominate over land–atmosphere interactions, while local feedbacks of the land on the atmosphere are comparatively small much of the time. [Ruiz-Barradas and Nigam \(2013\)](#) arrive at a similar conclusion through analysis of the North American Regional Reanalysis, where they find no significant relationship between daytime-average precipitation and the corresponding evaporative fraction (or soil moisture) averages. Moreover, there is debate

on the relationship between soil moisture and precipitation. In contrast to most model-based studies, which show a positive soil moisture–precipitation feedback over the arid–humid transitional regions, [Taylor et al. \(2012\)](#) found that convective precipitation falls preferentially over dry soils where the enhanced afternoon moist convection is driven by increased sensible heat flux. [Guilod et al. \(2015\)](#) suggest that spatial versus temporal variations in soil moisture allow the two mechanisms to operate simultaneously. Over the United States, [Tuttle and Salvucci \(2016\)](#) found a positive (negative) feedback of soil moisture on next-day precipitation probability in the western (eastern) United States. However, they concluded that a statistically significant soil moisture–precipitation feedback is not detectable over much of the Great Plains. This region may mark a transition between regimes (cf. [Findell and Eltahir 2003](#)) where the coupling regime is hard to detect climatologically, yet strong one way or the other on daily to subseasonal time scales. Compared with the default triggering mechanism, which is characterized by excessive land–atmosphere coupling strength, the HCF trigger provides a more physically based real-time representation of land–atmosphere interactions, with weaker and potentially more realistic coupling strength. However, observational studies are needed to further understand the role of θ_{def} in the land–atmosphere coupling process chain. The superiority of the θ_{def} criterion over the CAPE criterion for convective triggering also needs further observationally based investigation.

Another novel aspect of this study is the investigation of impacts of land-cover change on precipitation with two different convective triggering conditions in CESM. There is agreement in precipitation change over the northern Great Plains, but discrepancies over the southern plains, which cast uncertainties on previous land-cover change sensitivity studies. Within a single model, we find the adjustment of the deep convective parameterization has led to a different response of precipitation to land-cover change. [Brovkin et al. \(2013\)](#) examined six models in phase 5 of the Coupled Model Intercomparison Project (CMIP5) to assess the uncertainties in climatic effects of land-cover change due to differences in model parameterizations and implementation of LULCC data. However, only the parameterizations of land surface processes were discussed. [Hirsch et al. \(2015\)](#) show that the impacts of land-use change on regional temperature extremes depend on land–atmosphere coupling, which is modulated by the choice of planetary boundary layer schemes. Our results further demonstrate that the difference in convective parameterization of the coupled atmospheric model could also be responsible for the uncertainties in

land–atmosphere coupling strength and precipitation response. The magnitude and spatial patterns of precipitation response due to land-cover change can be greatly affected by the treatment of convection in ESMs, which also leads to different land–atmosphere coupling strengths. Therefore, caution should be taken when investigating the climatic effects of land-cover change even from multimodel experiments, such as the upcoming Land-Use Model Intercomparison Project (LUMIP; [Lawrence et al. 2016](#)).

Additionally, in this study the HCF trigger that we used does not consider the subgrid land surface heterogeneity, which could be important, especially for LULCC studies. It would be interesting to repeat our experiments with the subgrid HCF trigger.

5. Conclusions

This paper presents a follow-up study of [Chen and Dirmeyer \(2017\)](#) on the impacts of land-cover change on afternoon precipitation over North America, with the focus on how different convective triggering mechanisms influence the land cover–precipitation feedback in CESM. We have implemented a physically based convective trigger (HCF) as an additional triggering criterion to the CAPE-based criterion for deep convection in CESM. Compared with two observational datasets (CMORPH and MSWEP) with high spatial and temporal resolution, the simulation with the HCF trigger exhibits an improvement in the representation of the diurnal cycle of summer precipitation, but the model's dry bias has gotten worse over much of the United States.

To explore the mechanisms of land cover–precipitation feedback, we have examined the land–atmosphere coupling indices for both terrestrial and atmospheric legs. Overall, the atmosphere is less sensitive to the land surface over the Great Plains in the HCF simulation compared to the CTRL simulation. There is weakened coupling strength between soil moisture and surface turbulent fluxes. For the atmospheric leg, there is strong coupling between morning latent heat flux and CAPE, which is also found in [Chen and Dirmeyer \(2017\)](#). However, the additional HCF convective triggering criterion, the surface potential temperature deficit θ_{def} , is not significantly coupled to the surface fluxes over the central Great Plains, leading to low overall land–atmosphere coupling strength with the HCF trigger.

In contrast with the CTRL simulations, in which significant widespread increases in afternoon precipitation are detected over the whole Great Plains area, a region that has experienced extensive agricultural expansion

since preindustrial times (Chen and Dirmeyer 2017), the HCF-trigger simulations only suggest a significant increase in afternoon precipitation over the northern Great Plains with a significant decrease over the lower Mississippi River basin and no change in afternoon precipitation over the central plains due to the insensitivity of θ_{def} to surface fluxes over this region. The change in precipitation is mainly associated with the frequency at which deep convection is triggered, and the θ_{def} criterion is more dominant in convective triggering than the CAPE criterion. The discrepancy in the precipitation response to land-cover change depending on the convective trigger implies that care must be taken when investigating the impacts of land-cover changes on precipitation in ESMs.

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