The Role of Plant CO₂ Physiological Forcing in Shaping Future Daily-Scale Precipitation[Ⓐ]

CHRISTOPHER B. SKINNER AND CHRISTOPHER J. POULSEN

Department of Earth and Environmental Sciences, University of Michigan, Ann Arbor, Michigan

ROBIN CHADWICK

Met Office Hadley Centre, Exeter, United Kingdom

NOAH S. DIFFENBAUGH

Department of Earth System Science, and Woods Institute for the Environment, Stanford University, Stanford, California

RICHARD P. FIORELLA

*Department of Earth and Environmental Sciences, University of Michigan, Ann Arbor, Michigan, and
Department of Geology and Geophysics, University of Utah, Salt Lake City, Utah*

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ABSTRACT

Continued anthropogenic CO₂ emissions are expected to drive widespread changes in precipitation characteristics. Nonetheless, projections of precipitation change vary considerably at the regional scale between climate models. Here, it is shown that the response of plant physiology to elevated CO₂, or CO₂ physiological forcing drives widespread hydrologic changes distinct from those associated with CO₂ radiative forcing and has a role in shaping regional-scale differences in projected daily-scale precipitation changes. In a suite of simulations with the Community Climate System Model, version 4 (CCSM4), reduced stomatal conductance from projected physiological forcing drives large decreases in transpiration and changes the distribution of daily-scale precipitation within and adjacent to regions of dense vegetation and climatologically high transpiration. When atmospheric conditions are marginally favorable for precipitation, reduced transpiration dries the boundary layer and increases the likelihood of dry day occurrence. In CCSM4, the annual number of dry days increases by upward of 15 days yr⁻¹ over tropical land and the continental midlatitudes. Decreases in transpiration from physiological forcing also increase the number of heavy precipitation events by up to 8 days yr⁻¹ in many tropical forest regions. Despite reductions in the land surface contribution to atmospheric moisture, diminished surface latent heat fluxes warm the forest boundary layer and increase moisture convergence from nearby oceans, enhancing instability. The results suggest that consideration of the radiative impacts of CO₂ alone cannot account for projected regional-scale differences in daily precipitation changes, and that CO₂ physiological forcing may contribute to differences in projected precipitation characteristics among climate models.

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Corresponding author e-mail: Christopher B. Skinner, chrisbs@umich.edu

1. Introduction

Terrestrial evapotranspiration (ET) accounts for roughly 35% of the moisture that falls as precipitation on land (van der Ent et al. 2014). The contribution of regional evapotranspiration to total regional rainfall, the precipitation recycling ratio, varies in space and time

and is generally highest where surface and root zone soil moisture is abundant and when advective moisture fluxes are relatively small (Trenberth et al. 2003). Regions of high precipitation recycling over land rely on terrestrial moisture sources for precipitation and are particularly sensitive to any forcing that impacts local or upwind land ET rates, such as changes in land cover or greenhouse gas concentrations (Pielke 2001; Malhi et al. 2008; Seneviratne et al. 2006). Further, variations in the partitioning of total ET between transpiration and evaporation can impact surface energy flux variability and drive changes in the intensity and frequency of individual rainfall events, particularly in regions of high ET and precipitation recycling (Lawrence et al. 2007; Lee et al. 2012). Estimates of future changes in ET and its partitioning into evaporation and transpiration are essential for projecting rainfall changes and the associated impacts on humans and ecosystems worldwide.

Enhanced radiative forcing due to increased carbon dioxide (CO₂) concentrations, among other greenhouse gases, is expected to drive widespread changes in ET (evaporation over ocean, and evaporation plus transpiration over land) (Milly and Dunne 2016; Swann et al. 2016). Plant physiological responses to increased CO₂ also promote changes in ET unrelated to the radiative effects of CO₂. In response to high ambient CO₂ concentrations, many plants decrease stomatal conductance and transpiration in order to minimize water loss, increasing their plant water-use efficiency (the ratio of carbon assimilation rate to transpiration rate) (Field et al. 1995; Lammertsma et al. 2011). These decreases in transpiration affect the amount and variability of soil moisture and ET, both of which have been shown to influence the frequency, intensity, and duration of precipitation events (Eltahir 1998; Lawrence et al. 2007; C. Taylor et al. 2012; Lee et al. 2012). Given that transpiration accounts for roughly 64% of total terrestrial ET (Good et al. 2015), understanding the complex and often competing responses of vegetation–water interactions to CO₂ variations is critical for future hydrologic assessment.

Climate model experiments designed to isolate the impact of projected CO₂ physiological forcing (stomata changes) on future climate, with vegetation structure and vegetation location fixed, exhibit widespread reductions in transpiration in response to a doubling of atmospheric CO₂ (Betts et al. 1997; Douville et al. 2000; Cao et al. 2010; Andrews et al. 2011; Pu and Dickinson 2014). Reduced transpiration rates are highest in the midlatitudes during the summer season, and in tropical forest regions (Douville et al. 2000; Pu and Dickinson 2014). Reductions in transpiration impact local precipitation directly by diminishing a source of moisture to the atmosphere and indirectly by altering regional-scale circulation (Betts et al. 2004, 2007;

Boucher et al. 2009). Climate models project the greatest changes in precipitation (positive and negative) from future CO₂ physiological forcing in densely vegetated areas of the tropics (Betts et al. 1997; Swann et al. 2016).

Climate model sensitivity studies suggest that reduced transpiration leads to greater precipitation variability in regions of high transpiration and high precipitation recycling (Lawrence et al. 2007; Lee et al. 2012). Lee et al. (2012) set the transpiration flux to zero within modern-day climate simulations with the Community Climate System Model, version 3 (CCSM3), and found that tropical forests experience a higher frequency of both precipitation-free days and days with extremely heavy precipitation (>97th percentile). Similarly, Lawrence et al. (2007) modified vegetation and soil hydrology parameterizations in a present-day climate simulation with the Community Land Model, version 3 (CLM3), and found that local increases in transpiration at the expense of canopy evaporation shift the distribution of rainfall toward more frequent, lighter events. The results of Lawrence et al. (2007) and Lee et al. (2012) suggest that reductions in transpiration from CO₂ physiological forcing may contribute to increased precipitation variability in regions of high climatological transpiration and in regions that rely heavily on transpiration for rainfall. However, the prescribed transpiration changes in these studies are highly idealized, and the response of precipitation to projected, physically based variations in transpiration from CO₂ physiological forcing is unclear.

While elevated CO₂ is expected to reduce stomatal conductance, projected CO₂ fertilization and the associated changes in vegetation density and location further complicate the response of climate to CO₂ physiological forcing. In environments where CO₂ supply limits gross primary production, increases in CO₂ stimulate enhanced biomass production and greater leaf area index (LAI) (Cowling and Field 2003; Lloyd and Farquhar 2008; Zhu et al. 2016). Within climate model simulations of a high CO₂ climate, this CO₂ fertilization effect can lead to greater canopy-scale transpiration and partially or completely negate the diminished transpiration from reduced stomatal conductance, impacting regional-scale precipitation change (Betts et al. 1997; Levis et al. 2000; Kergoat et al. 2002).

Changes in precipitation characteristics have the potential to cause widespread impacts on natural and human systems (Field et al. 2012; Anderegg et al. 2013; Hirabayashi et al. 2013; Frank et al. 2015). Efforts to anticipate and prepare for these impacts are made easier when projections of precipitation change and the underlying processes that drive these changes are consistent among climate model simulations. At regional scales, there remains considerable uncertainty in projections of future precipitation characteristics and the processes that

TABLE 1. Simulation names and climate forcings.

Simulation	Atmospheric CO ₂	Land surface CO ₂	LAI
HistRad_HistStomata_HistLAI	367 ppm	367 ppm	End twentieth century
HistRad_FutStomata_FutLAI	367 ppm	900 ppm	End twenty-first century
FutRad_HistStomata_FutLAI	900 ppm	367 ppm	End twenty-first century
FutRad_FutStomata_HistLAI	900 ppm	900 ppm	End twentieth century
FutRad_FutStomata_FutLAI	900 ppm	900 ppm	End twenty-first century

shape them (O’Gorman and Schneider 2009). To date, the potential impact of CO₂ physiological forcing and CO₂ fertilization on future changes in precipitation characteristics remains largely unexplored. Here, we investigate the link between future CO₂–vegetation interactions and the hydrologic cycle with the goal of providing insights into the mechanisms that shape projected regional-scale changes in precipitation characteristics on daily time scales. We utilize a suite of sensitivity experiments with the Community Climate System Model, version 4 (CCSM4), to investigate the individual contributions of CO₂ radiative forcing, CO₂ physiological forcing, and CO₂ fertilization on future hydrologic changes. Our results suggest that consideration of CO₂ physiological forcing is critical for understanding regional differences in the projected response of precipitation characteristics to elevated CO₂ concentrations.

2. Data and methods

To analyze the contributions of CO₂ radiative forcing, CO₂ physiological forcing, and CO₂ fertilization to projected future hydrologic changes, we conduct a set of five 60-yr climate model simulations with the CCSM4 (Gent et al. 2011). Within CCSM4, the Community Atmosphere Model, version 4 (CAM4; Neale et al. 2010), and the Community Land Model, version 4 (CLM4; Oleson et al. 2010), are run on a $0.9^\circ \times 1.25^\circ$ finite-volume grid. Sea ice is modeled with the fully active Community Ice Code, version 4 (CICE4; Hunke and Lipscomb 2008). The ocean is represented with a mixed layer, slab ocean model run at a nominal resolution of 1° (Bitz et al. 2012). The atmosphere has 26 terrain-following vertical levels, and the soil column is divided into 15 vertical layers. All simulations are initialized from the same point in a preindustrial CCSM4 run (denoted as b40.1850.track1.1deg.006) and therefore have identical initial conditions (land surface, atmosphere, ocean, and sea ice). We analyze the final 30 years of each 60-yr simulation. Our experimental design is similar to that of Pu and Dickinson (2014), although here our goal is to better understand changes in daily precipitation characteristics from future CO₂–vegetation interactions.

We vary CO₂ concentrations in the land and atmosphere models separately to identify the relative contributions of

CO₂ radiative versus physiological forcing (Table 1). Changes in atmospheric CO₂ used in the atmosphere model capture the radiative effect, and are denoted as “Rad” in the simulation name. Changes in atmospheric CO₂ used in the land model capture the plant physiological effect, and are denoted as “Stomata” in the simulation name. The atmospheric CO₂ concentration in the atmosphere and land models is either prescribed at 367 ppm (historical, ~year 2000), denoted as “Hist” in the simulation name, or 900 ppm [future, ~year 2095 in the representative concentration pathway 8.5 (RCP8.5) scenario], denoted as “Fut” in the simulation name (van Vuuren et al. 2011). We also vary vegetation area between historical or future predicted values to estimate climate change due solely to changes in CO₂ fertilization. Climatological (monthly varying) LAI and stem area index (SAI) values in each grid box and plant functional type (PFT) are prescribed from an end of the twentieth-century time period [mean of years 1996–2005 from phase 5 of the Coupled Model Intercomparison Project (CMIP5) CCSM4 “historical” scenario] denoted as “HistLAI,” or end of twenty-first-century time period (mean of years 2091–2100 from the CMIP5 CCSM4 future RCP8.5 scenario) denoted as “FutLAI.” Changes in LAI or SAI between the Hist and Fut simulations arise from the direct effects of CO₂ fertilization on vegetation structure, and from the indirect effects of elevated CO₂ on vegetation structure through changes in climate (temperature, moisture, etc.) (Fig. S1 in the supplemental material). The changes in LAI and SAI do not include changes in vegetation location; the type and amount of PFTs in each grid cell are the same in the Hist and Fut simulations.

We isolate the influences of CO₂ changes on hydroclimate due to radiative forcing, stomatal conductance changes (physiological forcing), and LAI changes (CO₂ fertilization) by comparing simulations where one of these factors are changed to an historical control simulation (Table 2). By subtracting each experimental simulation from the future simulation (FutRad_FutStomata_FutLAI), we are able to assess the impact of both combined and individual forcings within the context of future end of twenty-first century climate. To assess the full impact of projected future changes in CO₂ (CO₂_Total), including the combined influences

TABLE 2. Experiment names and climate forcings.

Experiment	Calculated as difference between:
CO2_Total	FutRad_FutStomata_FutLAI minus HistRad_HistStomata_HistLAI.
CO2_Rad	FutRad_FutStomata_FutLAI minus HistRad_FutStomata_FutLAI.
CO2_Stomata	FutRad_FutStomata_FutLAI minus FutRad_HistStomata_FutLAI.
CO2_LAI	FutRad_FutStomata_FutLAI minus FutRad_FutStomata_HistLAI.

of radiative forcing, stomatal conductance, and LAI changes, we subtract the historical control simulation HistRad_HistStomata_HistLAI from the future simulation FutRad_FutStomata_FutLAI. To isolate the impact of future CO₂ radiative forcing (CO2_Rad), we subtract HistRad_FutStomata_FutLAI from simulation FutRad_FutStomata_FutLAI. To isolate the impact of future CO₂ physiological forcing (CO2_Stomata), we subtract simulation FutRad_HistStomata_FutLAI from simulation FutRad_FutStomata_FutLAI. Last, to isolate the impact of future CO₂ fertilization (CO2_LAI), we subtract FutRad_FutStomata_HistLAI from simulation FutRad_FutStomata_FutLAI.

CLM4 predicts leaf stomatal conductance separately for C₃ and C₄ plants using the Ball–Berry conductance

model (Oleson et al. 2010). The model relates stomatal conductance to the net photosynthesis rate, atmospheric relative humidity, and the CO₂ concentration at the leaf surface. Soil water indirectly impacts stomatal conductance via photosynthesis. Specifically, soil moisture stress (low soil moisture) reduces photosynthesis, which then limits stomatal conductance (Oleson et al. 2010). Increases in atmospheric CO₂ promote decreases in stomatal conductance, assuming other environmental variables do not change. All other greenhouse gases and aerosol concentrations in the simulations are prescribed and fixed at year 2000 values for all simulations (CH₄ = 1753 ppb and N₂O = 319 ppb). Therefore, radiative forcing in our experiments refers only to the projected energy perturbation from changes in CO₂. Consequently, our experiments underestimate the full radiative forcing expected from future greenhouse gas changes during the twenty-first century.

We follow the methodology of Pitman and Zhao (2000) and Skinner et al. (2012) to estimate the relative contribution of CO₂ radiative forcing, CO₂ physiological forcing, and CO₂ fertilization to the total simulated climate response to elevated CO₂. The calculation for the percent contribution of physiological forcing is as follows:

$$\frac{|\text{CO2_Stomata}|}{|\text{CO2_Rad}| + |\text{CO2_Stomata}| + |\text{CO2_LAI}| + |\text{NONLINEAR_INTERACTION}|} \times 100,$$

where

$$\begin{aligned} \text{NONLINEAR_INTERACTION} \\ = |\text{CO2_Total} - (\text{CO2_Rad} \\ + \text{CO2_Stomata} + \text{CO2_LAI})|. \end{aligned}$$

Likewise, we replace |CO2_Stomata| in the numerator with |CO2_Rad| to calculate the percent contribution of radiative forcing to the total climate response, etc. The final term in the denominator represents the nonlinear interaction between the CO₂ radiative forcing, CO₂ physiological forcing, and CO₂ fertilization. The use of absolute values in the calculation allows both negative and positive changes to contribute to the total change.

Additionally, to test the robustness of our results to the selection of climate model and to the role of ocean–atmosphere coupling, we make use of two similar climate model experiments with the Hadley Centre Global Environment Model, version 2, for atmosphere and land (HadGEM2-A; Jones et al. 2011) and the Community

Earth System Model, version 1, with biogeochemistry [CESM1(BGC); Lindsay et al. 2014]. First, we compare two 30-yr simulations from HadGEM2-A to estimate the combined impact of projected changes in CO₂ fertilization and CO₂ physiological forcing on future hydrologic change (Chadwick et al. 2017). The only difference between the two HadGEM2-A simulations is the concentration of atmospheric CO₂ seen by the land surface model. In the first simulation, the land model sees prescribed, time-varying CO₂ concentrations from the historical period 1979–2008 (335–385 ppm). In the second simulation, the time-varying CO₂ concentrations in the land model are quadrupled (1340–1540 ppm). Both simulations utilize the same prescribed historical period sea surface temperatures, and quadrupled time-varying atmospheric CO₂ concentrations (1340–1540 ppm). Unlike the CCSM4 simulations, the effects of changing LAI and stomatal conductance are included simultaneously in the HadGEM2-A simulation. Further details of the HadGEM2-A simulations are described in Chadwick et al. (2017).

We also compare our CCSM4 results to those from the coupled atmosphere–ocean CESM1(BGC) simulation, which is archived as part of the carbon–climate feedback experiment within CMIP5 (Arora et al. 2013; Swann et al. 2016). CESM1(BGC) shares many of the same components as CCSM4, including the atmosphere (CAM4) and land (CLM4) models. The primary difference between the CCSM4 simulations and the CESM1(BGC) simulation is the use of a dynamic ocean model in CESM1(BGC). By comparing our CCSM4 simulation with CESM1(BGC) we are able to assess the role of deep ocean circulation in shaping the climate response to CO₂ physiological forcing. We analyze a single, transient (140 yr) simulation, denoted CESM1-BGC_esmFixClim1 in the CMIP5 archive, where the CESM1(BGC) land and ocean models see an increase in atmospheric CO₂ concentration from 284 to 1140 ppm (increases at a rate of 1% yr⁻¹), but atmospheric CO₂ is fixed at 284 ppm in the atmosphere model. Stomatal conductance and LAI/SAI (as well as ocean biogeochemistry) are free to respond to the CO₂ increase. Similar to the HadGEM2-A analysis, the experimental design of the CESM1(BGC) simulation does not allow us to quantify the individual contributions of CO₂ fertilization and CO₂ physiological forcing to the total CO₂–vegetation forcing. Instead, we quantify the combined influence of CO₂ fertilization and CO₂ physiological forcing on future climate. To estimate the role of CO₂–vegetation interactions on projected hydroclimate changes in CESM1(BGC), we compare data from the first and final 30 years of the transient simulation.

Finally, we evaluate the simulation of precipitation in CCSM4 by comparing our simulations to the daily resolution Global Precipitation Climatology Project (GPCP), version 1.2 dataset (Huffman et al. 2001), and to 15 coupled atmosphere–ocean climate models from the CMIP5 historical and RCP8.5 experiments (K. Taylor et al. 2012) (see Table S1 in the supplemental material). The daily GPCP observation dataset covers the time period from 1997 through 2014 at a global 1° resolution. We interpolate all CMIP5 model data to a common 1° × 1° grid using a patch recovery method (Gu et al. 2004). All models from CMIP5 include the effects of future radiative forcing from CO₂, aerosols, and other greenhouse gases. The majority of CMIP5 models also include the effects of CO₂ physiological forcing (Table S1). We assess the statistical significance of changes (at the 95% confidence level) with the use of a permutation test. The permutation test is a nonparametric method of significance testing that makes no assumptions about the underlying distribution of the data.

3. Results

a. Evapotranspiration changes

The latitudinal distribution of historical period terrestrial ET (HistRad_HistStomata_HistLAI) and of the projected terrestrial ET change in CO₂_Total is consistent with historical period ET (1980–2005) and projected ET change [(2080–2100) – (1980–2005)] in 15 coupled atmosphere–ocean climate models from CMIP5 (Figs. 1a,b; see also Table S1). Although CCSM4 exhibits relatively high terrestrial ET change in the Southern Hemisphere and relatively low ET change in the Northern Hemisphere, it is important to note that unlike the RCP8.5 simulations, our CCSM4 simulations do not include projected changes in aerosol and non-CO₂ greenhouse gas concentrations, which may account for some of the disparity (Fig. 1b) (see section 2). For example, prescribed reductions in aerosol concentrations in the RCP8.5 simulations enhance Northern Hemisphere warming and may lead to greater ET change compared with our fixed aerosol CCSM4 simulations (Cox et al. 2008; Kloster et al. 2010). Overall, the simulated change in terrestrial evapotranspiration in CO₂_Total closely matches that from the fully coupled (dynamic ocean) CCSM4 experiment from CMIP5 (2080–2100) – (1980–2005) (Gent et al. 2011), which suggests our idealized slab ocean simulations serve as a good approximation for the fully coupled dynamic ocean CCSM4 model (Fig. S2 in the supplemental material).

Over land, projected future CO₂ radiative and CO₂ physiological forcing drive large but opposing changes in annual mean ET (Fig. 2). Surface warming from CO₂ radiative forcing (CO₂_Rad) enhances ET across most continental regions (Fig. 2b). In contrast, changes in terrestrial ET from CO₂ physiological forcing (CO₂_Stomata) are largely negative (Figs. 2c). Within CCSM4, the response of ET to future CO₂–vegetation interactions (CO₂ physiological forcing + CO₂ fertilization) is dominated by projected changes in stomatal conductance (as opposed to changes in LAI) (Figs. 2c,d). Changes in stomatal conductance reduce ET by –0.024 and –0.108 mm day⁻¹ averaged across the globe and land, respectively. In contrast, changes in LAI enhance global and land ET by 0.008 and 0.026 mm day⁻¹, respectively. The overall response of ET to future elevated CO₂ (CO₂_Total) resembles a balance between the individual responses to radiative forcing (CO₂_Rad) and stomatal conductance (CO₂_Stomata) (Fig. 2).

The relative contributions of radiative forcing and physiological forcing to the total CO₂–driven response of annual mean ET vary considerably at the regional scale

FIG. 1. Zonal mean of annual (a) evapotranspiration and (b) evapotranspiration change from CCSM4 and 15 CMIP5 models. Thick black line in (a) denotes the historical period CCSM4 simulation HistRad_HistStomata_HistLAI. Thick black line in (b) denotes the change ($\text{CO}_2_{\text{Total}}$) between CCSM4 simulations FutRad_FutStomata_FutLAI and HistRad_HistStomata_HistLAI. Blue lines in (a) denote the historical period average (1980–2005) from the CMIP5 models. Blue lines in (b) denote the change between the future RCP8.5 (2080–2100) and historical simulations from the CMIP5 models. Only land grid points are used in the zonal mean calculations. All models are regridded to a common $1^\circ \times 1^\circ$ grid. CMIP5 models are listed in Table S1.

(Figs. 3a,b). Throughout much of the densely vegetated tropics (10°S – 10°N), CO_2 -driven changes in stomatal conductance are as important as or more important than CO_2 radiative forcing in driving projected future ET change (Figs. 2 and 3). Over central and eastern North America, western and central Europe, and portions of Southeast Asia and the Congo River basin, stomata-driven reductions in ET largely counteract radiative-driven increases in ET, yielding little to no annual mean ET response to future elevated CO_2 . Over the oceans and sparsely vegetated land, future changes in ET are almost entirely a response to CO_2 radiative forcing. To more fully understand the ET responses, we next assess the individual contributions of transpiration and evaporation to ET change.

Differences between ET changes from future CO_2 radiative and physiological forcing are largely driven by changes in transpiration (Fig. 4). Future radiative forcing and reduced stomatal conductance yield transpiration changes of opposite sign in similar locations across the globe (Figs. 4c,e). In CCSM4, reduced stomatal conductance from elevated CO_2 diminishes transpiration in all vegetated areas, particularly in the tropics, Europe, southeastern Asia, the La Plata basin of Argentina, and the eastern United States (Fig. 4e). Warming from CO_2 radiative forcing increases the vapor pressure deficit and enhances transpiration in these same regions (Fig. 4c). In contrast, both CO_2

radiative forcing and reductions in stomatal conductance increase surface evaporation over most continental areas (with the prominent exception of radiative-driven evaporation changes in Central America and northern South America) (Figs. 4d,f). Evaporation increases with reduced stomatal conductance in part because diminished transpiration rates leave larger quantities of moisture available for evaporation from the soil (Figs. S4a–d in the supplemental material) (Swann et al. 2016).

Although stomatal conductance-driven changes in transpiration and evaporation are often found in the same locations (Figs. 4e,f), the overall response of ET to projected changes in stomatal conductance is dominated by the larger magnitude reductions in transpiration. In contrast, LAI-driven changes in evaporation and transpiration counteract one another, minimizing the impact of changes in CO_2 fertilization on future ET in our simulations (Figs. 2d, 3e, and 4g,h). Specifically, increases in LAI enhance transpiration at the expense of soil moisture, limiting evaporation. Additionally, greater vegetation coverage can limit evaporation by shading the soil from sunlight. Overall, our results indicate that future reductions in stomatal conductance contribute substantially to projected regional-scale changes in ET and that the combined influences of radiative forcing and physiological forcing dominate the future ET response to CO_2 .

Annual Mean Evapotranspiration Change

FIG. 2. Change in annual mean evapotranspiration from (a) CO₂_Total, (b) CO₂_Rad, (c) CO₂_Stomata, and (d) CO₂_LAI. Only statistically significant differences are shown.

b. Precipitation changes

The CCSM4 historical control simulation (HistRad_HistStomata_HistLAI) captures the observed global spatial pattern of annual mean precipitation well (Figs. 5a,b). Additionally, the latitudinal distribution of terrestrial precipitation resembles that of the 15 CMIP5 models (Fig. 5c). Precipitation biases in the CCSM4 historical control simulation are consistent with those found in the twentieth-century climate simulation from the fully coupled dynamic ocean CCSM4 model (Gent et al. 2011). Compared with GPCP data, CCSM4 overestimates precipitation along the intertropical convergence zones (ITCZs) of the Indian and Pacific Oceans and in topographically complex regions including the Himalayas, the Andes, and Rwenzori Mountains of eastern equatorial Africa (Figs. 5a,b). Gent et al. (2011) note that CCSM4 exhibits stronger than observed amplitude in El Niño–Southern Oscillation (ENSO) variability, and weaker than observed Madden–Julian oscillation (MJO) variability, both of which impact precipitation, particularly in the tropics. A detailed evaluation of twentieth-century climate in CCSM4 can be found in Gent et al. (2011). Terrestrial precipitation change in CO₂_Total is similar, although slightly greater in magnitude than that in the fully coupled

(dynamic ocean) CCSM4 experiment from CMIP5 (Fig. S3 in the supplemental material). Similar to ET change, the simulated precipitation change in the Southern Hemisphere in CO₂_Total is relatively high compared with the CMIP5 climate change simulations (RCP8.5 – historical) (Fig. 5d).

CO₂ radiative and CO₂ physiological forcing drive distinct patterns in regional precipitation change (Figs. 6b,c). Projected changes in CO₂ radiative forcing impact precipitation across most of the globe (Figs. 6b and 7a). Enhanced radiative forcing brings increased precipitation to the high latitudes and large portions of the midlatitudes, including much of Asia, Australia, North America, and South America. Rainfall decreases over portions of the subtropics, as well as northern South America. Future CO₂ radiative forcing drives particularly large precipitation increases in the tropics, including over the Maritime Continent, eastern equatorial Africa, and the Indian and Pacific Oceans. These tropical precipitation increases are in part driven by thermodynamic increases in atmospheric moisture convergence (the rich-get-richer hypothesis) (Held and Soden 2006) and in part by shifts in convergence zones due to sea surface temperature (SST) pattern changes and changes in regional-scale atmospheric dynamics (e.g., changes in regional

Percent Contribution to Evapotranspiration Change

FIG. 3. Percent contribution to annual mean evapotranspiration change from (a) CO₂_Rad, (b) CO₂_Stomata, (c) CO₂_LAI, and (d) the nonlinear interaction term (see section 2).

Walker circulations) (Chadwick et al. 2013; Kent et al. 2015).

Similar to changes in ET, the response of precipitation to CO₂–vegetation interactions is dominated by changes in stomatal conductance (Figs. 6c,d and 7b,c). The largest changes in precipitation from stomatal conductance are located in the tropics and in portions of the midlatitude Northern Hemisphere, where ET changes are of the greatest magnitude (Figs. 2c, 6c, and 7b). With the exception of the eastern Amazon and far northern South America, changes in stomatal conductance enhance mean annual rainfall over tropical land. Precipitation increases throughout central Africa, the Maritime Continent, southeastern Asia, the equatorial Andes, and the western Amazon. In these areas, enhanced precipitation combined with greater soil moisture from reduced transpiration leads to an increase in surface runoff (Fig. S4) (Gedney et al. 2006; Cao et al. 2010). Over the tropical oceans, annual mean rainfall decreases, particularly in the equatorial Atlantic and eastern Pacific. Drying in the Northern Hemisphere midlatitudes is primarily concentrated in the boreal summer season when vegetation LAI, and hence transpiration, is at its climatological peak (Fig. 8). Projected changes in LAI have little impact on future annual mean precipitation within

CCSM4 (Figs. 6d and 7c). Overall, although CO₂ radiative forcing is the primary driver of future precipitation change across much of the globe, CO₂ physiological forcing is critical in shaping the processes that control future precipitation in regions of high ET such as the tropics and the summer season midlatitudes.

Over most terrestrial regions, the individual precipitation responses to CO₂ radiative forcing, CO₂ physiological forcing, and CO₂ fertilization add nearly linearly to give the total precipitation response to future CO₂ (Fig. 7d and Fig. S5 in the supplemental material). This is consistent with the results of Peng et al. (2014), who found minimal interactive effects between CO₂ radiative and physiological forcing within a subset of CMIP5 models. However, there are some regions where interactions between the different forcings in the CO₂ Total simulation result in nonlinearities. In CCSM4, the largest magnitude deviations on land occur over the highlands of eastern Africa and in portions of southeastern Asia (Fig. S5). Because these regions are topographically complex and located within the humid tropics, even small changes in the location and magnitude of moisture convergence can drive large changes in precipitation, making the linear decomposition approximation less appropriate in these areas.

FIG. 4. Change in annual mean (left) transpiration and (right) evaporation from (a),(b) CO₂_Total, (c),(d) CO₂_Rad, (e),(f) CO₂_Stomata, and (g),(h) CO₂_LAI. Only statistically significant differences are shown. Evaporation over the ocean is masked out.

In addition to changes in annual mean precipitation, elevated CO₂ intensifies the hydrologic cycle, increasing the frequency of dry days and heavy precipitation days in many regions (Sun et al. 2007). Across most of the globe, CO₂ radiative forcing is the primary driver of projected changes in daily precipitation characteristics (Figs. 9 and 10). Future radiative forcing increases annual dry day occurrence (<0.1 mm) over northern South

America, portions of the terrestrial Northern Hemisphere subtropics (~20°–40°N), and along the northern margin of the SH extratropical storm track (~50°S) (Fig. 9b). Dry days become less common in the Sahel, the subtropical Southern Hemisphere, and the high latitudes. Radiative forcing increases the frequency of heavy rainfall days (>95th percentile event from the historical control simulation using only days with greater

Annual Mean Precipitation

FIG. 5. Annual mean precipitation from (a) GPCP (1997–2014) and (b) the 30-yr historical period CCSM4 simulation HistRad_HistStomata_HistLAI. Zonal mean of annual (c) precipitation and (d) precipitation change from CCSM4 and 15 CMIP5 models. Thick black line in (c) denotes the historical period CCSM4 simulation HistRad_HistStomata_HistLAI and in (d) denotes the change ($\text{CO}_2_{\text{Total}}$) between CCSM4 simulation FutRad_FutStomata_FutLAI and simulation HistRad_HistStomata_HistLAI. Blue lines in (c) denote the historical period average (1980–2005) from the CMIP5 models and in (d) denote the change between the future RCP8.5 (2080–2100) and historical simulations from the CMIP5 models. Only land grid points are used in zonal mean calculations for (c) and (d). All models are regridded to a common $1^\circ \times 1^\circ$ grid. CMIP5 models are listed in Table S1.

than 0.1 mm of precipitation) across broad stretches of the globe, including the mid- and high latitudes, tropical Pacific, tropical Africa, and Maritime Continent, and decreases heavy rainfall days over Central America and the tropical and subtropical Atlantic (Fig. 10b).

Despite its dominant influence on global precipitation change, radiative forcing alone does not fully account for the projected regional-scale changes in future precipitation characteristics in $\text{CO}_2_{\text{Total}}$ (Figs. 9 and 10). Within CCSM4, CO_2 physiological forcing plays an important role in shaping the geographic distribution of

projected daily-scale precipitation changes (Figs. 9 and 10). Future increases in the annual number of dry days in the Northern Hemisphere midlatitudes, including Europe and North America, as well as in the Amazon, Andes, and central Africa are largely a response to changes in stomatal conductance (Fig. 9c). These regions experience upward of 15 more dry days per year when future CO_2 -driven changes in stomatal conductance are included in CCSM4. Increases in the annual number of dry days are similar, although slightly more widespread and robust, when using a dry day threshold

Annual Mean Precipitation Change

FIG. 6. Change in annual mean precipitation from (a) CO₂_Total, (b) CO₂_Rad, (c) CO₂_Stomata, and (d) CO₂_LAI. Only statistically significant differences are shown.

of 0.01 mm (as opposed to 0.1 mm), a value closer to the dry day threshold of 0.024 mm used in Lau et al. (2013) (Fig. S6 in the supplemental material). Interestingly, in many regions, the increase in dry day occurrence comes at the expense of very light rainfall or drizzle days (0.1–2 mm) (Fig. 11a), suggesting that decreases in stomatal conductance reduce the likelihood of rainfall on days when atmospheric conditions are only marginally favorable for precipitation. The reductions in light rainfall days are consistent with decreases in low cloud fraction (Fig. 11b). In the midlatitudes, the increase in dry days exceeds the reduction in light precipitation days, suggesting that some dry day increases also come at the expense of heavier rainfall events. Although limited in geographic extent, changes in LAI also impact precipitation characteristics in CCSM4, reducing the number of dry days in parts of East Africa, the Sahel, northern Europe, and Southeast Asia (Fig. 9d). However, the geographic pattern of projected LAI change (Fig. S1) does not closely resemble the pattern of LAI-driven dry day change across the globe (Fig. 9d), highlighting the limited direct contribution of LAI changes in driving future dry day frequency in CCSM4.

In the tropics, physiological forcing also contributes to future changes in the frequency of heavy precipitation

events (Fig. 10c). The CO₂-driven reductions in stomatal conductance drive mean annual increases in excess of eight heavy rainfall days per year in large parts of the Andes, western Amazon, central Africa, and the Maritime Continent. In the eastern Amazon and far northeastern South America, projected changes in stomatal conductance are instrumental in reducing the number of future heavy rainfall days.

In a similar set of climate model experiments with HadGEM2-ES (see section 2), the projected combined impact of future CO₂ physiological forcing and CO₂ fertilization yields similar changes in daily precipitation characteristics to our CO₂_Stomata experiment (Figs. 12a,c). Specifically, the number of dry days increases over tropical land and the continental Northern Hemisphere midlatitudes (although dry day changes in North America are located north and west of those in CCSM4) (Fig. 12a). Additionally, as in CCSM4, the annual frequency of heavy rainfall events decreases over the eastern Amazon, and increases over the rest of the world's tropical forests (Fig. 12c). Similarly, in a coupled atmosphere–ocean simulation with CESM1(BGC), changes in daily precipitation characteristics from combined CO₂ physiological forcing and CO₂ fertilization (see section 2) resemble those of CCSM4,

Percent Contribution to Precipitation Change

FIG. 7. Percent contribution to annual mean precipitation change from (a) CO₂_Rad, (b) CO₂_Stomata, (c) CO₂_LAI and (d) the nonlinear interaction term (see section 2).

suggesting that the use of a slab ocean model in our CCSM4 simulations does not significantly impact the results (Figs. 12b,d).

c. Temperature, moisture, instability, and circulation changes

Previous theoretical and climate modeling work has generated considerable insights into the mechanisms that shape the response of precipitation characteristics to CO₂ radiative forcing (Trenberth et al. 2003; O’Gorman and Schneider 2009). Here, we seek to provide a physical explanation for changes in precipitation characteristics associated with future CO₂–vegetation interactions. We focus primarily on the climate system response to future CO₂ physiological forcing, as future CO₂ fertilization does not significantly contribute to future hydrologic changes in our CCSM4 experiments (Figs. 3 and 7).

Changes in annual mean precipitation and daily precipitation characteristics from future physiological forcing are located in or adjacent to regions of projected large ET reduction (cf. Figs. 2, 6, 9 and 10). Diminished ET arising from reduced stomatal conductance decreases the latent heat flux from the surface to the atmosphere warming the global and terrestrial surface by 0.26°

and 0.56°C, respectively (Figs. 13a,b). Warming on land occurs everywhere except in desert regions. Reductions in ET (Fig. 2) and surface latent heat flux (Fig. 13a) drive the greatest warming over tropical forests, the La Plata basin, southeastern Asia, and the continental Northern Hemisphere midlatitudes (Fig. 13b). Reductions in low cloud fraction (Fig. 11b), which are driven by increased atmospheric temperatures and reduced boundary layer relative humidity (Fig. S7 in the supplemental material), likely amplify surface warming in these regions. Warming also peaks in the high latitudes east of Greenland, far from any terrestrial ET and surface latent heat flux change, and is driven in part by increased low cloud cover during the winter season (Fig. 11b) (Pu and Dickinson 2014).

Despite reductions in surface latent heat flux, and hence a source of moisture to the atmosphere, total vertically integrated precipitable water increases considerably over tropical land (excluding eastern South America and the western Sahel), and along eastern Asia (Fig. 13c). As noted by Pu and Dickinson (2014), changes in stomatal conductance drive a negative surface moisture feedback, whereby reduced ET warms the land surface and draws in moisture from nearby oceans. This reallocation of moisture is most prominent over

Seasonal Mean Precipitation Change

FIG. 8. Change in seasonal mean precipitation from CO2_Stomata: (a) December–February, (b) March–May, (c) June–August, and (d) September–November.

western South America, central Africa, the Maritime Continent, and eastern Asia (Fig. 13c). In these regions, warmer temperatures and enhanced moisture convergence lead to increased convective available potential energy (CAPE) and a greater likelihood of heavy precipitation events (Figs. 10c and 13d). The large reduction in precipitation and heavy precipitation days over the eastern Amazon is unique within tropical forest regions. Anomalous westerly moisture flux directed toward the Amazon is partially blocked by the Andes, leading to substantial increases in annual precipitation and heavy precipitation events in the present day semiarid areas of western Ecuador and Peru, at the expense of the Amazon (Figs. 6c, 10c, and 13c). Further, greater Northern Hemisphere warming likely draws the ITCZ and associated moisture northward and away from the Amazon (although the northward ITCZ shift in the Atlantic is weaker than that in the Pacific) (Figs. 8 and 13c), and moisture export from the eastern Amazon toward the Andes and La Plata basin is enhanced (Fig. 13c). This reduction in eastern Amazon precipitation is a consistent response to CO₂ physiological forcing across climate models (Betts et al. 2004; Swann et al. 2016). In the mid- and high latitudes, moisture convergence does not balance the reduced source of ET

to the atmosphere, and mean precipitation, as well as the number of heavy rainfall events, remains constant or decreases slightly (Figs. 6c, 10c, and 13c).

Future physiological forcing increases the frequency of dry days in months when the local or upwind land surface acts as an important source of moisture for rainfall. In the tropics, dry days increase most when the ITCZ and associated moisture convergence are located outside the region of interest. For example, in the northern Congo and eastern Amazon, increases in dry days are concentrated during the dry season, and in particular during the transition from the dry season to the wet season (Figs. 14a,b). As noted by Lee et al. (2012), this is a time when atmospheric conditions are marginal for the development of rainfall and regional terrestrial ET may have an important role in supplying moisture to initiate the rainy season (Fu and Li 2004). Regions of the tropics that exhibit little to no increase in dry days, for instance the Maritime Continent and parts of Southeast Asia, rely less on terrestrial ET for precipitation (van der Ent et al. 2014). Instead, nearby oceanic regions supply nearly all of the moisture that falls as precipitation in these areas.

In the midlatitudes, increases in dry day frequency from physiological forcing are greatest during times of

Annual Mean Change Number of Dry Days

FIG. 9. Change in annual number of days without precipitation from (a) CO₂_Total, (b) CO₂_Rad, (c) CO₂_Stomata, and (d) CO₂_LAI. A day is considered to have no precipitation when the accumulated daily precipitation is less than 0.1 mm. Only statistically significant differences are shown.

the year when large-scale circulation is less conducive for rainfall and the frequency of dry days is already high. For example, in Europe and the South American mid-latitudes, dry days increase most during the summer months, when large-scale moisture advection is relatively low, and transpiration and precipitation recycling ratios are relatively high (Schär et al. 1999; Trenberth et al. 2003) (Figs. 14c,d). Overall, the diminished moisture source (Fig. 2c), as well as warmer temperatures (Fig. 13b) and associated increases in saturation vapor pressure, increase the likelihood of rain-free days when atmospheric conditions are only marginally favorable for precipitation. These results are consistent with a reduction in light precipitation days in many regions of diminished ET (Fig. 11a).

4. Discussion

a. Summary and implications

Previous idealized climate model sensitivity studies have shown that changes in transpiration flux can impact the frequency and intensity of rainfall in vegetated regions (Lawrence et al. 2007; Lee et al. 2012). The

imposed transpiration changes in these experiments were large [in the case of Lee et al. (2012), transpiration flux was completely removed] and not intended to reflect realistic changes in response to past or future climate forcings. Here, we demonstrate that transpiration changes from projected CO₂ physiological forcing are of sufficient magnitude to influence daily and annual precipitation characteristics and drive large regional-scale impacts in climate model simulations of the twenty-first century. Although warming from radiative forcing is the primary driver of future hydrologic changes at the global scale, regionally, physiological forcing can dominate the hydrologic response to changes in CO₂.

The identification of CO₂ physiological forcing as an important driver of daily-scale hydrologic change helps to more fully understand geographic and seasonal patterns of future projected changes in precipitation characteristics. Hydrologic changes from physiological forcing are primarily constrained to regions and times of the year with large transpiration fluxes, such as the tropics and the summer season Northern Hemisphere midlatitudes (Figs. 2, 4, 6, and 8–10). In many densely vegetated tropical regions, specifically those that exhibit high transpiration but rely on oceanic sources of

Annual Mean Change Number of Heavy Rainfall Days

FIG. 10. Change in the annual number of days with heavy precipitation from (a) CO₂_Total, (b) CO₂_Rad, (c) CO₂_Stomata, and (d) CO₂_LAI. Heavy rainfall days are those that exceed the 95th percentile event from the historical control simulation (year 2000 conditions). Only statistically significant differences are shown.

moisture for a large proportion of rainfall, future CO₂ physiological forcing enhances the likelihood of heavy rainfall events through a negative surface moisture feedback, whereby diminished transpiration from reduced stomatal conductance warms the surface and enhances moisture convergence from surrounding regions (Figs. 10 and 13). In these areas, annual ET decreases while annual precipitation increases. By definition, the local precipitation recycling ratio goes down, and these regions become more reliant on external sources of moisture for rainfall.

Further, in the tropics, reduced stomatal conductance and lower transpiration increases the likelihood of dry days during the transition between the dry and wet seasons, when moisture fluxes from the land surface are particularly important for initiating rainfall (Figs. 9 and 14) (Fu and Li 2004). Likewise, in vegetated areas of the midlatitudes, reduced stomatal conductance increases the frequency of dry days, particularly during the summer season when transpiration and precipitation recycling are relatively high (Figs. 9 and 14). Interestingly, the number of summer season dry days increases in midlatitude Europe, South America, and portions of North America despite

increases in summer soil moisture (Fig. S4). The increase in soil moisture reduces soil moisture stress on vegetation, promotes stomatal conductance, and serves as a negative feedback on the original reduction in transpiration from physiological forcing. However, the combined influences of enhanced stomatal conductance from low soil moisture stress and the moderate increase in soil evaporation from excess soil moisture (Fig. 4f) do not balance the much larger reduction in transpiration from the CO₂-induced limitation of stomatal conductance, leading to low relative humidity and an increase in dry days (Fig. 9; see also Fig. S7). In some midlatitude regions, including the central plains of the United States and southeastern Canada, drier soils from future physiological forcing (from temperature, cloud, and rainfall feedbacks) amplify the reduction in stomatal conductance and reduce soil evaporation, leading to more frequent dry day occurrence. Importantly, the changes in precipitation characteristics from future CO₂-vegetation interactions identified in this study are robust across several different climate model configurations (e.g., simulations with slab and dynamic ocean models, simulations where CO₂ physiological forcing and CO₂ fertilization

Annual Mean Change Number of Light Precipitation Days

Annual Mean Change Low Cloud Fraction

FIG. 11. Change in the (a) annual number of days with light precipitation and (b) annual mean low-cloud fraction from CO₂_Stomata. A day is considered to have light precipitation when the accumulated daily precipitation is greater than 0.1 mm and less than 2 mm. Only statistically significant differences are shown in (a). Brown shading indicates fewer days with light precipitation in (a).

are simulated separately and in tandem, and simulations with different CO₂ concentrations) (Figs. 9, 10, and 12).

Our results highlight the influential role of transpiration in shaping daily hydrologic characteristics, particularly in the tropics. Lee et al. (2012) propose that transpiration serves to reduce the variability of daily precipitation throughout tropical forests. In contrast, our findings suggest that the role of transpiration is not the same across all tropical forest regions. For example, in the eastern Amazon, reduced transpiration increases the annual number of dry days, but it also decreases the

annual number of heavy rainfall days (Figs. 9 and 10). We find that the role of transpiration in shaping daily precipitation depends on a number of factors, including regional geography and time of year. In addition to its impact on precipitation, transpiration also serves as an important control on surface runoff. Specifically, reduced transpiration from future CO₂ physiological forcing increases surface runoff, particularly in the tropics (Fig. S4) (Betts et al. 2007; Gedney et al. 2006; Cao et al. 2010). Our results suggest that enhanced runoff from CO₂ physiological forcing may also be a consequence of more frequent heavy rainfall events.

Annual Mean Change Number of Dry Days

Annual Mean Change Number of Heavy Rainfall Days

FIG. 12. Change in the annual number of (a),(b) dry days and (c),(d) days with heavy precipitation from CO₂ Stomata simulations with (left) HadGEM2-A and (right) CESM1-BGC_esmFixClim1 (see section 2). A day is considered to have no precipitation when the accumulated daily precipitation is less than 0.1 mm. Heavy rainfall days are those that exceed the 95th percentile event from the historical period (first 30 yr). Only statistically significant differences are shown.

b. Uncertainties, caveats, and future work

Although our results point to the important role of CO₂ physiological forcing in shaping projected future precipitation changes in climate model simulations, our findings should be interpreted cautiously. Many uncertainties regarding the interactions among vegetation, carbon dioxide, and hydrology remain (Wang et al. 2009). Most climate models, including those employed in this study, assume that plant water use strategies depend only on the plant's photosynthetic pathway (classified as C₃ or C₄) (Kala et al. 2016). However, observations indicate that plants within C₃ and C₄ groupings exhibit large species-specific differences in stomatal behavior (Lin et al. 2015). Kala et al. (2016) find substantial differences in future climate projections between climate model simulations that use PFT-level water use strategy parameterizations and simulations that use traditional

photosynthetic pathway-level parameterizations. Specifically, they find that the implementation of PFT-level water use strategies leads to a greater reduction in ET over Europe in response to future elevated CO₂, suggesting that our simulations may underestimate the regional hydrological impact of future CO₂ physiological forcing there (Kala et al. 2016). Further, our model experiments do not include the impacts of enhanced CO₂ concentrations on down-regulation of photosynthesis, which Pu and Dickinson (2014) have shown can substantially reduce stomatal conductance and transpiration within CCSM4.

Earth system models project widespread increases in LAI under elevated greenhouse gas concentrations (Mahowald et al. 2016). However, by the end of the twenty-first century, global mean LAI changes vary by a factor of about 14 across models under the RCP8.5 scenario (Mahowald et al. 2016). Relationships between

FIG. 13. Change in annual mean (a) surface latent heat flux, (b) surface temperature, (c) vertically integrated atmospheric moisture and moisture flux, and (d) CAPE from CO2_Stomata.

projected changes in LAI and changes in vegetation carbon or climate variables are not straightforward and the impact of projected LAI differences on the simulated spread in twenty-first-century precipitation projections across models is unclear. Although the hydroclimate impacts from projected LAI in our CCSM4 simulation are minor, future work should assess the contribution of LAI to hydrologic change within the full suite of CMIP climate models. Additionally, our simulations do not include dynamic vegetation models, and therefore do not capture the potential impacts of future changes in vegetation type or location from climate change or land use change on ET and precipitation, which may be large (Notaro et al. 2007; Pearson et al. 2013; Bathiany et al. 2014).

Last, it is important to note that climate models struggle to properly simulate the full distribution of observed precipitation rates (Lin et al. 2006; O’Gorman

and Schneider 2009). In particular, models tend to overestimate the annual number of days with very light precipitation (Lin et al. 2006). A proportion of the dry day (with less than 0.1 mm of precipitation) increases found in our simulation come at the expense of days with very light rainfall (days with 0.1–2 mm of precipitation). Although we provide a mechanistic explanation for the changes in rainfall distribution in CCSM4 [and find consistent results in HadGEM2-A and CESM1(BGC)], the robustness of our results should be tested with other models, particularly those with higher-resolution, resolved convection, and different land surface schemes.

5. Conclusions

Theoretical and climate modeling studies project a robust change in global precipitation characteristics in response to future increases in CO₂ (Trenberth et al. 2003;

FIG. 14. Box-and-whisker plots represent the number of dry days (<0.1 mm) for each month. Lines represent the average daily precipitation. Red (blue) coloring indicates data from a simulation with (without) future changes in stomatal conductance from CO_2 . Box-and-whisker plots show min, max, and interquartile range, and consist of 30 yr of data. Average daily precipitation is averaged over 30 years. Area weighted averages from (a) eastern Amazon (10°S – 0° , 65° – 50°W), (b) northern Congo (0° – 10°N , 15° – 30°E), (c) Europe (40° – 55°N , 0° – 15°E), and (d) midlatitude South America (30° – 40°S , 65° – 55°W).

O’Gorman and Schneider 2009; Lau et al. 2013). Given the potential for acute impacts on human and natural systems, understanding the physical mechanisms that drive these hydrologic changes is critical. To date, most studies have focused on the role of radiative forcing in driving these hydrologic changes. Our modeling results suggest that consideration of the radiative impacts of CO_2 alone cannot account for projected regional-scale differences in daily precipitation changes. Within our simulations with CCSM4, we find that projected CO_2 physiological forcing substantially reduces future transpiration in vegetated areas such as the midlatitudes during the summer season and the tropics. The reductions in transpiration are of sufficient magnitude to drive changes in the annual frequency of dry days and heavy rainfall days in these regions. The physiologically

driven changes have the potential to reinforce or greatly diminish the influence of radiative forcing on future precipitation characteristics. These results suggest that the representation of plant water use and its relationship to atmospheric CO_2 within climate models likely contributes to the persistent spread in future regional-scale precipitation projections. Given the potential for future CO_2 physiological forcing to shape projected changes in precipitation characteristics, future work should aim to improve the representation of the complex interactions among CO_2 , vegetation, and hydrology within climate models in order to refine estimates of hydrology and daily dry and wet extremes.

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