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Climate variability in monsoon and arid regions attributable to
dynamic vegetation in a global climate model

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Abstract

Climate variability in monsoon and arid regions attributable to dynamic vegetation is investigated by using NCAR’s Community Earth System Model (CESM) with the Dynamic Global Vegetation Model. Two present climate simulations, one with dynamic and the other with fixed vegetation cover, are carried out. A comparative analysis of the two simulations reveals that climate in monsoon and arid regions show different responses to the dynamic vegetation. On hemispheric scale, precipitation mainly exhibits increase in the Northern Hemisphere and decrease in the Southern Hemisphere in response to dynamic vegetation, and the surface temperature shows consistent decrease. On regional scale, the precipitation decrease caused by dynamic vegetation is the main trend in the monsoon regions except the Asian monsoon region, whereas the responses of precipitation to vegetation change is weak in the arid regions relative to monsoon regions. The surface temperature increased significantly due to dynamic vegetation only in the boreal winter Asian monsoon region, whereas the rest of monsoon regions and the arid regions mainly shows decreased surface temperature. Therefore, climate variability in the Asian monsoon region is obviously different from other regions. Further analysis shows that the dynamic vegetation can modulate the variations of the east-west sea level pressure gradient and the lower-level meridional winds in East Asia, and strengthen (weaken) East Asian summer (winter) monsoon. Mechanistic analysis reveals that the difference in hemispheric and regional climate variations may be due to the changes of dynamic vegetation-induced moisture flux and net surface radiative forcing.
Key words Dynamic vegetation; Monsoon region; Arid region; Climate variability;

East Asian monsoon
1. Introduction

Precipitation is an example of a natural phenomenon that can restrict global socio-economic development. Around two thirds of the world’s human population lives in the monsoon regions, where precipitation imposes significant socio-economic impacts (Wang et al. 2012). Precipitation also plays an important role in arid regions. Due to their fragile ecosystems, arid regions are more sensitive to global change (Rotenberg and Yakir 2010). In recent years, observed precipitation has increased in monsoon regions and decreased in arid regions under ever increased global warming (Wang et al. 2012).

Vegetation modifies the energy exchange of the land surface–atmosphere system by slowly altering land surface characteristics and imposing intricate impacts on the terrestrial hydrologic cycle. Thus, vegetation–climate interaction attracts a great deal of attention in the climate research community (Claussen 1998; Notaro et al. 2007, 2011, 2017; O’ishi and Abe-Ouchi 2009; Quillet et al. 2010; Delire et al. 2011; Maneta and Silverman 2013; Jiang et al. 2015). Using dynamic global vegetation models coupled with atmospheric general circulation models, numerous authors have studied the impact of dynamic vegetation on climate. Through analyzing the climate variability resulted from the responses of global land to the dynamic vegetation, Sun and Wang (2014) found that dynamic vegetation significantly influences the precipitation variation over the Amazon region. They also found that the impacts on temperature were strong over the US Great Plains region in all four seasons, and in the Amazon region during the dry and dry-to-wet transition seasons. In addition, many studies have focused on the
impacts of dynamic/interactive vegetation at regional scales, e.g., a present-day strengthening of the South Asian summer monsoon and weakening of the western North Pacific monsoon (Li et al. 2009), enhancement of the interdecadal variation of rainfall in the West African Sahel (Zeng et al. 1999), and positive temperature feedback over Asiatic Russia (Notaro and Liu 2008). However, few studies focused on comparing the climate sensitivity to dynamic vegetation between monsoon and arid regions. Dynamic vegetation affects the albedo, evapotranspiration and roughness of the land surface (Delire et al. 2011), and the land area is larger in the Northern Hemisphere (NH) than in the Southern Hemisphere (SH), hence the land–atmosphere feedback may be much more intricate in the NH than in the SH. We hypothesize that the land–sea area ratios between the NH and SH induces distinct impacts of dynamic vegetation on climate between the global monsoon and arid regions. Based on this assumption, we first describe the model experimental setup, model validation and the study regions in section 2. Then, in section 3, we examine the differences and similarities of the dynamic vegetation–induced precipitation and surface temperature responses between monsoon and arid regions at the hemispheric and regional scales, and we also clarify the possible mechanism. In section 4, we analysis how the dynamic vegetation affect the large-scale circulation in East Asian monsoon region. Section 5 provides a summary of the study and offers some further discussion.

2. Experiments and methodology

2.1 Model and experiments

We perform the numerical experiments using the NCAR’s Community Earth
System Model, version 1.0.4 (CESM), with a dynamic global vegetation model (DGVM) included in the land component, CLM4 (Community Land Model, version 4) (Hurrell et al. 2013). It is run at a horizontal resolution of approximately 0.9° latitude × 1.25° longitude (192 × 288 grid cells) and with 26 atmospheric levels in the vertical direction. Meanwhile, in the ocean component, the default gx1v6 resolution (a displaced pole grid with a 1° resolution) is used. Using a DGVM coupled with CESM has been successful in studying the feedback of dynamic vegetation to climate variability and carbon flux (Jiang et al. 2013; Peng and Dan 2015; Qiu and Liu 2015). Further details of CESM and the coupled CLM4-DGVM module can be found in Hurrell et al. (2013) and Oleson et al. (2010).

Two equilibrium climate simulations are executed under 2000 AD conditions, i.e., with the solar constant (1365 W/m²), CO₂ concentration (367 ppm) and other greenhouse gases, and sea surface temperature (SST) held constant at the level of 2000 AD. One simulation is performed with dynamic vegetation which changes in time and space and interacts with the climate (DV run), and another with fixed vegetation cover (CTL run). Both simulations are run for 60 years after spin-up, and the last 30 years are analyzed. Using prescribed SST allows us to eliminate the climate response related to the ocean and ensure that the differences in climate between the DV and CTL runs are attributable only to the interaction between the dynamic vegetation and the atmosphere. Hence, based on the model outputs, we can evaluate the effects of dynamic vegetation on present-day climate.
2.2 Validation of the model

To assess the performance of the model, the simulated patterns of the climatological annual mean of precipitation and surface temperature are compared with the CPC (Climate Prediction Center) Merged Analysis of Precipitation (CMAP) (Xie and Arkin 1997) and NCEP–DOE (National Centers for Environmental Prediction–Department of Energy) Reanalysis 2 data (NCEP2) (Kanamitsu et al. 2002). Only a small difference in precipitation distribution is observed in the tropics, showing that the simulated precipitation in DV run is in good agreement with the observed climate, with the correlation coefficient of 0.84 (Fig. 1b). The model represents the pattern of surface temperature excellently, and the correlation coefficient between the simulated and observed surface temperature reaches 0.98 (Fig. 1d). For CTL run (figures not shown), the correlation coefficient between the simulated and observed precipitation (surface temperature) is 0.85 (0.99). These results demonstrate that the simulated precipitation and surface temperature climatology in the model are comparable to the climate observations in CMAP and NCEP2. In short, the performance of the model is acceptable, and this builds confidence for further analysis.

2.3 Definition of Monsoon and arid regions

Following Wang et al. (2012), in terms of precipitation characteristics, the global monsoon region (MR) and arid region (AR) are defined as follows: The MR is the area in which the local summer-minus-winter precipitation exceeds 2 mm/day and the local summer precipitation exceeds 55% of annual rainfall. The AR, meanwhile, is defined as the area in which local summer precipitation is less than 1 mm/day. Here, the local
summer and winter are defined as May–September (MJJAS) and November–March (NDJFM) for the NH, and vice versa in the SH. Note that the monsoon regions and arid regions are all located between 60°S and 60°N, and we therefore focus on this range (Fig. 2). Besides the Asian monsoon region, including the tropical, subtropical and extratropical regions, five other monsoon regions are centered in the tropics. The arid regions in the NH contain the tropical and subtropical regions; while in the SH, they are positioned only in the tropics. We survey the climatological (30-yr mean) response to dynamic vegetation for the whole year, and during boreal summer (MJJAS) and boreal winter (NDJFM). Student’s $t$-test is employed to test the significance of the dynamic vegetation-induced precipitation and surface temperature differences.

3. Dynamic vegetation-induced climate variabilities

3.1 Precipitation

3.1.1 Responses in monsoon regions

As shown in the Fig. 3 (left column), it can be seen that the presence of dynamic vegetation increases the precipitation over the Asian monsoon region to a greater extent than in the other monsoon regions, both in winter and summer, and the area with increased precipitation extends farther north in winter. Amongst the arid regions, precipitation synchronously decreases in the whole region of the South Atlantic, while spatial heterogeneity in the precipitation changes is found in the other four arid regions. Given that each of the six monsoon regions and five arid regions is unique, we attempt to reveal to what extent the presence of dynamic vegetation could impact upon the climate in each region. The precipitation differences between the DV and CTL runs are
computed and marked on Fig. 3. In terms of the annual and seasonal precipitation over
the monsoon regions, the area-average impact of dynamic vegetation ranges from −0.28
to 0.08 mm/day, and the effect in the arid regions is comparatively small, with a range
from −0.05 to 0.10 mm/day.

For the monsoon regions, except the Asian monsoon region and boreal summer
North American and Australian monsoon region, annually and seasonally precipitation
are reduced in other monsoon regions. The amplitudes of precipitation change in boreal
winter are larger than in boreal summer, with the exception of the North African
monsoon region. Among the six monsoon regions, the responses of the North African
(0.27 mm/day) and the Australian monsoon region (0.28 mm/day) in local summer are
the most evident.

3.1.2 Responses in arid regions

The arid regions show greater inter-regional differences with respect to the
precipitation response to dynamic vegetation. Decreased precipitation is evident in the
largest arid region, which spans most of Eurasia, and the arid region in southern Pacific,
and the reductions are less in local summer than in local winter. The two American arid
regions located in the eastern Pacific feature an obvious increase in precipitation in
local summer and a decrease in local winter. The only arid region with enhanced
precipitation in summer and winter is situated over the Southwestern Australian, and
the simulated change is smaller in local summer than in local winter. The phases of
annual mean precipitation deviation are consistent with the changes in local summer in
each arid region. The maximum simulated precipitation deviation (0.1 mm/day) occurs
in North American arid region in boreal summer.

3.1.3 Intercomparison between hemispheric monsoon and arid regions

From the global and hemispheric mean perspective, we calculate the absolute differences in the precipitation response between the DV and CTL runs (Table 1). The responses of the precipitation to dynamic vegetation range from −0.131 to 0.024 mm/day. The amplitudes of the absolute precipitation change are larger in the monsoon regions than in the arid regions, except in the NH in boreal summer; combined with Fig. 3e, the enhanced summer monsoon precipitation in the NH can be mostly attributed to the precipitation increasing over land in the Asian monsoon region. We also observe that the amplitudes in the SH are greater than in the NH. The impact of dynamic vegetation leads to increased annual mean precipitation in the NH but decreased that in the SH, in both the monsoon and arid regions. The global changes are in line with those in the SH due to the larger amplitude of reduction in the SH. In boreal winter, the precipitation response in the SH monsoon region has the largest amplitude (−0.131 mm/day), and the reduced global mean precipitation is still reconcilable with that in the SH in the monsoon and arid regions. In boreal summer, the enhanced global precipitation appears to agree with the augmented change in the NH. In short, the impact of dynamic vegetation on precipitation is one of an increase (decrease) in the NH (SH) monsoon and arid regions, with a larger amplitude in the SH than in the NH, as well as in the monsoon regions than in the arid regions.

Further, we analyze the precipitation differences (percentage) caused by dynamic vegetation compared to the CTL run (Table 2). The percentage changes in precipitation
caused by dynamic vegetation range from −2.20% to 3.03%. In terms of the annual mean, dynamic vegetation results in a 0.23% and 0.54% (−1.33% and −1.53%) increment in the monsoon and arid regions in the NH (SH), respectively. The relative changes in the SH are generally larger than those in the NH, with the exception of the arid regions in boreal summer. The relative changes are larger in the monsoon (arid) regions in boreal winter (summer), and the annual changes are dominated by the variability in boreal summer on the hemispheric scale. It is also noticeable that the precipitation responses in the global monsoon (arid) regions show the regions with more (less) precipitation having bigger (smaller) changes, indicating that wetter regions tend to show a larger precipitation response to dynamic vegetation than drier regions. Meanwhile, there are no clear indications of such patterns on the hemispheric scale.

3.2 Surface temperature

3.2.1 Responses in monsoon regions

Different from the precipitation changes, the surface temperature change caused by the presence of dynamic vegetation show better regional consistency (Fig.3, right column). It is evident that the cooling effect dominates most areas, while the Asian–Australian monsoon region and some scattered areas in South America and North America stand out as domains where dynamic vegetation enhances the surface temperature warming.

The differences in monsoon regions range from −0.49°C to 0.23°C. The surface temperature in the Asian monsoon region increases by 0.23°C in boreal winter,
significant at the 99% confidence level, and the enhancements in the South American monsoon region are quite limited and not statistically significant. The decrements of surface temperature (close to 0.5°C) in the North African monsoon region are most evident in the monsoon regions, and are all statistically significant at the 99% confidence level. The surface temperature responses of the monsoon regions in the NH are larger in amplitude than those in the SH.

3.2.2 Responses in arid regions

The arid regions become almost consistently cooler under the impact of dynamic vegetation, with differences ranging from −0.35°C to 0.01°C, the only exception is the Eurasian arid region in boreal winter (0.01°C), and the maximum amplitude of surface temperature change also occurs in the Eurasian arid region in boreal summer (−0.35°C).

The simulated change in Southwestern Australian arid region is smaller in local summer (−0.06°C) than in local winter (−0.33°C). The dynamic vegetation-induced annual and boreal summer mean surface temperature decreased statistically significant at the 99% confidence level in Eurasian, South Atlantic and Southwestern Australian arid region, while responses in the two American arid regions are statistically insignificant. In boreal winter, only the Northern American and South Atlantic arid region show statistically significant differences. Notably, the annual and seasonal anomalies with statistically significant at the 99% confidence level are only found in the South Atlantic arid region, and the dynamic vegetation caused annual (local summer/local winter) mean surface temperature decreasing by about 0.26°C (0.18°C/0.33°C).
3.2.3 Intercomparison between hemispheric monsoon and arid regions

Similarly, the area-average surface temperature differences are computed separately for the monsoon and arid regions (Table 1). The surface temperature differences range from $-0.23$ to $0.05^\circ C$, and the most significant changes occur in boreal summer over the arid regions. The surface temperature responses resulting from dynamic vegetation show a consistent cooling in the NH, SH and GH, with the exception of the NH monsoon region in boreal winter, with an average cooling of $0.09^\circ C$ and $0.16^\circ C$ for the global annual mean in the monsoon and arid regions, respectively. The decrements of surface temperature changes are larger in amplitude in the arid regions than in the monsoon regions, implying that the surface temperature in arid regions is more susceptible, and the change in surface temperature is less remarkable in boreal winter than in boreal summer. Notably, climatological responses with statistically significant at the 95% confidence level are found in more regions for surface temperature than they are for precipitation, indicating that surface temperature is more sensitive to the presence of dynamic vegetation than precipitation.

3.3 Potential mechanism

Dynamic vegetation could influence the circulation and the surface radiation and impact the change of precipitation and surface temperature. In the following, we analyze the dynamic vegetation-induced regional circulation and surface radiative forcing changes and their climate effects.

3.3.1 Impact of regional circulation on precipitation

Moist flux is a significant component in the precipitation cycle, and the
precipitation amount is closely related to the vertically integrated moisture flux convergence. The water vapor flux is larger in monsoon regions than that in arid regions, and the vertically integrated moisture flux convergent (divergent) in local summer (winter) in monsoon regions, and divergent in arid regions usually (not shown). To assess how the dynamic vegetation-induced precipitation was affected by regional circulation, we calculate the differences of the vertically integrated moisture flux divergence/convergence from the surface to top of troposphere (300hPa) between the DV and CTL runs (Fig. 4).

Generally, the moist flux is suppressed (promoted) in the region with decreased (increased) precipitation under the dynamic vegetation effect in most regions. The convergence anomalies are evident in the Asian monsoon and the South African monsoon region, and divergence anomalies occur more frequent in other four monsoon regions and all the arid regions. It is reasonable to speculate that an increased water vapor transfer to Asian monsoon region and a promotion of convergence leads to the instability of the atmosphere, thus increased the precipitation.

In the divergence/convergence term, it shows inconsistent change with the precipitation in the South African monsoon region, boreal summer North American and South American monsoon region, boreal winter North African monsoon region, annual South American arid region, and boreal summer Australian arid region. As for the combination of specific humidity (not shown), the decreased precipitation in South African monsoon region is attributed to the decrement of specific humidity in the DV run, although the promoted convergence is contradict with the decreased precipitation.
Generally, the inconformity of moist flux divergence/convergence and precipitation changes in monsoon regions can gain reasonable support from specific humidity changes; while the differences of specific humidity also can’t match the increased annual precipitation in the South Pacific arid region.

3.3.2 Impact of radiative forcing on surface temperature

Dynamic vegetation can affect the energy balance of the earth and further induce climate variability. In order to illustrate the linkage between dynamic vegetation and the surface temperature, we calculate the differences of regional average surface total radiative forcing (TRF) in each monsoon and arid regions between the DV and CTL runs (Fig.5).

In monsoon regions, the differences of TRF range from $-1.86 \text{W/m}^2$ to $3.77 \text{W/m}^2$. It is noted that the significant warming effects correspond to the enhanced TRF in Asian monsoon region. The enhanced TRF lead to the increased surface temperature in South American monsoon region, while reduced TRF differences coincide well with the decreased precipitation in North American and Australian monsoon region. In arid regions, the TRF differences range from $-1.91 \text{W/m}^2$ to $0.99 \text{W/m}^2$. The main regimes of dynamic vegetation-induced TRF in arid regions are weakened, and hence cooling the area. The synchronous changes between TRF and surface temperature response are found in most monsoon and arid regions, except in the African monsoon region and the Eurasian arid region. We further check the distribution of the TRF differences, and found that the asynchronous changes between the TRF and surface temperature fall in African and the inner Eurasian, and the climate responses to dynamic vegetation are
more complicated in these regions.

Leaf area index (LAI), a key parameter of vegetation dynamics, characterizes canopy density. Since surface radiative forcing can be affected by LAI, causing surface temperature difference between the DV and CTL run, the variations of LAI are investigated (Fig. 6). The results show that LAI decreases in most land areas due to dynamic vegetation, while LAI increases in some Equatorial regions, North Africa, Inner Asia, South Asia and North America. The LAI difference ranges from −0.61 to −0.12 and −0.21 to −0.02, over monsoon and arid region, respectively (Table 3). In terms of the dynamic vegetation-induced annual and seasonal LAI, the area-average LAI ranges from −0.63 to 0.45 over the monsoon regions, while the LAI in the arid regions is relatively small, ranging from −0.43 to 0.02 (Fig. 6). In a word, the LAI decreases more evidently in monsoon regions than that in arid regions.

To some extent, there is a positive correlation between the change of LAI and surface temperature. Combined with Fig. 5 and Fig 3, in the north land of 50°N over East Asian monsoon region, and Sahara and middle Asia arid region, the decreased LAI (Fig.6c) leads to more albedo and less absorption of downward shortwave, resulting in decreased TRF and the corresponding cooling variation in these regions in boreal summer. In the middle land of the East Asian monsoon region, the boreal summer surface temperature and TRF increased while the LAI decreased, thus it shows warming effect which is likely attributable to a decrease in evapotranspiration as LAI and precipitation decrease, leading to an increase of the surface temperature. In boreal summer India and Southeast Asia monsoon region and Arabia arid region and boreal
winter North African monsoon region, the increased LAI encounters the decreased surface temperature and increased TRF, and the possible reason of the cooling effect is that the increased LAI not only absorbs more downward shortwave but also leads to more evapotranspiration, and the cooling effect of evapotranspiration may be stronger than the TRF warming, resulting in cooling change in these regions. Such in-phase or out-phase variations between LAI and surface temperature in these regions were similar with the previous study (Zhang et al. 2002). The cause of different effects of LAI on surface temperature mentioned above may be due to the variation in vegetation types which are represented by plant functional types (PFTs) in DGVM, and the responses of PFTs to radiative forcing are different (Qiu et al. 2017).

In addition, as shown in Fig. 6, each monsoon and arid region to some extent contains sea area, whereas the dynamic vegetation-induced LAI differences occur only on land, and associated climate variability is global. This furthermore illustrates the diversity of dynamic vegetation effects on different regions and highlights the importance of dynamic vegetation in global climate system.

4 East Asian monsoon changes

As mentioned above, when comparing the regional results in Asian monsoon region and other regions, we can conclude that the climate variabilities in the Asian monsoon region are obviously different from other regions, not only with respect to precipitation enhancement, but the surface temperature increment also. Here, we focus on East Asian monsoon to investigate how the dynamic vegetation affects regional atmospheric circulation and influences monsoonal climate variability.
4.1 East Asian summer monsoon and precipitation

Precipitation is one of the dominant features of the monsoon climate, and the summer precipitation in East Asia is largely attributed to the change of the East Asian summer monsoon (EASM) (Wang 2006). Fig. 7a presents the change of precipitation response to dynamic vegetation, from which we can find that the precipitation change has a relatively clear contrast between approximately north and south of 30°N, and the precipitation is enhanced in the Northern China and is suppressed in the Yangtze River Basin. It is noted that the precipitation change induced by dynamic vegetation (Fig. 7a) is in agreement with the observed precipitation pattern difference between strong and weak monsoon years in boreal summer East Asia (Fig. 7b).

The EASM is a subtropical monsoon, the low-level winds reverse primarily from southerlies. To explore how the dynamic vegetation influence the circulation, we examine the change in the summer 850hPa winds and sea level pressure (Fig. 8). It is clear that the dynamic vegetation results in an augment of the east-west pressure gradient, and the surface temperature difference (Fig. 3f) also reveal an increasing trend of the continental East Asian surface temperature, leading to an augment of the land-sea thermal contrast. The LAI almost decreased in all of East Asia (Fig. 6c) which may leads to the reduction in evapotranspiration and the surface temperature rising. The lower-level southerly winds are strengthened over East Asia, thus bring more water vapor from the ocean into East China. In a word, the enhanced summer precipitation in East Asia is associated with the intensification of the EASM circulation induced by the dynamic vegetation.
4.2 East Asian winter monsoon and surface temperature

The East Asian winter monsoon (EAWM) is characterized by northerly winds in the low troposphere and bring cold air from Siberian to the East Asia (Gong et al., 2014). As seen in the Fig. 3d and Fig. 6b, the dynamic vegetation-induced surface temperature shows a significant increase in East Asia; spatially, LAI is increased in north and decreased in most of south region of East Asia, and the whole East Asia exhibits warming trend. In the north, the increased LAI leads to more TRF warming while in the south the warming can be ascribed to the reduction in evapotranspiration which is caused by decreased LAI. In order to explore how the dynamic vegetation influences the EAWM, we also examine the change in the 850hPa winds and sea level pressure (Fig. 9). The decreases in the east-west pressure gradient and the lower-level northerly winds reveal the weakening of EAWM. The weakening of EAWM restrained the cold air southward, thus the surface temperature increased in East Asia. In addition, the increasing of land surface temperature may enhance the instability of the lower atmosphere and promote air convection, thereby lead to an augment of the winter precipitation in East Asian monsoon region.

5. Discussion and conclusion

In this study, with six monsoon regions and five arid regions characterized by precipitation variability, we detect the climate variabilities in global monsoon and arid regions in response to dynamic vegetation in two simulations using the NCAR’s CESM-DGVM. The CESM-DGVM model captures the main spatial feature of the precipitation and surface temperature, and the performance of the model is acceptable.
There is much discrepancy in the dynamic vegetation–induced regional climate variabilities, and the surface temperature responses are more statistically robust and consistent among the monsoon and arid regions than the precipitation. It is concluded that the dynamic vegetation-induced annual precipitation is mainly increased in the NH and decreased in the SH, in both monsoon and arid regions, but with a larger amplitude in the SH compared to the NH. Furthermore, the absolute amplitude in the monsoon regions is larger than in the arid regions. Seasonally, in local summer, the precipitation in monsoon and arid regions universally shows an increasing trend in the NH, while the responses in the SH decreased. In local winter, the precipitation increases in monsoon regions and decreases in arid regions, in both the NH and SH. The precipitation anomalies in monsoon regions in local summer might suggest that the dynamic vegetation enhances the summer monsoon precipitation in the NH, but restrains it in the SH. The responses of area-average surface temperature at the hemispheric scale show a consistent reduction, and the decrement of surface temperature changes are larger in amplitude in the arid regions than those in the monsoon regions. The change of surface temperature is less remarkable in boreal winter than in boreal summer. On the whole, the precipitation responses in monsoon regions are stronger than that in arid regions, while the surface temperature response is opposite. This might indicate that precipitation in monsoon regions is more sensitive to dynamic vegetation, while in arid regions the surface temperature is more susceptible. In addition, the seasonal climate differences may be associated with the stronger land–atmosphere interaction induced by dynamic vegetation in the vegetation growing season, especially in boreal summer.
Regionally, with the dynamic vegetation effect considered, the precipitation in the six monsoon regions generally undergoes an annual and seasonal reduction, except in East Asian monsoon region. The local summer precipitation change in the monsoon regions confirms the impact of dynamic vegetation in terms of strengthening the summer monsoon precipitation in the NH while restraining it in the SH. The response of precipitation to dynamic vegetation in South American monsoon region is consistent with the results reported by Sun and Wang (2014), while there are differences in other regions. This may be due to the varied SST used in their studies, while the SST in this study is fixed. The surface temperature responses of the monsoon (arid) regions in the NH are larger (smaller) in amplitude than those in the SH. Our results, in terms of the positive feedback on surface temperature in East Asian monsoon region, are consistent with the results reported by Notaro and Liu (2008).

The dynamic vegetation can alter moisture flux convergence and surface total radiative forcing, and is critical in precipitation and surface temperature responses. Results from the differences of moist flux indicate that the convergence anomalies are evident in the Asian monsoon and the South African monsoon region, and divergence anomalies occur more frequent in other four monsoon regions and all the arid regions. The dynamic vegetation-induced specific humidity and the vertically integrated moist flux divergence/convergence can attribute to the precipitation changes with the effect of dynamic vegetation in each monsoon and arid regions. Comparing the dynamic vegetation-induced surface total radiative forcing and surface temperature changes, it is noted that the surface total radiative forcing affects most of the monsoon and arid
regions, most notably for the warming effect in Asian monsoon region. There are also some inconsistent changes between the surface total radiative forcing and surface temperature, and it is found that most of the synchronous changes fall in the arid regions, while asynchronous changes are more evident in monsoon regions, mainly in boreal winter South African monsoon region. The decreased surface temperature can be partly explained by the cooling effect of LAI reductions which absorbs less downward shortwave, however the cooling changes conflict with the warming effect of SRF increasing. Such uncertainties indicate the different regional responses of surface temperature to LAI and reveal the complexity of dynamic vegetation associated with LAI change and the corresponding climate variability. In order to obtain more precise regional climate responses to the dynamic vegetation, the regional climate model should be used to conduct further investigation.

The findings of this study are also of interest in terms of the dynamic vegetation-induced large-scale circulation in East Asian monsoon region. The spatial pattern of summer precipitation and winter surface temperature response to dynamic vegetation are similar to the observed regimes of strong EASM and weak EAWM. Further analysis on the 850hPa winds and sea level pressure confirm that the dynamic vegetation can modulate the east-west sea level pressure gradient and the lower-level meridional winds in East Asia, and strengthen EASM and weaken EAWM, respectively.

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List of Figures

1. Comparison of the climatological annual mean precipitation rate (unit: mm/day) and surface temperature (unit: °C) between the (a, c) observation (CMAP, NCEP2, 1979–2008 AD) and (b, d) 30-yr mean of the DV run. The pattern correlation coefficients and root mean square errors with respect to the observation are shown in the upper- and lower-left corners, respectively, of (b, d).

2. Distribution of the global monsoon precipitation region (stippled, blue) and arid region (hatched, brown), based on CMAP data. The labels MN1 to MS3 and AN1 to AS3 indicate the monsoon and arid regions, respectively.

3. Climatological differences in precipitation (left column, unit: mm/day) and surface temperature (right column, unit: °C) between the DV and CTL runs: annual (a, b); boreal winter (c, d); boreal summer (e, f). The areas within the blue and red lines are the monsoon and arid domains, respectively. The numbers on each monsoon and arid regions are the area-average precipitation and surface temperature, the italics bold font and stippling indicate the differences that are statistically significant at the 95% confidence level.

4. Climatological differences in precipitation (black bars) and the vertically integrated moisture flux divergence/convergence from the surface to top of troposphere (300hPa) (red bars) between the DV and CTL runs in each monsoon and arid regions.

5. Same as Fig. 4, but for surface temperature (black bars) and the surface total
radiative forcing (red bars).

Leaf area index differences between the DV and CTL runs: (a) annual, (b) boreal winter and (c) boreal summer. The areas within the blue and red lines are the monsoon and arid domains, respectively. The numbers on each monsoon and arid regions are the area-average leaf area index.

Climatological differences in summer precipitation between the DV and CTL runs (a); Composite strong-minus-weak monsoon year summer rainfall (years chosen as in Yin et al. 2014) (b).

Climatological differences in boreal summer SLP (contours; units: hPa), and 850 hPa winds (vectors; units: m/s) between the DV and CTL runs.

Same as Fig.8, but for the changes in boreal winter.
Fig. 1. Comparison of the climatological annual mean precipitation rate (unit: mm/day) and surface temperature (unit: °C) between the (a, c) observation (CMAP, NCEP2, 1979–2008 AD) and (b, d) 30-yr mean of the DV run. The pattern correlation coefficients and root mean square errors with respect to the observation are shown in the upper- and lower-left corners, respectively, of (b, d).
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Fig. 8. Climatological differences in boreal summer SLP (contours; units: hPa), and 850 hPa winds (vectors; units: m/s) between the DV and CTL runs.
Fig. 9. Same as Fig. 8, but for the changes in boreal winter.
| Table | Description |
|-------|-------------|
| 1 | Precipitation and surface temperature differences in the DV run compared to the CTL run (°C) in Northern Hemispheric (NH), Southern Hemispheric (SH) and global (GLO) monsoon region (MR) and arid region (AR). Single asterisks (*) and double asterisks (**) indicate that the differences are statistically significant at the 95% and 99% confidence levels, respectively. |
| 2 | Precipitation differences (%) in the DV run compared to the CTL run in Northern Hemispheric (NH), Southern Hemispheric (SH) and global (GLO) monsoon region (MR) and arid region (AR). |
| 3 | Leaf area index differences in the DV run compared to the CTL run in Northern Hemispheric (NH), Southern Hemispheric (SH) and global (GLO) monsoon region (MR) and arid region (AR). |
Table 1. Precipitation (mm/day) and surface temperature (°C) differences in the DV run compared to the CTL run in Northern Hemispheric (NH), Southern Hemispheric (SH) and global (GLO) monsoon region (MR) and arid region (AR). Single asterisks (*) and double asterisks (**) indicate that the differences are statistically significant at the 95% and 99% confidence levels, respectively.

|                | Annual | NDJFM | MJJAS |
|----------------|--------|-------|-------|
|                | MR     | AR    | MR    | AR    | MR    | AR    |
| Precipitation  |        |       |       |       |       |       |
| (mm/day)       |        |       |       |       |       |       |
| NH             | 0.009  | 0.005 | 0.014 | −0.002| 0.016 | 0.020 |
| SH             | −0.056 | −0.015| −0.131| −0.012| 0.024 | −0.023|
| GLO            | −0.021 | −0.003| −0.052| −0.006| 0.013 | 0.002 |
| Surface        |        |       |       |       |       |       |
| temperature   |        |       |       |       |       |       |
| (°C)           |        |       |       |       |       |       |
| NH             | −0.09* | −0.13**| 0.05  | −0.05 | −0.20**| −0.23**|
| SH             | −0.09  | −0.20**| −0.09 | −0.15**| −0.10* | −0.23**|
| GLO            | −0.09* | −0.16**| −0.00 | −0.09 | −0.17**| −0.23**|
Table 2. Precipitation differences (%) in the DV run compared to the CTL run in Northern Hemispheric (NH), Southern Hemispheric (SH) and global (GLO) monsoon region (MR) and arid region (AR).

|       | Annual | NDJFM | MJJAS |
|-------|--------|-------|-------|
|       | MR     | AR    | MR    | AR    | MR    | AR    |
| NH    | 0.23   | 0.54  | 0.75  | −0.17 | 0.26  | 3.03  |
| SH    | −1.33  | −1.53 | −1.97 | −1.46 | 1.26  | −2.20 |
| GLO   | −0.51  | −0.37 | −1.30 | −0.63 | 0.31  | 0.20  |
Table 3. Leaf area index differences in the DV run compared to the CTL run in Northern Hemispheric (NH), Southern Hemispheric (SH) and global (GLO) monsoon region (MR) and arid region (AR).

|            | Annual | NDJFM | MJJAS |
|------------|--------|-------|-------|
|            | MR     | AR    | MR    | AR    | MR    | AR    |
| NH         | −0.24  | −0.02 | −0.12 | −0.03 | −0.44 | −0.02 |
| SH         | −0.36  | −0.21 | −0.61 | −0.21 | −0.12 | −0.21 |
| GLO        | −0.30  | −0.05 | −0.33 | −0.05 | −0.31 | −0.04 |