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Comprehensive Study on the Influence of Evapotranspiration and Albedo on Surface Temperature Related to Changes in the Leaf Area Index

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ABSTRACT

Many studies have investigated the influence of evapotranspiration and albedo and emphasize their separate effects but ignore their interactive influences by changing vegetation status in large amplitudes. This paper focuses on the comprehensive influence of evapotranspiration and albedo on surface temperature by changing the leaf area index (LAI) between 30°–90°N. Two LAI datasets with seasonally different amplitudes of vegetation change between 30°–90°N were used in the simulations. Seasonal differences between the results of the simulations are compared, and the major findings are as follows. (1) The interactive effects of evapotranspiration and albedo on surface temperature were different over different regions during three seasons [March–April–May (MAM), June–July–August (JJA), and September–October–November (SON)], i.e., they were always the same over the southeastern United States during these three seasons but were opposite over most regions between 30°–90°N during JJA. (2) Either evapotranspiration or albedo tended to be dominant over different areas and during different seasons. For example, evapotranspiration dominated almost all regions between 30°–90°N during JJA, whereas albedo played a dominant role over northwestern Eurasia during MAM and over central Eurasia during SON. (3) The response of evapotranspiration and albedo to an increase in LAI with different ranges showed different paces and signals. With relatively small amplitudes of increased LAI, the rate of the relative increase in evapotranspiration was quick, and positive changes happened in albedo. But both relative changes in evapotranspiration and albedo tended to be gentle, and the ratio of negative changes of albedo increased with relatively large increased amplitudes of LAI.

Key words: surface temperature, evapotranspiration, albedo, leaf area index, comprehensive influence

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1. Introduction

As the interface between the land surface and the atmosphere, vegetation plays a significant role in regulating the land–atmosphere interaction, thereby affecting regional and global climates. Its variations influence the exchange of heat, mass, and momentum between the land surface and the lower atmosphere (Dickinson et al., 1992; Bonan, 1994; Guillevic et al., 2002; Kang et al., 2007; Li and Xue, 2010). Pielke et al. (1998) suggested that vegetation dynamics might be as important for climate as other forcings, such as atmospheric dynamics and composition, ocean circulation, ice sheet extent, and orbit perturbations.

In recent decades, many studies have been conducted to detect the effects of vegetation on albedo (Charney et al., 1975; Sud and Fennessy, 1982; Bonan et al., 1992; Claussen et al., 2001), evapotranspiration (ET; Shukla and Mintz, 1982; Chen and Zeng, 2012; Zhu and Zeng, 2014), and surface roughness length (Sud et al., 1988). Generally speaking, increased vegetation over the Northern Hemisphere is expected to reduce surface albedo because of an increase in the amount of solar radiation absorbed by vegetation, which may result in further regional or even global climate warming. For example, Levis et al. (2000) showed that this process could increase global warming during the 21st century when considering the northward expansion of the boreal forests as a result of climatic warming. Bonfils et al. (2012) also determined that an invasion of the tundra by tall shrubs tended to systematically warm the soil. However, increased vegetation also accompanies intensified evapotranspiration, which is favorable for cloud development and causes surface cool-

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ing (Bounoua et al., 2000; Buermann et al., 2001). These opposing biophysical effects of vegetation—warming through increased energy absorption (relatively low albedo) and cooling through increased evaporation—tend to dominate at different latitudes, depending on the regional climate characteristics and geographical features (Meir et al., 2006; Xue et al., 2010). Many studies emphasize one aspect of the effects of increased vegetation, but ignore its interactive influences. Therefore, the goal of this paper is to discuss the comprehensive impacts of evapotranspiration and albedo on surface temperature. Throughout this entire paper, we will always maintain the concept that the importance of evapotranspiration and albedo for surface temperature is on the same level.

Many previous studies used relatively large amplitude changes to investigate the influence of vegetation on temperature. For instance, Dickinson and Henderson-Sellers (1988) and Shukla et al. (1990) replaced tropical forests with grass, which led to warmer and drier conditions. Over the mid–high latitudes, simulations with boreal forest result in warmer conditions than those with bare ground or tundra (Bonan et al., 1992). More recently, both Bounoua et al. (2000) and Buermann et al. (2001) applied maximum and minimum leaf area index (LAI) values from yearly satellite records to investigate the impact of extreme variability in the amount of vegetation on temperature. These experiments provide valuable references for understanding the influence of vegetation on climate but also may bracket these influences with relatively large changes in the amount of vegetation. However, in this paper, the results are discussed with different ranges of LAI changes. Furthermore, changes in climate are expected to be larger over the mid–high latitudes relative to the tropics. The entire poleward shift of the boreal ecosystem that arises from regional warming and increasing CO_2 levels may have significant impacts on climate, including changes in precipitation patterns, carbon levels, and the energy balance (Lucht et al., 2006; Schaphoff et al., 2006; Alo and Wang, 2008; O'Ishi and Abe-Ouchi, 2009). Thus, the region over 30°–90°N was selected as the research focus.

In the next section, the model and LAI datasets that were applied in this study, as well as the experimental design, are described. Section 3 presents the results and discussion, emphasizing the comprehensive effects of evapotranspiration and albedo on surface temperature. Finally, concluding remarks are given in section 4.

2. Model, data, and experimental design

2.1. *Model description*

In this study, we employed the coupled model of the Community Earth System Model (CESM), including the atmosphere and land components of the Community Atmosphere Model version 4 (CAM4; Neale et al., 2013) and the Community Land Model version 4 (CLM4; Oleson et al., 2010; Lawrence et al., 2011), respectively. CLM4 represents the fundamental physical, chemical, and biological processes of the terrestrial ecosystem and describes the water, energy, and carbon–nitrogen processes by coupling with the carbon–nitrogen model (CN) and the dynamic vegetation model (CNDV, Castillo et al., 2012). When the CN or CNDV model is inactive, the prescribed LAI data is used as the boundary condition.

2.2. *Leaf area index*

Two sets of LAI data were applied in this study. One was the LAI used in CLM4 (the default LAI), which is the climatological mean with monthly variability at a 0.9° lat ×1.25° lon horizontal resolution. Figure 1 shows its spatial differences between March–April–May (MAM) and December–January–February (DJF), June–July–August (JJA) and DJF, as well as September–October–November (SON) and DJF over 30°–90°N. It is in line with expectations that the average LAI in MAM, JJA and SON should be larger than that in DJF. By and large, the amplitudes of the differences between JJA and DJF are largest, followed by those between SON and DJF, and then MAM and DJF. Therefore, for the purposes of investigating the influence of a change in the LAI in different and reasonable ranges, we constructed another dataset, the idealized LAI data, from the default LAI dataset. For the idealized LAI, each monthly value over 30°–90°N was the averaged value of DJF of the default LAI, whereas over other regions, it was the same as the default LAI. Consequently, by analyzing the seasonal results of the simulations using the two LAI datasets, the influence of a change in the seasonal amplitudes of the LAI could be investigated.

2.3. *Experimental design*

Two simulations, CTL and IDL, utilizing the default LAI and the idealized LAI, respectively, were performed. Beyond this, the two simulations were totally identical. For example, the atmosphere and land components were active in both simulations, and they were both driven by historical sea surface temperatures for 1979–2003 (Hurrell et al., 2008). Both simulations were run at a 0.9° lat × 1.25° lon horizontal resolution with the carbon and nitrogen cycles turned off.

3. Results and discussion

In this section we discuss the seasonal changes to evapotranspiration and albedo, which were directly influenced by the changes to the LAI, and emphasize their comprehensive influences in determining surface temperature. When referring to differences in the following sections, we mean CTL minus IDL averaged over the last 23 years of the simulations (1981–2003).

3.1. *MAM*

Generally, the CTL's surface temperature increased over high latitudes and decreased over mid latitudes when compared to IDL (Fig. 2). The main statistically significant regions, indicated by the three boxes in the figure, were selected as the objective areas. Over the southeastern United States (area 1), the increased LAI led to an increase in both

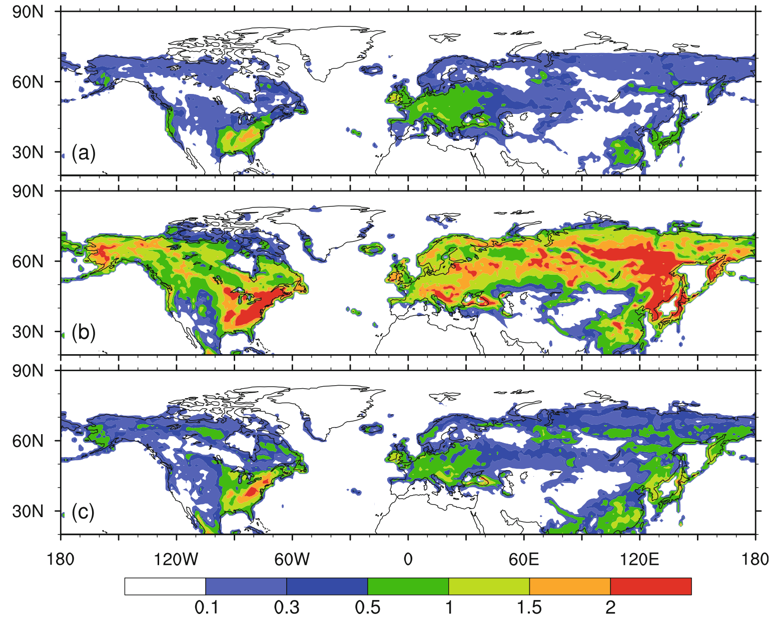

Fig. 1. The differences of LAI between (a) MAM and DJF, (b) JJA and DJF, and (c) SON and DJF. The LAI data are from the surface dataset of CLM4 and all units are $m^2\ m^{-2}$.

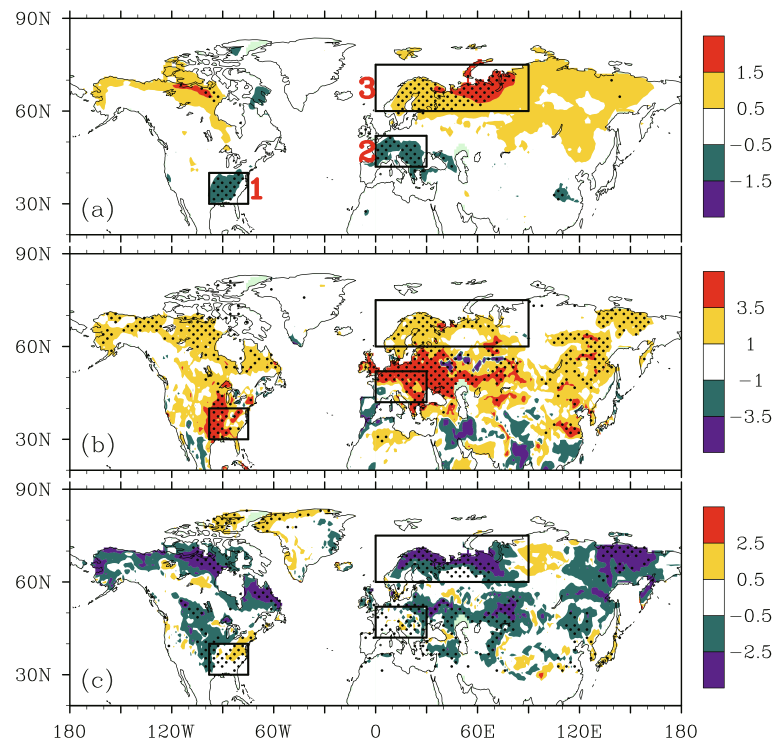

Fig. 2. Average differences between CTL and IDL (CTL−IDL) in (a) surface temperature (K), (b) evapotranspiration (mm $month^{-1}$), and (c) albedo (%) during MAM. The stippled areas represent areas of statistical significance ($P < 0.05$). The boxes named area 1, area 2, and area 3 indicate the regions over which the area averaging was made to construct Table 1.

evapotranspiration (6.37 mm $month^{-1}$, or 8.7%) and albedo (0.025, or 2.1%), which together contributed to the decrease in surface temperature (Table 1). Over southern Europe (area 2), the surface temperature in CTL also decreased compared to IDL, which mainly resulted from the cooling effects of increased evapotranspiration (4.76 mm $month^{-1}$, or 8.7%)

Table 1. Area averaged differences (CTL−IDL) in LAI (units: $m^2\ m^{-2}$), evapotranspiration (ET, mm $month^{-1}$), albedo (%), and surface temperature (T, units: K) over selected regions during MAM, JJA, and SON. Only these grids, over which the differences of surface temperature, evapotranspiration, and albedo are all statistically significant ($P < 0.05$) were selected for calculation. The number in parentheses is the relative differences compared to IDL.

	Regions	Δ LAI	Δ ET	Δ Albedo	ΔT
MAM	Area 1	1.29 (104.0%)	6.37 (8.7%)	0.25 (2.1%)	−0.81
	Area 2	0.84 (160.0%)	4.76 (8.7%)	−0.2 (−1.4%)	−0.84
	Area 3	0.22 (19.0%)	2.51 (15.0%)	−4.28 (−12.2%)	1.58
JJA	Area 1	2.08 (174.0%)	14.92 (14.4%)	0.03 (0.3%)	−1.89
	Area 2	1.57 (223.2%)	17.68 (25.3%)	−0.88 (−6.2%)	−1.91
	Area 3	1.62 (183.1%)	14.92 (27.7%)	−0.49 (−3.8%)	−1.84
	Area 4	0.96 (226.9%)	18.38 (30.2%)	−3.16 (−16.7%)	−1.36
	Area 5	1.28 (226.8%)	14.43 (22.2%)	−2.81 (−16.3%)	−1.62
	Area 6	2.29 (1031.2%)	29.96 (94.8%)	−1.58 (−11.1%)	−2.74
SON	Area 1	1.54 (121.2%)	4.69 (9.8%)	0.74 (6.0%)	−1.58
	Area 2	0.74 (110.7%)	2.63 (9.1%)	0.24 (1.7%)	−0.87
	Area 5	0.29 (241.4%)	−0.36 (−1.9%)	−2.70 (−11.3%)	0.98

that dominated the warming impacts of the slightly decreased albedo (−0.02, −1.4%). Over northwestern Eurasia (area 3), however, the albedo in CTL decreased by 0.043, or by about 12.2%, compared to the change in IDL, which dominated the influence of increased evapotranspiration (2.51 mm $month^{-1}$, or 15.0%) on surface temperature. Consequently, the CTL's surface temperature increased 1.58 K on average. In general, over the three regions, the changes to the surface temperature resulted from the comprehensive influence of evapotranspiration and albedo. In this process, their roles were different, with evapotranspiration being dominant over area 1 and area 2 but albedo being dominant over area 3.

3.2. *JJA*

Compared to IDL, CTL's evapotranspiration increased significantly, whereas its albedo decreased over most regions in 30°–90°N (Fig. 3). Correspondingly, surface temperature of CTL significantly decreased in comparison to that of IDL. Therefore, it is implied that the cooling effects of increased evapotranspiration dominated the warming influences of decreased albedo over most regions in the range 30°–90°N. However, albedo still played a role in regulating surface temperature, as illustrated by the spatial heterogeneity of the differences in surface temperature, exceeding 2 K over eastern United States and eastern Russia and below 2 K over other regions, such as central North America and central Eurasia. The changes in albedo also showed a large spatial heterogeneity, but they were large over central North America and central Eurasia and exceeded 2.5%. Relatively, the differences of evapotranspiration over these regions were less heterogeneous than those of albedo, and all exceeded 15 mm $month^{-1}$. Therefore, the decreased albedo of CTL reduced, to some extent, the decrease in amplitudes of the surface temperature induced by the increase in evapotranspiration.

To further compare the effects of evapotranspiration and albedo, three more areas—central North America (area 4), central Eurasia (area 5), and eastern Russia (area 6) in Fig. 3—were selected in addition to the three areas chosen during MAM. Over area 4 and area 5, the changes to albedo were relatively large, while over area 6, changes to evapotranspiration were relatively large. Over area 1, the negligibly increased albedo (0.3%) together with the increased evapotranspiration (14.92 mm $month^{-1}$, or 14.4%) resulted in a decrease in surface temperature (1.89 K). This was consistent with what happened during MAM, although the increased amplitudes of albedo were smaller than those of MAM. Over area 2 and area 3, the averaged effects of evapotranspiration and albedo on surface temperature were opposite. Finally, cooling effects of increased evapotranspiration (25.3%, 27.7%) offset, to some extent, the warming impacts of the decrease in albedo (6.2%, 3.8%) and led to a decrease in the surface temperature. Over area 2 and area 3, the albedo did not change much, and thus the average surface temperature decreased up to 1.91 and 1.84, respectively. Over area 4 and area 5, however, the albedo in CTL compared to IDL decreased by 16.7% and 16.3%, respectively, which weakened the cooling effects of the increase in evapotranspiration to a large extent. As a result, the average decrease in the surface temperature was 1.36 K and 1.62 K, respectively, which were reduced by 28.8% and 15.2%, respectively, compared to that of area 2. Over area 6, although the decrease in the albedo also exceeded 10%, the absolute and relative increase in the evapotranspiration was too large, and consequently the average decrease in the surface temperature reached up to 2.74 K. In short, when changes in the LAI happened in relatively large amplitudes during JJA, the cooling effects of increased evapotranspiration dominated over almost all regions in the range of 30°–90°N. However, the corresponding albedo changes still played a significant role in regulating the surface temperature, which contributed to the heterogeneity of the changes in surface temperature.

3.3. *SON*

In general, compared to IDL, surface temperature in CTL significantly decreased over eastern North America, western Europe, and eastern China, while it increased over central

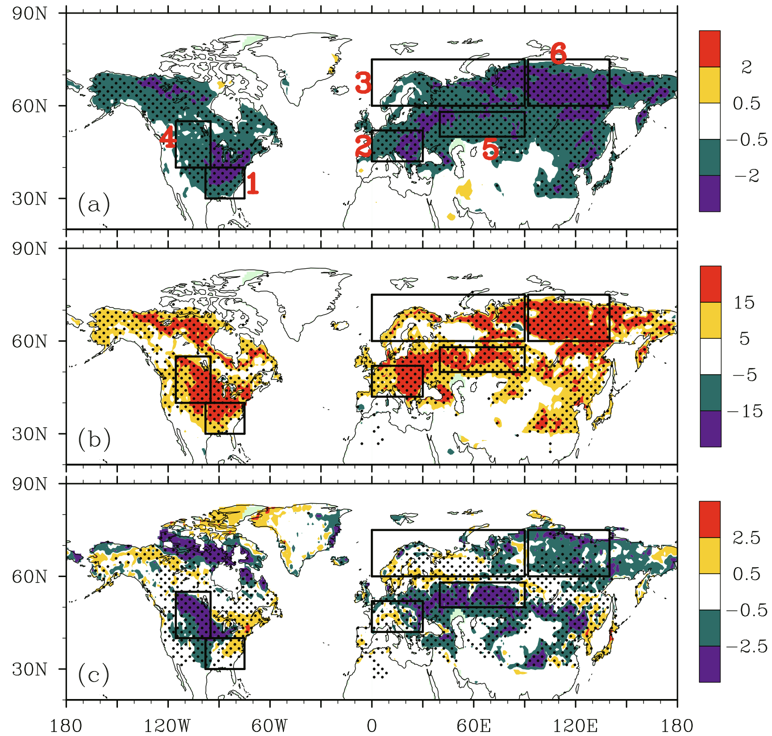

Fig. 3. As in Fig. 2, but for JJA.

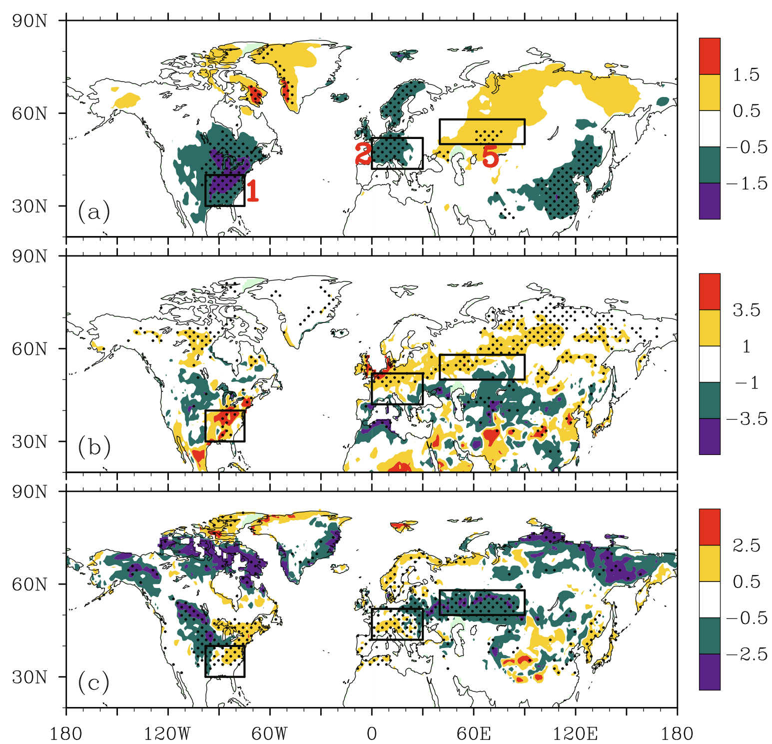

Fig. 4. As in Fig. 2, but for SON.

Eurasia (Fig. 4). Therefore, the southeastern United States (area 1), southern Europe (area 2), and central Eurasia (area 5), discussed in section 3.2, were also selected for comparison to the above results during MAM and JJA.

Over area 1, the results of SON were also in line with observations from during MAM and JJA, i.e. the increased evapotranspiration (4.69 mm month^{-1}, or 9.8%) and the increased albedo (6.0%) were both responsible for the average decrease in the surface temperature (1.58 K). Over area 2, the increased LAI led to an increase in the evapotranspiration and

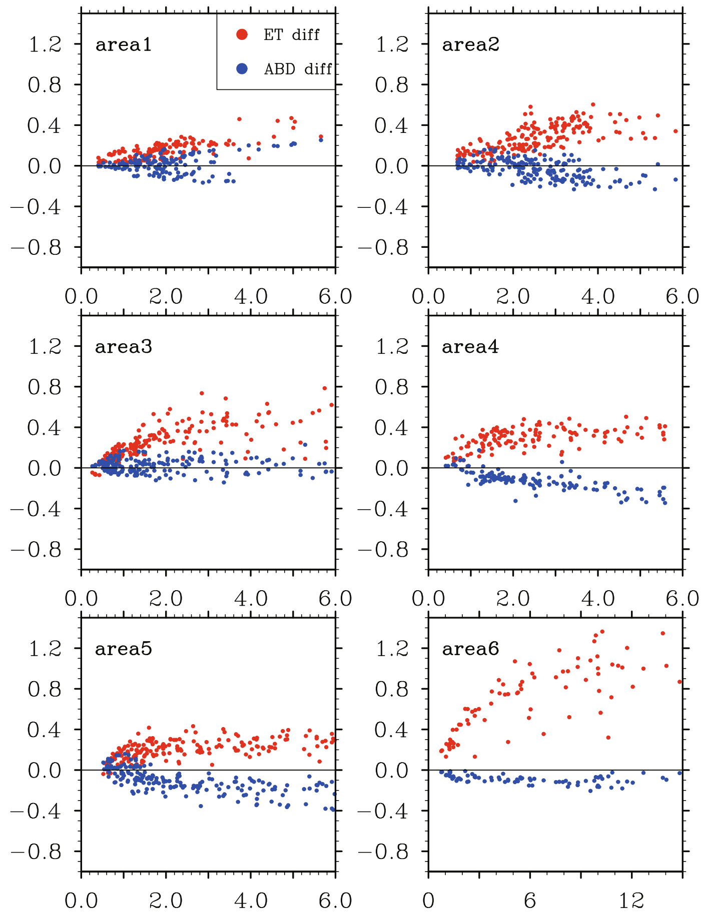

Fig. 5. Scatterplots of the relative changes of CTL's evapotranspiration (ordinate, ET, red) and albedo (ordinate, ABD, blue) over the six selected areas during JJA, as well as the relative amplitudes of changes in the LAI (abscissa). The relative changes mean (CTL−IDL)/IDL. Only these grids, over which the differences of surface temperature, evapotranspiration, and albedo are all statistically significant ($P < 0.05$) were selected.

albedo, which both contributed to the decrease in surface temperature. This was different from the changes during MAM and JJA, when the effects of changes in the evapotranspiration and albedo on the surface temperature were opposite. Over area 5 during SON, CTL's surface temperature increased by an average of 0.98 K in comparison to that of IDL. This was mainly because of a significant decrease in albedo (11.3%). Meanwhile, the slight decrease in the evapotranspiration also made a certain contribution. Overall, during SON, the effects of changes in the evapotranspiration and albedo on the surface temperature of area 1 were consistent with the changes during MAM and JJA, whereas over area 2 and area 5, their roles were different from those observed during MAM and JJA.

3.4. *Discussion*

The scatterplots of the relative changes to CTL's evapotranspiration and albedo compared to those of IDL over six selected areas during JJA (Fig. 5) show consistent results with Fig. 3b, in which over almost all of the grids CTL's evapotranspiration increased when compared to IDL. This was mainly because the presence of vegetation was favorable for increasing the evapotranspiration of the area by extracting additional water from the soil via roots (Meir et al., 2006). Furthermore, our results also show that, along with changes to the LAI, the relative changes in the evapotranspiration did not increase at the same pace. Over the six selected areas (Fig. 5), when the relative amplitudes of changes to the LAI were below a value of about 200% for area 1 to area 5 and a value of 600% for area 6, the relative increase in the amount of evapotranspiration was quick, but it tended to be gentle when the relative amplitudes of the changes to the LAI exceeded those values. Guillevic et al. (2002) suggested that the impact of LAI variability on canopy evapotranspiration is determined by both external environmental factors and internal vegetation factors. They emphasized the effects of saturation over dense vegetation and water stress. Therefore, Fig. 5 further reveals that the changes in the evapotranspiration were limited by LAI when the relative amplitudes of the changes to the LAI were smaller; whereas when they became larger, the changes to the evapotranspiration were limited mainly by other external environmental factors, i.e., precipitation or soil moisture. On the other hand, the albedo also showed different results that corresponded to different relative amplitudes of changes to the LAI. Positive changes happened to the albedo when the relative amplitudes of the changes to the LAI were smaller; and when the amplitudes of the changes became larger, the ratio of grids over which negative changes happened to the albedo increased. Similar to the results observed in changes to the evapotranspiration, the changes to the albedo also tended to be gentle, as the relative amplitudes of the changes to the LAI were sufficiently large.

Therefore, the different effects of evapotranspiration and albedo on surface temperature in different regions and seasons can be explained by the combination of above changed responds and external environmental factors. First, when the amplitudes of the increases to the LAI are relative small, the quick increase in the evapotranspiration and the positive changes to the albedo together contribute to a decrease in surface temperature. For example, over area 1 during MAM and SON and area 2 during SON, the amplitudes of the relative changes to the LAI were around 100%, and consequently both evapotranspiration and albedo increased and then together resulted in a decrease to the surface temperature (Table 1). However, the external environment also plays important roles. For example, over area 3 during MAM, although the average amplitude of the relative change to the LAI was small (19%), the change to the albedo was still negative. In this region boreal trees and shrubs stand above snow cover, the large contrast between leaf and snow albedo may have strong impact on land surface energy balance, further increase the albedo difference between the CTL and IDL experiments. Second, when the amplitudes of the increase to the LAI were relatively large, the warming effects of the decrease in albedo were reversed with the cooling influences from the increase in evapotranspiration. Either evapotranspiration or albedo tended to be dominant depending on the climate's characteristics and relative changes of themselves. For instance, over almost all regions in the range of 30°–90°N during JJA, evapotranspiration played a dominant role in regulating the surface temperature in comparison to albedo, which is also because of the relative sufficient precipitation. In contrast, during SON, the significant decrease in the albedo dominated area 5 when compared to the slight changes in the evapotranspiration due to not sufficient precipitation.

Of course, the characteristics of LAI change are strong correlated to the feature of local vegetation composition, in particular the dominant plant functional type (PFT). For example, evergreen and deciduous plants have different phenology, and woody plants (trees and shrubs) behave different with grasses during snow cover. Our current work focuses on the regional effects of changed LAI, further study may investigate the different impacts of PFTs within a gridcell, as well as the regional impacts based on the distribution of different plant functional types.

This paper emphasizes the impacts of local land–atmosphere interactions on surface temperature. However, local climate (e.g., temperature, precipitation, etc) is also influenced by the change in large scale atmospheric circulation, which may represent the regional and even global LAI changes between the two experiments. Further experiments are needed to quantitate the contributions from different regions.

4. Conclusion

The CESM was used to investigate the comprehensive influence of evapotranspiration and albedo on surface temperature under seasonal situations with different ranges of changes to the LAI.

The interactive effects of evapotranspiration and albedo on the surface temperature were different for six selected areas during the three chosen seasons (MAM, JJA, and SON). For the southeastern United States (area 1) during all three seasons, as well as for southern Europe (area 2) during SON, increases to both evapotranspiration and albedo contributed to cooling effects and resulted in decreases to the surface temperature. However, for the other five regions during JJA, the average albedo decreased, resulting in warming effects, which were reversed with the cooling influences of the increases to the evapotranspiration. In addition, either evapotranspiration or albedo tended to be dominant over different areas and different seasons. During JJA, the cooling effects of increases in the evapotranspiration dominated the warming influences of decreases in albedo over almost all regions in the range of 30°–90°N. However, over northwestern Eurasia (area 3) during MAM and central Eurasia (area 5) during SON, the changes to the albedo played a dominant role in determining the surface temperature in comparison to changes in the evapotranspiration.

The response of the evapotranspiration and albedo to different increases in the LAI showed distinct paces and signals. When the amplitudes of the increase to the LAI were relatively small, the rate of the relative increase in evapotranspiration was quick; whereas, it tended to be gentle when the amplitudes of increase in the LAI became relatively large. Meanwhile, positive changes happened to the albedo when the amplitudes of increase in the LAI were relatively small. When the amplitudes of the changes became larger, the ratio of grids over which the negative changes happened to the albedo increased, and ultimately the decreases in the amplitudes also tended to be gentle. Therefore, these effects explain the changing interactive and dominant influences of evapotranspiration and albedo on surface temperature.

This paper has highlighted the comprehensive influence of evapotranspiration and albedo on surface temperature and discussed their changes in relation to a range of changes to the LAI. The results were based on six selected areas during three seasons. More studies are necessary to investigate the effects over other regions. Furthermore, the characteristics of the response of evapotranspiration and albedo to changes in the LAI using different ranges need to be further investigated so as to identify the possible relationships between the results and climatic characteristics or vegetation types.

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