Climate Variability in Monsoon and Arid Regions Attributable to Dynamic Vegetation in a Global Climate Model

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Abstract

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

Keywords 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. About two-thirds of the world’s human population live in monsoon regions, where precipitation imposes significant socio-economic impacts (Wang et al. 2012). Precipitation also plays an important role in arid regions. Because of their fragile ecosystems, arid regions are more sensitive to global change (Rotenberg and Yakir 2010). In recent years, the observed precipitation has increased in monsoon regions and decreased in arid regions under the ever-increasing global warming (Wang et al. 2012).

Vegetation modifies the energy exchange of the land surface–atmosphere system by slowly altering the land surface characteristics and imposing intricate impacts on the terrestrial hydrologic cycle. Thus, vegetation–climate interactions attract 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 the climate. Through analyzing the climate variability resulting from the responses of global land to dynamic vegetation, Sun and Wang (2014) found that dynamic vegetation significantly influences the variation in precipitation 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., the 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 have 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); thus, 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 induce distinct impacts of dynamic vegetation on the climate between global monsoon and arid regions. Based on this assumption, we first describe the model experimental setup, model validation, and 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 hemispheric and regional scales, and we also clarify the possible mechanism. In Section 4, we analyze how dynamic vegetation affects the large-scale circulation in the 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 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) 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. The use of a DGVM coupled with the 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 2016). Further details of the 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 m\(^{-2}\)), CO\(_2\) concentration (367 ppm) and other greenhouse gases, and sea surface temperature (SST) kept constant at the 2000-AD level. One simulation is performed with dynamic vegetation, which varies in time and space and interacts with the climate (DV run), and another simulation is performed with fixed vegetation cover (CTL run). Both simulations are run for 60 years after spin-up, and the last 30 years are analyzed. Using the 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 atmosphere. Hence, based on the model outputs, we can evaluate the effects of dynamic vegetation on the 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 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, indicating that the simulated precipitation in the DV run is in good agreement with the observed climate, with a correlation coefficient of 0.84 (Fig. 1b). The model excellently represents the pattern of surface temperature, and the correlation coefficient between the simulated and observed surface temperature reaches 0.98 (Fig. 1d). For the 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 climatologies 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.

Fig. 1. Comparison of the climatological annual mean precipitation rate (unit: mm day\(^{-1}\)) 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, respectively. 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).
2.3 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\(^{-1}\) 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\(^{-1}\). Here, the local summer and winter are defined as May–September (MJJAS) and November–March (NDJFM) for the NH, respectively, and vice versa in the SH. Note that the MRs and ARs are all located between 60°S and 60°N, and we therefore focus on this range (Fig. 2). In addition to the Asian MR, including the tropical, subtropical, and extratropical regions, five other MRs are centered in the tropics. The ARs 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). A 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

a. Responses in MRs

As shown in Fig. 3 (left column), it can be seen that the presence of dynamic vegetation increases the precipitation over the Asian MR to a greater extent than in the other MRs, both in winter and summer, and the area with increased precipitation extends farther north in winter. Among the ARs, the precipitation synchronously decreases in the whole South Atlantic region, while spatial heterogeneity with the change in precipitation is found in the other four ARs. Given that each of the six MRs and five ARs is unique, we attempt to reveal to what extent the presence of dynamic vegetation could impact 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 MRs, the area-average impact of dynamic vegetation ranges from −0.28 to 0.08 mm day\(^{-1}\), and the effect in the ARs is comparatively small, with a range from −0.05 to 0.10 mm day\(^{-1}\).

For the MRs, except for the Asian MR and boreal summer North American and Australian MRs, annual and seasonal precipitations are reduced in the other MRs. The amplitudes of precipitation change in boreal winter are larger than in boreal summer, with the exception of the North African MR. Among the six MRs, the responses of the North African (0.27 mm day\(^{-1}\)) and Australian MRs (0.28 mm day\(^{-1}\)) in local summer are the most evident.

b. Responses in ARs

The ARs exhibit greater inter-regional differences with respect to the precipitation response to dynamic vegetation. Decreased precipitation is evident in the largest AR, which spans most of Eurasia and the AR in the southern Pacific, and the reductions are smaller in local summer than in local winter. The two
American ARs located in the eastern Pacific feature a clear increase in precipitation in local summer and a decrease in local winter. The only AR with enhanced precipitation in summer and winter is situated over the southwest Australia region, and the simulated change is smaller in local summer than in local winter. The phases of the annual mean precipitation deviation are consistent with the changes in local summer in each AR. The maximum simulated precipitation deviation (0.1 mm day$^{-1}$) occurs in the North American AR in boreal summer.

c. Intercomparison between hemispheric MRs and ARs

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$^{-1}$. The amplitudes of the absolute change in precipitation are larger in the MRs than the ARs, 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 increase over
land in the Asian MR. We also observe that the amplitudes in the SH are greater than in the NH. The impact of dynamic vegetation leads to an increased annual mean precipitation in the NH but a lower annual mean precipitation in the SH, in both the MRs and ARs. The global changes are in line with those in the SH because of the larger amplitude of reduction in the SH. In boreal winter, the precipitation response in the SH MR has the largest amplitude (−0.131 mm day\(^{-1}\)), and the reduced global mean precipitation is still reconcilable with that in the SH in the MRs and ARs. 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 increases (decreases) in the NH (SH) MRs and ARs, with a larger amplitude in the SH than in the NH, as well as in the MRs than in the ARs.

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 MRs and ARs in the NH (SH), respectively. The relative changes in the SH are generally larger than those in the NH, with the exception of the ARs 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 exhibiting larger (smaller) changes, indicating that wetter regions tend to exhibit 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

#### a. Responses in MRs

Unlike the precipitation changes, the surface temperature change caused by the presence of dynamic vegetation shows a better regional consistency (Fig.3, right column). It is evident that the cooling effect dominates most areas, while the Asian–Australian MR 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 MRs range from −0.49°C to 0.23°C. The surface temperature in the Asian MR increases by 0.23°C in boreal winter, significant at the 99 % confidence level, and the enhancements in the South American MR are quite limited and not statistically

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### Table 1. Precipitation (mm day\(^{-1}\)) 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  |        |       |       |       |       |       |
| 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    |        |       |       |       |       |       |
| 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  |
significant. The decrements of surface temperature (close to 0.5°C) in the North African MR are most evident in the MRs and are all statistically significant at the 99% confidence level. The surface temperature responses of the MRs in the NH are larger in amplitude than those in the SH.

b. Responses in ARs

The ARs become almost consistently cooler under the impact of dynamic vegetation with differences ranging from −0.35°C to 0.01°C, with the only exception being in the Eurasian AR in boreal winter (0.01°C), and the maximum amplitude of surface temperature change also occurs in the Eurasian AR in boreal summer (−0.35°C). The simulated change in the Southwestern Australian AR 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 temperatures decrease, statistically significant at the 99% confidence level, in the Eurasian, South Atlantic, and Southwestern Australian ARs, while the responses in the two American ARs are statistically insignificant. In boreal winter, only the North American and South Atlantic ARs exhibit statistically significant differences. Notably, annual and seasonal anomalies with a statistical significance at the 99% confidence level are only found in the South Atlantic AR, and the dynamic vegetation causes an annual (local summer/local winter) mean surface temperature decreasing by about 0.26°C (0.18°C/0.33°C).

c. Intercomparison between hemispheric MRs and ARs

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

3.3 Potential mechanism

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

a. 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 MRs than in ARs, and the vertically integrated moisture flux converges (diverges) in local summer (winter) in MRs and usually diverges in ARs (not shown). To assess how the dynamic vegetation-induced precipitation is affected by regional circulation, we calculate the differences in the vertically integrated moisture flux divergence/convergence from the surface to the top of the troposphere (300 hPa) between the DV and CTL runs (Fig. 4).

Generally, the moist flux is suppressed (amplified) in the region with decreased (increased) precipitation under the dynamic vegetation effect in most regions. Convergence anomalies are evident in the Asian monsoon and South African MRs, while divergence anomalies occur more frequently in the other four regions.

Fig. 4. Climatological differences in precipitation (black bars) and the vertically integrated moisture flux divergence/convergence from the surface to the top of the troposphere (300 hPa) (red bars) between the DV and CTL runs in each monsoon and arid region.
MRs and all the ARs. It is reasonable to speculate that an increased water vapor transfer to the Asian MR and an amplification of convergence lead to an instability of the atmosphere, thus increasing the precipitation.

The divergence/convergence term exhibits an inconsistent change with the precipitation in the South African MR, boreal summer North American and South American MRs, boreal winter North African MR, annual South American AR, and boreal summer Australian AR. As for the combination of specific humidity (not shown), the decreased precipitation in the South African MR is attributed to the decrement of specific humidity in the DV run, although the amplified convergence contradicts the decreased precipitation. Generally, the nonconformity of the moist flux divergence/convergence and the precipitation changes in MRs can gain reasonable support from specific humidity changes, while the differences in specific humidity also cannot match the increased annual precipitation in the South Pacific AR.

b. 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 surface temperature, we calculate the differences in the regional average surface total radiative forcing (TRF) in each MR and AR between the DV and CTL runs (Fig. 5).

In the MRs, the differences in 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 the Asian MR. The enhanced TRF leads to an increased surface temperature in the South American MR, while the reduced TRF differences coincide well with the decreased surface temperature in the North American and Australian MR. In ARs, 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 the ARs are weakened, thus cooling the area. The synchronous changes between the TRF and surface temperature response are found in most MRs and ARs, except in the African MR and Eurasian AR. We further check the distribution of the TRF differences and found that the asynchronous changes between the TRF and surface temperature fall in the African and inner Eurasian regions, and the climate responses to dynamic vegetation are more complicated in these regions.

The leaf area index (LAI), a key parameter of vegetation dynamics, characterizes canopy density. As surface radiative forcing can be affected by the LAI, causing surface temperature differences between the DV and CTL run, the variations in LAI are investigated (Fig. 6). The results indicate that the LAI decreases in most land areas because of dynamic vegetation, while it 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 the MRs and ARs, 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 MRs, while the LAI in the ARs is relatively small, ranging from $-0.43$ to $0.02$ (Fig. 6). In other words, the LAI decreases more evidently in MRs than in ARs.

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

![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 region are the area-average leaf area index.](image)

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

4. East Asian monsoon changes

As mentioned above, from the comparison of the regional results in the Asian MR and other regions, we can conclude that the climate variabilities in the Asian MR are clearly different from those in the other regions, not only with respect to precipitation enhancement but also surface temperature increment. Here, we focus on the East Asian MR 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 in the East Asian summer monsoon (EASM) (Wang 2006). Fig. 7a presents the change in 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 region and suppressed in the Yangtze River Basin. It is noted that the precipitation change induced by dynamic vegetation (Fig. 7a) agrees 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, and the low-level winds reverse primarily from the southerly direction. To explore how the dynamic vegetation influences the circulation, we examine the change in the summer 850-hPa winds and sea-level pressure (Fig. 8). It is clear that the dynamic vegetation results in an increase in the east–west pressure gradient, and the surface temperature difference (Fig. 3f) also reveals an increasing trend of continental East Asian surface temperature, leading to an increase in the land–sea thermal contrast. The LAI almost decreases in all of East Asia (Fig. 6c), which may lead to a reduction in evapotranspiration and rise in surface temperature. The lower-level southerly winds are strengthened over East Asia, thus bringing more water vapor from the ocean to East China. In other words, 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 brings cold air from Siberia to the East Asian region (Gong et al. 2014). As seen in Figs. 3d and 6b, the dynamic vegetation-induced surface temperature shows a significant increase in East Asia; spatially, the LAI increases in the north and decreases in most
of the south region of East Asia, and the whole of East Asia exhibits a 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 a lower LAI. In order to explore how the dynamic vegetation influences the EAWM, we also examine the change in the 850-hPa 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 the EAWM. The weakening of the EAWM restrains the cold air southward, thus increasing the surface temperature in East Asia. In addition, the increasing land surface temperature may enhance the instability of the lower atmosphere and amplify air convection, thereby leading to an increase in the winter precipitation in the East Asian MR.

5. Discussion and conclusion

In this study, with six MRs and five ARs characterized by precipitation variability, we detect the climate variabilities in global MRs and ARs 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 a considerable amount of discrepancy in the dynamic vegetation-induced regional climate variabilities, and the surface temperature responses are more statistically robust and consistent among the MRs and ARs 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 MRs and ARs, but with a larger amplitude in the SH compared to the NH. Furthermore, the absolute amplitude in the MRs is larger than in the ARs. Seasonally, in local summer, the precipitation in the MRs and ARs universally shows an increasing trend in the NH, while the responses in the SH decrease. In local winter, the precipitation increases in the MRs and decreases in the ARs in both the NH and SH. The precipitation anomalies in the MRs 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 the area-average surface temperature at the hemispheric scale show a consistent reduction, and the decrement of the change in surface temperature is larger in amplitude in the ARs than in the MRs. The change in surface temperature is less remarkable in boreal winter than in boreal summer. Overall, the precipitation responses in the MRs are stronger than those in the ARs, while the surface temperature response is opposite. This might indicate that precipitation in MRs is more sensitive to dynamic vegetation, while in ARs, the surface temperature is more susceptible. In addition, the seasonal climate differences may be associated with a 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 MRs generally undergoes an annual and seasonal reduction, except in the East Asian MR. The local summer precipitation change in the MRs confirms the impact of dynamic
vegetation in terms of the strengthening of the summer monsoon precipitation in the NH and its restraint in the SH. The response of precipitation to dynamic vegetation in the South American MR is consistent with the results reported by Sun and Wang (2014), while there are differences in the other regions. This may be due to the varied SST used in their studies, while the SST used 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 the East Asian MR, are consistent with the results reported by Notaro and Liu (2008).

Dynamic vegetation can alter the moisture flux convergence and surface TRF and is critical in precipitation and surface temperature responses. Results from the differences in moist flux indicate that convergence anomalies are evident in the Asian MR and South African MR, and divergence anomalies occur more frequently in the other four MRs and all the ARs. The dynamic vegetation-induced specific humidity and the vertically integrated moist flux divergence/convergence can be attributed to the precipitation changes with the effect of dynamic vegetation in each MR and AR. Comparing the dynamic vegetation-induced surface TRF and surface temperature changes, it is noted that the surface TRF affects most of the MRs and ARs, most notably for the warming effect in the Asian MR. There are also some inconsistent changes between the surface TRF and surface temperature, and it is found that most of the synchronous changes fall within the ARs, while asynchronous changes are more evident in the MRs, mainly in the boreal winter South African MR. The decreased surface temperature can be partly explained by the cooling effect of the LAI reductions, which absorbs less downward shortwave radiation; however, the cooling changes conflict with the warming effect of the increasing SRF. Such uncertainties indicate the different regional responses of surface temperature to the LAI and reveal the complexity of dynamic vegetation associated with the 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 the East Asian MR. The spatial pattern of the summer precipitation and winter surface temperature response to dynamic vegetation are similar to the observed regimes of the strong EASM and weak EAWM. Further analysis of the 850-hPa winds and sea-level pressure confirm that dynamic vegetation can modulate the east–west sea-level pressure gradient and lower-level meridional winds in East Asia, and it can strengthen the EASM and weaken the EAWM, respectively.

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