Vegetation Feedbacks to Climate in the Global Monsoon Regions*

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(Manuscript received 1 December 2010, in final form 30 March 2011)

ABSTRACT
Vegetation feedbacks on climate, on the subannual time scale, are examined across six monsoon regions with a fully coupled atmosphere-ocean-ice-land model with dynamic vegetation. Initial value ensemble experiments are run in which the total vegetation cover fraction across the six monsoon regions is reduced and the climatic response assessed. Consistent responses among the regions include reductions in leaf area index, turbulent fluxes, and atmospheric moisture; enhanced subsidence; and increases in ground and surface air temperature. The most distinct changes in vertical motion, precipitable water, and precipitation occur along the flanks of the monsoon season, with small changes in midmonsoon rainfall. Unique responses to reduced vegetation cover are noted among the monsoon regions. While the monsoon is delayed and weaker over north Australia owing to diminished leaf area, it occurs earlier over China and the southwest United States. The subtropical monsoon regions are characterized by a larger decrease in sensible heat than latent heat flux, while the opposite is true for tropical monsoon regions. North Australia experiences the most substantial decline in both moisture flux convergence and precipitation.

1. Introduction
Vegetation impacts the atmosphere by modifying the partitioning of evapotranspiration, roughness length, Bowen ratio, surface albedo, emissivity, and carbon fluxes (McPherson 2007). The sign and strength of vegetation feedbacks depend on the geographical region and background climatology. Deforestation might induce cooling at the high latitudes, due to an albedo feedback (Betts 2000; Govindasamy et al. 2001), and warming in the tropics, due to an evapotranspiration feedback (Costa and Foley 2000; Gibbard et al. 2005). This is largely attributed to the efficient partitioning of heat into evapotranspiration at low latitudes related to the exponential relationship between temperature and saturation vapor pressure (Gibbard et al. 2005). Hoffmann and Jackson (2000) examined the climatic impacts of converting five savanna regions into grasslands in a coupled climate model. The imposed land use change caused both robust responses, such as an increase in surface air temperature and decline in evapotranspiration in all regions, and unique responses, such as a decline in precipitation in four regions and no change in the North African savanna. Zheng and Eltahir (1998) found that simulated desertification along the southern edge of the Sahara produced minimal impact on the African monsoon while simulated coastal deforestation triggered a collapse of the African monsoon circulation owing to reduced boundary layer entropy. Simulations by Osborne et al. (2004) suggested that vegetation may influence the Southeast Asian monsoon precipitation but has minimal impact on the intense South Asian monsoon system, indicating differential vegetation feedbacks on monsoon systems.

Individual biomes, such as forests versus grasslands, produce unique feedbacks to climate (Snyder et al. 2004; Gibbard et al. 2005; Liu et al. 2010; Notaro and Gutzler 2011). By individually removing global biomes in a coupled climate model, Snyder et al. (2004) found that the boreal forests have the greatest influence on global temperatures, savannas have the greatest local influence on precipitation, and tropical forests induce the largest remote impacts on precipitation.

Modeling studies of tropical deforestation have generally shown local surface warming and diminished evapotranspiration and precipitation (Eltahir and Bras
1993; Lean and Rowntree 1997; Hahmann and Dickinson 1997; Dickinson and Henderson-Sellers 1988; Costa and Foley 2000; Osborne et al. 2004; Nosetto et al. 2005). Local land surface changes can impact the large-scale atmospheric circulation pattern and trigger remote climatic responses. Hahmann and Dickinson (1997) found that most modeling studies of tropical deforestation simulate a decrease in moisture convergence. The spatial extent of atmospheric subsidence anomalies can extend beyond the deforested area, as shown by Delire et al. (2001) for Indonesia. These local and remote responses are not uniform across all modeling studies, however. Polcher and Laval (1994) actually simulated an increase in moisture convergence and precipitation in response to Amazonian and tropical African deforestation. Deforestation-related reductions in roughness length can lead to either decreases (Pitman et al. 1993; Hahmann and Dickinson 1997) or increases in precipitation (Zeng et al. 1996; Lean and Rowntree 1993).

Few studies have focused on, and compared, vegetation feedbacks on climate in the global monsoon regions. Lee et al. (2010, manuscript submitted to J. Geophys. Res.) applied observational data and the Predicting Ecosystem Goods and Services Using Scenarios (PEGASUS) ecosystem model to assess the role of turbulent heat fluxes in the monsoons of North Africa, India, and China. They concluded that sensible heat fluxes affect monsoon onset, while latent heat fluxes affect the monsoon intensity during the wet season. The conversion of potential natural vegetation to bare ground in these regions leads to diminished summertime latent heat fluxes and a weaker monsoon (Lee et al. 2010, manuscript submitted to J. Geophys. Res.). Hoffmann and Jackson (2000) simulated the effects of converting savanna to grassland over north Australia, North Africa, and South America in the National Center for Atmospheric Research (NCAR) Community Climate System Model, version 3–Land Surface Model (CCSM3-LSM) and found the largest reductions in moisture convergence and precipitation over north Australia and minimal precipitation response over North Africa. Both sensible and latent heat fluxes are reduced, particularly the latter.

Studies on vegetation feedbacks over North Africa during the mid-Holocene reached opposing conclusions of both positive (Kutzbach et al. 1996; Braconnot et al. 1999) and negative (Wang et al. 2007; Notaro et al. 2008) feedbacks on precipitation. In response to the removal of tropical Asian vegetation, Mabuchi et al. (2005) simulated an increase in latent heat flux due to greater bare soil evaporation, which is a similar unexpected result as with Notaro et al. (2008). Using a regional climate model, Matsui et al. (2005) found that higher vegetation cover did not consistently increase the evaporative fraction of the North American monsoon system, with land–atmosphere exchanges highly dependent on the canopy stress index parameterization of the land surface model. Miller et al. (2005) suggested that the observed failure of the Australian monsoon to penetrate the continental interior during the Holocene was the result of burning by early human civilization, which converted the tree–shrub–grassland mosaic across the semiarid zone into desert shrub and triggered desertification. By replacing the Australian desert with forest, they simulated enhanced moisture recycling, decreased albedo, increased roughness, greater surface heating, and greater convection over the continental interior, which intensified onshore flow.

Rather than focus on the equilibrium, long-term response of climate to a permanent, imposed change in vegetation cover (e.g., greening in the Sahara or Amazonian deforestation), the present study focuses on the short-term, intra-annual to interannual feedback to climate (Notaro and Liu 2008; Notaro et al. 2008; Notaro and Gutzler 2011) of an imposed variation in vegetation cover. The latter is critical for seasonal to annual climate prediction in the monsoon regions and will likely be more local, with less remote impacts. This modeling study aims to address the following questions. 1) How do variations in vegetation cover impact the global monsoon systems? 2) What are the similarities and differences in vegetation feedbacks among the monsoon regions?

The data and methods are outlined in section 2, results of the modeling experiments are described in section 3, and the discussion and conclusions are presented in section 4.

2. Data and methods

2.1. Model description

The applied model is the NCAR Community Climate System Model Version 3.5 (CCSM3.5) (Collins et al. 2006; Gent et al. 2010) with dynamic atmosphere, ocean, sea ice, and land surface, including interactive vegetation. NCAR CCSM3.5 is applied without use of flux adjustment. The active model components include the Community Atmosphere Model (CAM) with 26 levels in the vertical, Parallel Ocean Program (POP) with 60 levels in the vertical, Community Sea Ice Model (CSIM), and Community Land Model–Dynamic Global Vegetation Model (CLM3.5-DGVM) with 10 soil layers. The CLM3.5-DGVM (Levis et al. 2004) is based on the NCAR Land surface Model (LSM) (Bonan et al. 2003),
with annual processes based on the Lund–Potsdam–Jena (LPJ) DGVM (Sitch et al. 2003). Ten plant functional types (PFTs) are simulated, including seven types of trees and three types of grasses. In CCSM3.5, the CAM spectral transform dynamical core in CCSM3 was replaced with a Lin–Rood finite volume dynamical core (Lin 2004). The standard atmospheric and land resolution for the finite volume (FV) core is applied, with $144 \times 96$ grid cells ($1.9^\circ$ latitude $\times$ $2.5^\circ$ longitude).

Numerous improvements (Gent et al. 2010) were made in CCSM3.5 over its predecessors. The CAM convective scheme represents both shallow convection (Hack 1994) and deep convection (Zhang and McFarlane 1995), and the effects of deep convection are incorporated into the momentum equation (Richter and Rasch 2008). A dilute approximation is applied in plume calculations of convective available potential energy (Raymond and Blythe 1986). These CAM modifications substantially improved the simulated frequency of El Niño–Southern Oscillation events (Deser et al. 2006; Neale et al. 2008), while the spatial pattern of teleconnection responses to ENSO is better represented in version 3.5. Due to these modifications to the convective scheme, the double intertropical convergence zone (ITCZ) bias in the western tropical Pacific was partially alleviated and the excessively strong trade winds were weakened (Gent et al. 2010).

In developing CCSM3.5, several modifications were made to the land component, CLM (Oleson et al. 2008; Stöckli et al. 2008). The partitioning of evapotranspiration (ET) was significantly improved, resulting in wetter soils, reduced water stress on plants, increased transpiration, enhanced photosynthesis, and a better representation of the annual cycle of total water storage (Oleson et al. 2008). A resistance term was added to reduce excessive soil evaporation, while scaling of the canopy interception was included (Lawrence et al. 2007) to alleviate the excessive interception in CLM3 (Hack et al. 2006). A factor was introduced to account for nitrogen limitation on plant growth.

### b. Simulations

A 200-yr modern-day control simulation (CTL) of CCSM3.5 is produced, and the last 80 years are analyzed. Restart files are saved at the end of each year, which include the fractional cover of each plant function type. A set of 80 initial value ensemble experiments (ENS) is run, each one year in duration. At the start of each ensemble member, a restart file is obtained from CTL and the fraction of total vegetation cover is reduced by 0.2 over the six global monsoon regions of North America, South America, North Africa, India, China, and north Australia (Fig. 1). The total cover of all PFTs in a grid cell is reduced, while maintaining the original proportions of PFTs.

The climate in each ensemble member is compared against its relevant year in the control simulation to reveal the impact of reduced vegetation cover on the climate of a monsoon region. A similar approach was previously applied over North Africa and northern Eurasia with the global climate model, Fast Ocean Atmosphere Model (FOAM)–LPJ, by Notaro et al. (2008) and Notaro and Liu (2008), respectively. The large number of ensemble members allows the individual climatic responses of each monsoon region to reduced vegetation cover to emerge from the noise when comparing the climate of ENS and CTL. Results over the North American monsoon region were assessed by Notaro and Gutzler (2011).

### c. Observations and reanalysis

Simulated precipitable water and precipitation are compared with the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al. 1996) and the Climate Prediction Center Merged Analysis of Precipitation dataset (Xie and Arkin 1996), respectively. Simulated vegetation types are compared against the potential natural vegetation dataset of Ramankutty and Foley (1998).
3. Results

a. Mean simulated precipitation and vegetation

The ability of the model’s predecessor, CCSM3 (without dynamic vegetation), and CAM3 at simulating the global monsoon systems was examined by Meehl et al. (2006). Consistent with that study, CCSM3.5-DGVM generally simulates global monsoon systems that are reasonable in seasonal timing. Simulated and observed precipitation peak during July–August in India and China, August–September in North Africa, and February–March in South America. The simulated peaks in precipitation during September–October across North America and February–March across north Australia represent one-month delays compared to observed. The simulated North American monsoon precipitation persists through late autumn over Mexico.

All six regions are characterized by a seasonal peak in precipitable water that is 8%–21% higher than in the NCEP–NCAR reanalysis, most notably over north Australia and North Africa. In conjunction with these high precipitable water biases, the model simulates even greater precipitation biases. The largest precipitation biases are +158% and +136% across North America and north Australia, respectively, while the other four regions have simulated precipitation within 25% of observed annual precipitation. The simulated monsoons generally penetrate too far inland, especially for north Australia.

The noted precipitation biases in the atmospheric model of CCSM3.5 influence the simulated vegetation, as evident by comparing the CTL simulation with an offline, observation-driven CLM3.5-DGVM simulation. Because of a simulated wet bias, more trees and less grasses are produced across North America and north Australia as compared to CTL than offline, while a simulated dry bias over South America causes the opposite vegetation response.

b. Characteristics of the six monsoon regions

Given that each of the six monsoon regions is unique, it is expected that the feedback response to a variation in vegetation cover will differ by region. The monsoons of China and North America are positioned in the subtropics, while the other four monsoons are centered in the tropics. The most extratropical monsoon is over China and the most equatorward is over North Africa (Fig. 1). The regions’ simulated background climatologies vary dramatically. Annual mean temperatures range from 16°C in China to 28°C in South America, and annual mean precipitation varies from 85 cm in North America to 175 cm in north Australia. Only China (18 cm of actual snow depth) and North America (13 cm) receive measurable simulated snowfall.

Vegetation categories are unique among the six monsoon regions (Fig. 2). Simulated mean percent tree cover in CTL ranges from 24% in South America to 87% in China. The simulated natural vegetation cover is densely distributed, with limited bare space ranging from 1% in China and India to 14% in South America. The simulated vegetation types generally agree favorably with the potential natural vegetation dataset of Ramankutty and Foley (1998) for the six regions. The main exception is across North America, where the model simulates a mix of temperate needleleaf evergreen trees and grasses, while the dataset of Ramankutty and Foley (1998) suggests that the primary type should be open shrublands. The model does not include a shrub PFT but can simulate short trees in such a semiarid environment. There is evidence that the natural vegetation type over portions of the North American monsoon region is actually dense grassland and not shrubland (Cornelius et al. 1991; Tweit 1995; Reynolds et al. 2000; Laliberte et al. 2004), making the absence of a shrub category in CCSM3.5 less of a concern.

c. Consistency among regions

First, the study will focus on climatic responses to land cover variations that are consistent among the monsoon regions. To quantify how robust the climatic responses are to reduced vegetation cover across the six regions, a consistency index is introduced as the number of regions with the same sign change in a variable (ENS − CTL). Responses with a consistency index of at least five are highlighted in Table 1.

The responses to reduced vegetation cover are generally quite consistent prior to the summer monsoon season and then the consistency diminished in summer–autumn. Of the 33 variables listed in Table 1, 26 of them display high consistency (>5) in winter but only 15 do so in autumn. Thermal responses to reduced vegetation cover among the monsoon regions tend to be more consistent than hydrological responses.

Consistent annual responses to reduced vegetation cover include decreases in leaf area index (LAI), surface stress, sensible and latent heat fluxes, evapotranspiration, precipitable water, large-scale precipitation, and low cloud fraction and increases in ground and surface air temperature, boundary layer depth, low-level divergence, and subsidence (Table 1). The LAI is reduced year-round on average by −0.79 m² m⁻², with the greatest reduction during the peak growing season in summer, −0.86 m² m⁻². The response in surface albedo is minimal among the regions, reaching −0.004 on average in summer–autumn. Given less vegetation cover, the surface stress is reduced year-round, particularly in autumn when climatological wind speeds are strong. In
response to the decline in LAI, sensible heat flux consistently decreases in winter and latent heat flux consistently decreases in winter–spring, with larger reductions in the latter variable. Given that both turbulent fluxes decrease annually, the Bowen ratio does not consistently change among the regions. The smallest response in turbulent fluxes occurs during the summer monsoons when oceanic forcings swamp those of land forcings.

Modest increases in annual ground, surface air, and vegetation temperatures of 1.28°C, 0.48°C, and 0.34°C, respectively, result from reduced vegetation cover among the monsoon regions. The warming of the ground and air is limited to nighttime and peaks in spring. With the springtime warming and enhanced wind speeds, the atmospheric boundary layer deepens by 14 m on average. Consistent decreases in evapotranspiration, along with reduced moisture convergence, lead to diminished precipitable water during winter–summer and precipitation (mainly convective) during winter–spring. Subsidence and associated low-level divergence are robust responses during winter–summer, particularly in spring. Subsidence and drying support a consistent decrease in low cloud cover during spring–autumn. However, the fractional cover increases and top pressure decreases for summertime convective clouds, indicating deeper, more abundant convective clouds with reduced vegetation cover.

d. Surface radiation budget

The responses of each component in the surface radiation budget to reduced vegetation cover are analyzed, including shortwave radiation, longwave radiation, sensible heat flux, latent heat flux, and soil heat flux (Fig. 3). For all regions, the surface gains energy through decreases in sensible and latent heat fluxes and increases in downward shortwave radiation flux and loses energy through increases in heat flux into the soil layers and upward longwave radiation flux. The change in each component will be discussed in the order of their magnitude. The largest change among the regions is an increase in net surface longwave radiation flux, with modifications to the net flux dominated by its upward component. The increase in ground temperature leads to the emission of more longwave radiation to the atmosphere. The largest longwave response is found across the North American monsoon region where an annual increase in ground temperature of 1.6°C, due to reduced vegetation cover, leads to +2.8 W m⁻² more net surface longwave radiation flux (Fig. 3b2), which is a sink of radiation to the atmosphere. Among the six regions, the North American monsoon region is characterized by the largest increases in net longwave and shortwave radiation and decreases in sensible heat flux.

FIG. 2. Pie charts of the simulated (CTL run) mean percent cover of evergreen trees (blue), deciduous trees (red), grasses (green), and bare ground (brown) for the monsoon regions of (a) China, (b) North America, (c) North Africa, (d) India, (e) north Australia, and (f) South America.
TABLE 1. Mean changes (ENS — CTL) in 33 different variables, averaged among the six monsoon regions, shown annually and by season, with increases and decreases that achieve a consistency index of 5–6 shown with up and down arrows, respectively. For each variable, bold italic font indicates the season with the largest change.

<table>
<thead>
<tr>
<th>Consistency index</th>
<th>Annual</th>
<th>Winter</th>
<th>Spring</th>
<th>Summer</th>
<th>Autumn</th>
</tr>
</thead>
<tbody>
<tr>
<td>Leaf area index (m² m⁻²)</td>
<td>−0.79 ± 0.37 ↓</td>
<td>−0.68 ± 0.36 ↓</td>
<td>−0.80 ± 0.41 ↓</td>
<td>−0.86 ± 0.37 ↓</td>
<td>−0.78 ± 0.39 ↓</td>
</tr>
<tr>
<td>Ground temperature (°C)</td>
<td>1.28 ± 0.22 ↑</td>
<td>1.19 ± 0.34 ↑</td>
<td>1.65 ± 0.45 ↑</td>
<td>1.22 ± 0.48 ↑</td>
<td>1.06 ± 0.24 ↑</td>
</tr>
<tr>
<td>Maximum ground temperature (°C)</td>
<td>−0.44 ± 0.25 ↓</td>
<td>−0.40 ± 0.32 ↓</td>
<td>−0.35 ± 0.40 ↓</td>
<td>−0.43 ± 0.22 ↓</td>
<td>−0.53 ± 0.18 ↓</td>
</tr>
<tr>
<td>Minimum ground temperature (°C)</td>
<td>0.97 ± 0.34 ↑</td>
<td>1.02 ± 0.32 ↑</td>
<td>1.05 ± 0.37 ↑</td>
<td>0.78 ± 0.40 ↑</td>
<td>1.06 ± 0.36 ↑</td>
</tr>
<tr>
<td>Vegetation temperature (°C)</td>
<td>0.34 ± 0.13 ↑</td>
<td>0.40 ± 0.24 ↑</td>
<td>0.31 ± 0.17 ↑</td>
<td>0.21 ± 0.15 ↑</td>
<td>0.44 ± 0.18 ↑</td>
</tr>
<tr>
<td>2-m surface air temperature (°C)</td>
<td>0.48 ± 0.12 ↑</td>
<td>0.51 ± 0.21 ↑</td>
<td>0.52 ± 0.14 ↑</td>
<td>0.12 ± 0.15 ↑</td>
<td>0.50 ± 0.16 ↑</td>
</tr>
<tr>
<td>Maximum 2-m surface air temperature (°C)</td>
<td>−0.10 ± 0.09 ↓</td>
<td>−0.16 ± 0.17 ↓</td>
<td>−0.08 ± 0.16</td>
<td>−0.08 ± 0.09</td>
<td>−0.09 ± 0.08 ↓</td>
</tr>
<tr>
<td>Minimum 2-m surface air temperature (°C)</td>
<td>0.73 ± 0.18 ↑</td>
<td>0.79 ± 0.25 ↑</td>
<td>0.80 ± 0.28 ↑</td>
<td>0.57 ± 0.27 ↑</td>
<td>0.75 ± 0.21 ↑</td>
</tr>
<tr>
<td>Diurnal temperature range (°C)</td>
<td>−0.83 ± 0.17 ↓</td>
<td>−0.95 ± 0.11 ↓</td>
<td>−0.88 ± 0.37 ↓</td>
<td>−0.65 ± 0.32 ↓</td>
<td>−0.83 ± 0.17 ↓</td>
</tr>
<tr>
<td>Sensible heat flux (W m⁻²)</td>
<td>−0.72 ± 0.26 ↓</td>
<td>−1.18 ± 0.42 ↓</td>
<td>−0.90 ± 0.85</td>
<td>−0.21 ± 1.14</td>
<td>−0.60 ± 0.85</td>
</tr>
<tr>
<td>Latent heat flux (W m⁻²)</td>
<td>−0.89 ± 0.53 ↓</td>
<td>−1.82 ± 1.58 ↓</td>
<td>−1.40 ± 0.65</td>
<td>−0.24 ± 0.27</td>
<td>−0.10 ± 0.79</td>
</tr>
<tr>
<td>Bowen ratio</td>
<td>−0.004 ± 0.020</td>
<td>0.014 ± 0.038</td>
<td>−0.013 ± 0.037</td>
<td>−0.021 ± 0.051</td>
<td>0.006 ± 0.040</td>
</tr>
<tr>
<td>Precipitable water (kg m⁻²)</td>
<td>−0.15 ± 0.10 ↓</td>
<td>−0.20 ± 0.29 ↓</td>
<td>−0.25 ± 0.31 ↓</td>
<td>−0.15 ± 0.13 ↓</td>
<td>−0.01 ± 0.46</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>−1.49 ± 1.89</td>
<td>−0.26 ± 0.21</td>
<td>−0.84 ± 1.76</td>
<td>−0.30 ± 0.94</td>
<td>−0.10 ± 1.39</td>
</tr>
<tr>
<td>(cm period⁻¹)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Large-scale precipitation (cm period⁻¹)</td>
<td>−0.72 ± 0.65 ↓</td>
<td>−0.06 ± 0.09 ↓</td>
<td>−0.19 ± 0.49 ↓</td>
<td>−0.25 ± 0.60</td>
<td>−0.22 ± 0.50</td>
</tr>
<tr>
<td>Convective precipitation (cm period⁻¹)</td>
<td>−0.78 ± 1.45</td>
<td>−0.20 ± 0.12</td>
<td>−0.65 ± 1.32</td>
<td>−0.04 ± 0.44</td>
<td>0.12 ± 97</td>
</tr>
<tr>
<td>Net evaporation (cm period⁻¹)</td>
<td>−1.11 ± 0.67 ↓</td>
<td>−0.57 ± 0.49 ↓</td>
<td>−0.44 ± 0.20 ↓</td>
<td>−0.08 ± 0.08</td>
<td>−0.03 ± 0.25</td>
</tr>
<tr>
<td>Precipitation-evaporation (cm period⁻¹)</td>
<td>−0.38 ± 1.98</td>
<td>0.31 ± 0.52</td>
<td>−0.40 ± 1.67</td>
<td>−0.22 ± 0.96</td>
<td>−0.07 ± 1.57</td>
</tr>
<tr>
<td>Column volumetric soil water (mm/m³)</td>
<td>0.0004 ± 0.0007</td>
<td>0.0008 ± 0.0008 ↑</td>
<td>0.0007 ± 0.0013</td>
<td>0.0004 ± 0.0011</td>
<td>−0.0005 ± 0.0016</td>
</tr>
<tr>
<td>Total cloud cover fraction</td>
<td>−0.001 ± 0.003</td>
<td>0.000 ± 0.004</td>
<td>−0.002 ± 0.007</td>
<td>−0.001 ± 0.006</td>
<td>−0.002 ± 0.012</td>
</tr>
<tr>
<td>Low cloud cover fraction</td>
<td>−0.005 ± 0.003</td>
<td>−0.001 ± 0.004</td>
<td>−0.007 ± 0.006</td>
<td>−0.008 ± 0.007</td>
<td>−0.004 ± 0.007 ↓</td>
</tr>
<tr>
<td>Medium cloud cover fraction</td>
<td>0.000 ± 0.002</td>
<td>−0.001 ± 0.003</td>
<td>−0.001 ± 0.006</td>
<td>0.001 ± 0.005</td>
<td>−0.001 ± 0.007</td>
</tr>
<tr>
<td>High cloud cover fraction</td>
<td>0.001 ± 0.004</td>
<td>0.001 ± 0.003</td>
<td>0.000 ± 0.007</td>
<td>0.002 ± 0.006</td>
<td>0.000 ± 0.011</td>
</tr>
<tr>
<td>Convection top pressure</td>
<td>−3.56 ± 3.19</td>
<td>1.45 ± 1.05</td>
<td>−3.15 ± 6.61</td>
<td>−8.81 ± 4.08</td>
<td>−3.71 ± 6.38</td>
</tr>
<tr>
<td>(hPa)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Surface albedo</td>
<td>−0.001 ± 0.002</td>
<td>0.001 ± 0.005</td>
<td>0.000 ± 0.003</td>
<td>−0.004 ± 0.004 ↓</td>
<td>−0.004 ± 0.003 ↓</td>
</tr>
<tr>
<td>Depth of the planetary boundary layer (m)</td>
<td>9.68 ± 3.75 ↑</td>
<td>8.17 ± 9.34 ↑</td>
<td>14.44 ± 8.78 ↑</td>
<td>7.28 ± 5.26 ↑</td>
<td>8.80 ± 5.86 ↑</td>
</tr>
<tr>
<td>700-hPa omega (Pa s⁻¹)</td>
<td>0.0009 ± 0.0006 ↑</td>
<td>0.0007 ± 0.0009 ↑</td>
<td>0.0011 ± 0.0016 ↑</td>
<td>0.0009 ± 0.0007 ↑</td>
<td>0.0007 ± 0.0015</td>
</tr>
<tr>
<td>850-hPa divergence (s⁻¹)</td>
<td>0.56 ± 0.61</td>
<td>0.50 ± 0.82</td>
<td>0.82 ± 0.73</td>
<td>0.53 ± 0.65</td>
<td>0.39 ± 70</td>
</tr>
<tr>
<td>Surface stress (kg m⁻²)</td>
<td>−0.0059 ± 0.0046</td>
<td>−0.0067 ± 0.0059</td>
<td>−0.0032 ± 0.0032</td>
<td>−0.0045 ± 0.0046</td>
<td>−0.0091 ± 0.0079</td>
</tr>
<tr>
<td>Flux of CO₂ to the atmosphere (umol/m²/s)</td>
<td>0.36 ± 0.08 ↑</td>
<td>0.22 ± 0.17 ↑</td>
<td>0.38 ± 0.24 ↑</td>
<td>0.49 ± 0.12 ↑</td>
<td>0.33 ± 0.09 ↑</td>
</tr>
<tr>
<td>TOA upward shortwave flux (W m⁻²)</td>
<td>0.35 ± 0.32</td>
<td>−0.27 ± 0.98</td>
<td>0.28 ± 1.14</td>
<td>0.76 ± 1.30</td>
<td>0.60 ± 1.73</td>
</tr>
<tr>
<td>TOA upward longwave flux (W m⁻²)</td>
<td>0.36 ± 0.54</td>
<td>0.61 ± 0.51 ↑</td>
<td>0.56 ± 1.18</td>
<td>−0.07 ± 0.87</td>
<td>0.35 ± 1.51</td>
</tr>
<tr>
<td>TOA net shortwave + longwave flux (W m⁻²)</td>
<td>0.02 ± 0.48</td>
<td>0.88 ± 1.12 ↑</td>
<td>0.28 ± 0.90</td>
<td>−0.42 ± 1.41</td>
<td>−0.25 ± 72</td>
</tr>
</tbody>
</table>

(Figs. 3a–c). Owing to diminished evapotranspiration, the tropospheric moisture and cloud cover are generally reduced, which increases net shortwave radiation flux to the surface for all regions. Small surface albedo changes also influence the upward component of shortwave radiation.

Latent heat flux decreases annually for all regions, with the greatest reduction of −1.8 W m⁻² across North
FIG. 3. Mean changes (ENS − CTL) in surface radiation budget components due to reduced vegetation cover fraction for (row 1) China, (row 2) North America, (row 3) North Africa, (row 4) India, (row 5) north Australia, and (row 6) South America. Changes in (a) shortwave radiation, (b) longwave radiation, (c) sensible heat flux, (d) latent heat flux, and (e) soil heat flux are shown. The net fluxes are further divided into (a),(b) upward and downward fluxes for shortwave and longwave radiation, (c) ground and vegetation components for sensible heat flux, and (d) ground evaporation, transpiration, and canopy evaporation for latent heat flux. Red dots indicate statistically significant (90%) differences, based on t tests.
Africa. Ground evaporation increases as diminished leaf area exposes more bare soil, with the largest increases during the monsoon season when the soils are climatologically wet; the increase in ground evaporation leads to year-round drying of the topmost soil layer in all regions (not shown). Canopy evaporation decreases with less leaf area to intercept falling precipitation, most notably during the monsoon season when precipitation is climatologically greatest. Wetter regions, such as north Australia, tend to have greater reductions in canopy evaporation than drier regions, like North America. The timing of the greatest increase in ground evaporation and decrease in canopy evaporation is largely climate driven, based on the wet season. However, the decrease in plant transpiration tends to peak during the season of maximum LAI reduction, which varies by region. Regions with the greatest tree cover, such as China, experience the largest decline in transpiration and increase in ground evaporation. All regions undergo an annual reduction in sensible heat flux, with decreases in the sensible heat flux from vegetation exceeding increases in the sensible heat flux from the ground. With the loss of leaf area, heat flux into the soil increases annually about the same amount for all regions, which supports anomalously high nighttime temperatures.

e. Climatic responses by region and their robustness

Throughout this section, a robustness index will be discussed, which, for a specific region, is the percentage of ensemble members with the same sign change in a climatic variable between ENS and CTL.

By reducing the vegetation cover fraction for each monsoon region, the LAI and stem area index (SAI) are decreased year-round, particularly in regions with substantial evergreen tree cover such as North Africa, China, and North America. Decreases in LAI range from $-1.25 \text{ m}^2 \text{ m}^{-2}$ in North Africa and $-1.18 \text{ m}^2 \text{ m}^{-2}$ in China to $-0.31 \text{ m}^2 \text{ m}^{-2}$ in South America, which is largely grassland (Figs. 4a,i,u).

Reduced LAI supports a year-round rise in ground temperature for all regions, ranging from +1.0°C in India to +1.6°C in North America (Figs. 4f,n). The most pronounced seasonal response is an increase in ground temperature of +2.3°C during spring in North America. For all regions, 99%–100% of ensemble members show an increase in annual ground temperature, making this response extremely robust. Generally, regions with greater evergreen forest cover are characterized by larger ground warming when vegetation cover is reduced. In response to the ground warming, surface air temperatures also increase year-round in all regions, although modestly, ranging from +0.3°C in India to +0.7°C in North America (Figs. 4g,o). The robustness of the increase in annual surface air temperature is quite high, ranging from 84% in China to 98% in North Africa. The low latitude regions of South America, north Australia, and North Africa undergo the most robust increase in surface air temperature.

For all regions, the largest precipitation responses to reduced vegetation cover are always along the flanks (spring or autumn) of the monsoon season and never significant during the actual monsoon season. For five regions, the change in monsoon precipitation is less than 1%, while north Australia has a decline of $-3\%$ ($-1.9 \text{ cm month}^{-1}$ during December–March). The most robust precipitation changes are decreases during spring in north Australia (71%) and during autumn in India (60%) and increases during spring in China (61%). Clearly, the intraensemble spread in response to reduced vegetation cover is much greater for precipitation than temperature.

North Australia is the only region with a statistically significant change in annual precipitation due to reduced vegetation cover, with a decrease of $-0.40 \text{ cm month}^{-1}$ (Fig. 4t). The largest seasonal response in precipitation is a decline of $-1.35 \text{ cm month}^{-1}$ in October–December (OND) across north Australia, including a reduction of $-1.84 \text{ cm}$ in December. This statistically significant reduction in spring precipitation and an insignificant reduction in summer precipitation ($-0.64 \text{ cm month}^{-1}$) in north Australia indicates a delayed and weaker monsoon in ENS compared to CTL. Alternatively, reduced vegetation cover favors an earlier monsoon, with little change in intensity, across China, with a significant increase of $+0.46 \text{ cm month}^{-1}$ in spring and insignificant decrease of $-0.29 \text{ cm month}^{-1}$ in summer.

Annual large-scale precipitation declines in all monsoon regions (not shown), particularly north Australia ($-1.61 \text{ cm}$) and North America ($-1.39 \text{ cm}$). However, the response in annual convective-scale precipitation varies among regions (not shown), ranging from a significant decrease of $-3.17 \text{ cm}$ in north Australia to an insignificant increase of $+0.53 \text{ cm}$ in China. The two climatologically wettest regions, north Australia and India, experience the largest reductions in convective precipitation.

Annual sensible heat flux is diminished in all monsoon regions, particularly outside of the summer monsoon season. The reduction in sensible heat flux is greatest in the subtropical monsoon region of North America, $-1.22 \text{ W m}^{-2}$, and least in the tropical monsoon region of North Africa, $-0.48 \text{ W m}^{-2}$, peaking at $-2.16 \text{ W m}^{-2}$ during summer in North America (Fig. 3c). The most robust decreases in annual sensible heat flux (67%) occur in North America. Annual latent heat flux is also diminished in all regions, ranging from $-0.43 \text{ W m}^{-2}$ in
North America to $-1.80 \text{ W m}^{-2}$ in North Africa (Fig. 3d). Regions with large reductions in sensible heat flux tend to have small reductions in latent heat flux, and vice versa, implying that some monsoon regions have more of a thermal response and others have more of a moisture response to altered vegetation cover. The decrease in annual latent heat flux is most robust in North Africa (83%), peaking in winter, $-4.70 \text{ W m}^{-2}$, with 99%
robustness. The most robust decreases in evapotranspiration occur in the three wettest regions, North Africa, India, and north Australia. In general, the largest decreases in transpiration and canopy evaporation and increases in ground evaporation occur in summer–autumn, with forested areas experiencing larger reductions in transpiration and increases in ground evaporation than grassy areas. For all regions, the largest and most robust decreases in evapotranspiration occur in winter–spring, mainly attributed to transpiration loss.

The annual decrease in latent heat exceeds that of sensible heat in the three wettest regions (North Africa, north Australia, and India), while the opposite is true of the three driest regions (North America, China, and South America). Therefore, the Bowen ratio increases in the climatologically wet regions, such as +0.024 in North Australia, and decreases in the dry, subtropical regions, such as −0.032 and −0.020 in North America and China, respectively.

Although the annual changes in surface albedo are statistically significant, they are quite small and vary in sign by region, ranging from −0.004 in South America to +0.003 in North Africa (Figs. 5j,v). The largest reduction in albedo is −0.008 during summer in South America due to reduced grass LAI. The increase in surface albedo in North Africa is attributed to a wintertime drying of the top soil layer related to enhanced ground evaporation, given that drier soils have a larger albedo.

Based on annual 700-hPa omega, there is greater atmospheric subsidence in all regions due to reduced vegetation cover, particularly across North America and north Australia (Figs. 5g,s). The most notable changes in 700-hPa omega are diminished ascent during spring (+0.0038 Pa s⁻¹, 74% robustness) and summer (+0.0014 Pa⁻¹) in North Australia, indicating a weaker, delayed monsoon; this corresponds with anomalous divergence within the boundary layer.

Diminished evapotranspiration and enhanced subsidence contribute to annual reductions in precipitable water for all regions, with the largest annual reductions in India and north Australia (Figs. 5p,t). The largest seasonal response in precipitable water is a decrease of −0.78 kg m⁻² over north Australia in spring, supporting a delayed monsoon. China receives a notable and significant reduction in precipitable water of −0.36 kg m⁻² during the summer monsoon season (Fig. 5d). The wettest regions, north Australia and India, defined as having the greatest climatological annual precipitation, are characterized by the largest ratio of reduced precipitable water to reduced LAI of the six regions.

f. Vertical profile of atmospheric response

While an annual enhancement of lower-middle tropospheric subsidence is consistent among the regions, the seasonal responses of vertical motion to reduced vegetation cover are complex. Of the six regions, the most pronounced vertical motion anomalies are induced over north Australia, China, and North America.

In China, deep anomalies of ascent in May and descent in September throughout the troposphere support an earlier monsoon season (Fig. 6a). Likewise, both precipitation and tropospheric condensational heating from convection (largest diabatic heating term) increase in May–July and decrease in August–October. The negative anomaly in turbulent fluxes peaks in May at −3.5 W m⁻², leading to the largest ground warming of +1.9°C and an intensified ocean–land thermal contrast in the premonsoon season. The inverse relationship between ground temperature and sea level pressure peaks during May in CTL. With the surface heating, pressure is reduced, moisture convergence increases, precipitable water increases, and ascent is enhanced, leading to an earlier monsoon onset. In response to the subsidence anomaly in the late monsoon season, the specific humidity of the lower-middle troposphere is reduced in China during August–September (Fig. 7a). Convective cloud cover fraction actually increases and outgoing longwave radiation (OLR) decreases during June–July, related to a steeper lapse rate and anomalous ascent during the early monsoon season. This mid-monsoon increase in convective cloud amount is noted for all regions.

Anomalous subsidence occurs in north Australia during October–January, with a delayed, weaker monsoon (Fig. 6e) that represents the opposite response to reduced vegetation cover than in China. Along with this subsidence anomaly, there are significant decreases in precipitation, total cloud cover, total column specific humidity (Fig. 7e), and moisture flow from the ocean in north Australia during November–December. With the dampened monsoon system, condensational heating from convection is reduced throughout the troposphere during October–February. OLR during the premonsoon period of November–December increases by +3.20 W m⁻², which is further evidence of reduced convective activity.

Anomalous subsidence in November–December in North America (Fig. 6b) coincides with an anomalous anticyclone and diminished moisture flux convergence, which lowers precipitation and precipitable water over Mexico. Deep tropospheric drying occurs in November–December (Fig. 7b) with diminished cloud cover in November. Convective cloud cover fraction increases in July–August with a steeper lapse rate and a localized
area of enhanced ascent in the mid–upper troposphere (Fig. 6b).

The tropospheric response to reduced vegetation cover is less intense over North Africa, South America, and India. With diminished evapotranspiration in winter, specific humidity in the lower-middle troposphere is reduced during January–March in North Africa (Fig. 7c). However, anomalous ascent during August (Fig. 6c) favors more convective cloud amount during the monsoon. Similarly, in South America, convective
precipitation and convective cloud fraction increase and OLR decreases in January–May, with an insignificant enhancement in ascent.

g. Nonlocal circulation responses

Velocity potential anomalies between ENS and CTL reveal that variations in vegetation cover in a monsoon region may alter the large-scale circulation pattern and even induce cross-equatorial impacts on the divergence fields over other monsoon regions (Fig. 8). Associated with a delay in the Australian monsoon, reduced vegetation cover triggers a negative anomaly in velocity potential, meaning anomalous divergence, at 925–700 hPa and a positive anomaly in velocity potential at 250 hPa during October–December; these divergence anomalies correspond to anomalous subsidence (Figs. 8a–c). The pattern of low-level divergence and upper-level convergence spreads from north Australia across the equator into East Asia in Northern Hemisphere autumn. Reductions in precipitable water are evident across Australia, Indonesia, East Asia, and South Asia.

Besides north Australia, the other monsoon region that appears to trigger a widespread response in velocity potential is over North America (Figs. 8d–f). Likewise, during January–March, anomalous subsidence over North America corresponds to anomalous divergence at 850–700 hPa and convergence at 250 hPa. The upper-level convergence into North America extracts air from the South American monsoon region during its corresponding wet season.

4. Discussion and conclusions

To investigate intra-annual vegetation feedbacks on climate in the global monsoon regions, both a control simulation and initial value ensemble experiments with reduced vegetation cover are run using NCAR CCSM3.5-DGVM. The model is capable of simulating the seasonal monsoon evolution over each of

FIG. 6. Vertical profile (height AGL in meters) of mean monthly changes (ENS − CTL) in vertical motion (Pa s⁻¹) for (a) China, (b) North America, (c) North Africa, (d) India, (e) north Australia, and (f) South America. The seasons are identified as WIN for winter, SPR for spring, SUM for summer, and AUT for autumn. The yellow highlight identifies the simulated monsoon season. Red (blue) indicates anomalous descent (ascent). Hatching indicates 90% significance, based on t tests.
six monsoon regions but generally the monsoons are too wet and penetrate too far inland.

The following are the key findings of this study:

1) Reductions in vegetation cover induce both consistent responses among the monsoon regions, such as diminished turbulent fluxes, surface warming, and a dampened hydrological cycle, and unique responses, with changes in the timing of the monsoon season that differs by tropical versus subtropical locations.

2) Changes in the hydrological cycle and vertical motion fields are most distinct during the onset/decay of the monsoon season (spring or autumn).

3) A local reduction in vegetation cover can trigger a large-scale circulation response with anomalous low-level divergence and subsidence, which decrease moisture flux convergence and may be even more critical than the change in local moisture recycling.

Given that each monsoon region has its own background climatology, latitudinal position, and vegetation types, unique climatic responses to reduced vegetation cover are produced among the six monsoon regions. Generally, the thermal responses (e.g., temperature, diurnal temperature range) are more statistically robust and consistent among the regions than the hydrological responses (e.g., precipitation, precipitable water, clouds). The climatic responses to reduced vegetation cover are most consistent among the six regions prior to the monsoon season and then diminish in consistency during the summer–autumn wet period, mostly attributed to the less consistent hydrological responses to reduced vegetation cover.

With reduced vegetation cover, LAI is decreased, particularly in regions with more evergreen tree cover. Reduced LAI causes diminished leaf area available for transpiration and interception, conductance (Hales et al. 2004), and roughness length—all of which favor a reduction in turbulent fluxes, including both sensible and latent heat fluxes. Kirk-Davidoff and Keith (2008) found that a decrease in roughness causes a reduction in the aerodynamic drag coefficient and thus less sensible and latent heat fluxes. The reductions in turbulent fluxes and roughness lead to a drier atmosphere with anomalous subsidence, less precipitation, and higher ground and surface air temperatures (Foley et al. 2003). The increase in ground temperatures is most distinct in regions

FIG. 7. As in Fig. 6 but for specific humidity (g kg$^{-1}$). Red (blue) indicates anomalously moist (dry) air.
with dense evergreen tree cover while the drying is strongest in woody regions, both of which are further evidence that trees and grasses induce differing feedbacks, as noted by Liu et al. (2010). The decrease in both sensible and latent heat fluxes, particularly the latter, with reduced LAI, is consistent with studies by Hoffmann and Jackson (2000) and Oyama and Nobre (2004) for tropical regions. The reduction in latent heat flux is largest in the wetter tropical monsoon regions and the reduction in sensible heat flux is largest in the drier subtropical monsoon regions, with a decreasing Bowen ratio in the subtropical regions. The findings of local surface warming and diminished evapotranspiration and precipitation agree with tropical deforestation studies (Eltahir and Bras 1993; Lean and Rowntree 1997; Hahmann and Dickinson 1997; Dickinson and Henderson-Sellers 1988; Costa and Foley 2000; Osborne et al. 2004; Nosetto et al. 2005). Durbidge and Henderson-Sellers (1993) also produce enhanced subsidence in response to deforestation. In addition to reductions in sensible and latent heat flux, changes to the surface radiation budget include increases in upward longwave radiation (due to a warmer ground), downward shortwave radiation (due to less clouds and moisture), and heat flux into the soil (due to more exposed soil). In response to the increased flux of heat into the soil with diminished LAI, the ground and surface air temperatures rise at night, reducing the diurnal temperature range, particularly in the drier subtropical monsoon regions of North America and China. This is consistent with the study of Zhou et al. (2007). Lawrence and Chase (2010), in an investigation of land use impacts on climate using CCSM3.5, attributed surface warming to reduced evapotranspiration as in the present study.

FIG. 8. Mean changes (ENS − CTL) in velocity potential (m² s⁻¹ scaled by 10⁶) for (a) 925 hPa, (b) 700 hPa, and (c) 250 hPa during OND, shown for the area around the north Australian monsoon region. Mean changes in velocity potential for (d) 850 hPa, (e) 700 hPa, and (f) 250 hPa during JFM for the area around the North American monsoon region. Red (blue) indicates anomalous convergence (divergence).
As shown in prior studies, deforestation tends to produce cooling at high latitudes through an albedo feedback (Betts 2000; Govindasamy et al. 2001) and warming in the low latitudes through an evapotranspiration feedback (Costa and Foley 2000; Osborne et al. 2004; Gibbard et al. 2005). The present study demonstrates the latter processes are involved in the global monsoon regions, both tropical and subtropical. As noted by Gibbard et al. (2005), the partitioning of heat into evapotranspiration is more efficient at lower latitudes owing to the exponential relationship between temperature and saturation vapor pressure, which partly explains the dominant role of the vegetation–evapotranspiration feedback instead of the albedo feedback. Reduced vegetation cover fraction produces only small surface albedo changes in the monsoon regions, consistent with studies by Osborne et al. (2004) and Watts et al. (2007). The high LAI values simulated by CCSM3.5-DGVM in CTL might partly explain the low sensitivity of surface albedo to LAI changes (Bonan 1997).

In agreement with Osborne et al. (2004), reduced LAI induces year-round warming in all regions but the impacts on precipitation vary by season and region. Nonetheless, the general finding is a dampened hydrologic cycle. The increases in ground and surface air temperatures are predominantly local, within the monsoon regions, while large-scale responses in velocity and temperature are predominantly local, within the monsoon regions, while large-scale responses in velocity and temperature are predominantly local, within the monsoon regions. Further study is warranted, including the possibility of running individual experiments of reduced vegetation cover for each monsoon region, although this would be computationally expensive. In addition, experiments in which the LAI is reduced in specific months would help clarify seasonally unique vegetation feedbacks.

Most of the significant responses in precipitable water, omega, and precipitation to reduced vegetation cover occur along the flanks of the monsoon season, with influence on the monsoon timing but little impact on actual monsoon precipitation for most regions. Similar to Osborne et al. (2004), the present study shows that vegetation feedbacks are seasonally dependent with the greatest impacts outside of the wet season when soil moisture availability is limited. The impacts of the vegetation feedbacks extend beyond the boundary layer, into the mid–upper troposphere, in terms of changes in rising motion and atmospheric moisture (Bonan 1997).

In response to reduced vegetation cover, the north Australian monsoon is delayed and weakened, while the monsoon season shifts earlier in the year over China and the southwest United States. Over north Australia (tropics), local recycling is weakened with less evapotranspiration leading up to onset, and moisture convergence is diminished with low-level divergence and anomalous tropospheric subsidence. The drying extends into the continental interior, similar to the study by Miller et al. (2005). In the case of China and the southwest United States (subtropics), sensible heat flux is reduced in winter–spring, allowing the ground to warm and the land–ocean thermal contrast to intensify, which produces an earlier monsoon onset (Lee et al. 2010, manuscript submitted to J. Geophys. Res.). In general, latent heat anomalies are critical in the tropics and sensible heat anomalies are critical in the subtropics with vegetation feedbacks in the monsoon regions. Like Osborne et al. (2004), we found significant vegetation influences on the China monsoon but minimal impacts on the Indian monsoon, given that the latter is much more controlled by ocean forcings in CCSM3.5. Across north Australia and North America, notable moisture flux divergence anomalies are induced such that the decline in precipitation exceeds the decline in evapotranspiration by three to four times. Hahmann and Dickinson (1997) likewise concluded that most models produce a decrease in moisture convergence in response to tropical deforestation. The reduction in precipitation and moisture convergence in north Australia, along with the minimal response in precipitation and increase in moisture convergence in North Africa, match the findings of Hoffmann and Jackson (2000), who impose LAI reductions in tropical savannas.

Several broader implications emerge from this modeling study. It is apparent that vegetation dynamics are important elements of the climate system, especially over northern Australia, and need to be included in future stages of the Climate Model Intercomparison Project (CMIP). Future heat stress could reduce vegetation cover in certain regions, which could further enhance the surface warming. In the case of northern Australia, drought events can trigger vegetation loss, which through positive vegetation–precipitation feedbacks could amplify the drought intensity. Continued agricultural practices in China and India and grazing in North America and northern Australia likely reduce the total leaf area and could lead to surface heating and a dampened hydrological cycle.

Acknowledgments. This study was funded by DOE NICCR, NOAA CPPA, and USDA Forest Service. Experiments were performed using NERSC resources. The authors are grateful for assistance with CCSM3.5 from Sam Levis, Jon Wolfe, Steve Vavrus, and Feng He and appreciate the discussions with Dr. Gallimore and Professor Gutzler. Comments by two anonymous reviewers were greatly appreciated.
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