Future Intensification of the Water Cycle with an Enhanced Annual Cycle over Global Land Monsoon Regions

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(Manuscript received 21 September 2018, in final form 9 May 2019)

ABSTRACT

An integrated picture of the future changes in the water cycle is provided focusing on the global land monsoon (GLM) region, based on multimodel projections under the representative concentration pathway 8.5 (RCP8.5) from phase 5 of the Coupled Model Intercomparison Project (CMIP5). We investigate the reservoirs (e.g., precipitable water, soil moisture) and water fluxes (e.g., precipitation P , evaporation E , precipitation minus evaporation $P - E$, and total runoff) of the water cycle. The projected intensification of the water cycle with global warming in the GLM region is reflected in robust increases in annual-mean P (multimodel median response of $0.81\% \text{ K}^{-1}$), E ($0.57\% \text{ K}^{-1}$), $P - E$ ($1.58\% \text{ K}^{-1}$), and total runoff ($2.08\% \text{ K}^{-1}$). Both surface ($-0.83\% \text{ K}^{-1}$) and total soil moisture ($-0.26\% \text{ K}^{-1}$) decrease as a result of increasing evaporative demand. Regionally, the Northern Hemispheric (NH) African, South Asian, and East Asian monsoon regions would experience an intensified water cycle, as measured by the coherent increases in P , $P - E$, and runoff, while the NH American monsoon region would experience a weakened water cycle. Changes in the monthly fields are more remarkable and robust than in the annual mean. An enhanced annual cycle (by $\sim 3\% - 5\% \text{ K}^{-1}$) with a phase delay from the current climate in P , $P - E$, and runoff is projected, featuring an intensified water cycle in the wet season while little changes or slight weakening in the dry season. The increased seasonality and drier soils throughout the year imply increasing flood and drought risks and agricultural yields reduction. Limiting global warming to 1.5°C , the low warming target set by the Paris Agreement, could robustly reduce additional hydrological risks from increased seasonality as compared to higher warming thresholds.

1. Introduction

The global water cycle is expected to intensify in a warmer climate in terms of its cycling water fluxes (Sun et al. 2007). One aspect of the changes is often described as “wet gets wetter and dry gets drier” in terms of

changes in annual precipitation minus evaporation ($P - E$) (Allen and Ingram 2002; Neelin et al. 2006; Allan and Soden 2007; Chou et al. 2009). There are many ways to quantify “wet” and “dry” for terrestrial systems, including water fluxes (e.g., $P - E$) and reservoirs (e.g., soil water storage). In the long-term climatological mean, when the terrestrial water storage changes negligibly, the $P - E$ flux equals the local rate of production of water resources (i.e., runoff and groundwater flow). This water resource is crucial for humans, and the change of which is usually referred to as “wetting” or

Denotes content that is immediately available upon publication as open access.

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DOI: 10.1175/JCLI-D-18-0628.1

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“drying” by stakeholders. For some other applications that are not related to human use of water resources, the water storage may be more important in indicating the wetness or dryness.

As the climate warms, the enhancement in the existing pattern of $P - E$ has been seen in both observations and generations of global climate models (Held and Soden 2006; Durack et al. 2012). To the first order, this large-scale response of $P - E$ to greenhouse gas (GHG)-induced warming can be accounted for by the simple thermodynamic control due to atmospheric moistening. As the atmospheric water vapor increases under global warming, over climatologically wet regions (with $P > E$) mainly dominated by ascending motion, the mean upward motion induces increases in moisture convergence and enhances $P - E$. In contrast, over climatologically dry regions (with $P < E$) mainly dominated by descending motion, the mean downward motion leads to anomalous moisture divergence and thus negative $P - E$ changes. Thus, this thermodynamic effect tends to enhance the existing $P - E$ pattern in a warmer climate (Chou et al. 2009; Chou and Lan 2012). Additionally, this thermodynamic contribution is more robust among climate models than the dynamic contribution related to circulation changes (Chou et al. 2009; Chou and Lan 2012). However, this simple thermodynamic control is ocean-dominated and does not hold over land (Greve et al. 2014; Byrne and O’Gorman 2015; Greve and Seneviratne 2015). Other mechanisms have been proposed to explain the decreasing precipitation over the dry subtropics (the current dry region) that contributes to the “dry gets drier” part (Dai et al. 2018).

The annual variation of precipitation and associated global water cycle (Trenberth et al. 2007) is greatly affected by the global monsoon system, a planetary-scale circulation system driven and maintained by the seasonal cycle of solar radiation (Trenberth et al. 2000; Wang and Ding 2006, 2008; Wang et al. 2012). The global land monsoon (GLM) regions are characterized with abundant monsoon rainfall, a distinct wet–dry season contrast, and an active water cycle. In addition, monsoon rainfall plays an important role in shaping the distribution of freshwater resources across the globe and sustains nearly two-thirds of the world’s population.

Future changes of the monsoon system, especially monsoon precipitation and circulation, have been studied extensively. As the climate warms, the global monsoon area, total amount, and intensity of monsoon precipitation are projected to increase robustly, mainly due to increases in moisture convergence (Hsu et al. 2012, 2013; Kitoh et al. 2013). Regionally, overall it is projected that the Northern Hemisphere monsoon would gain more precipitation than that in the Southern

Hemisphere, while the Eastern Hemisphere monsoon would get more precipitation than that in the Western Hemisphere, primarily due to the increased land–ocean and interhemispheric thermal gradients (Lee and Wang 2014). Specifically, the South and East Asian summer monsoon precipitation would increase significantly (Kitoh et al. 2013; Wang et al. 2014). In contrast, a robust reduction in monsoonal precipitation is projected for the Northern Hemisphere American monsoon region, largely caused by the increased atmospheric stability and weakened convection, due to the sea surface warming patterns (Pascale et al. 2017).

In addition to the changes in the annual-mean and monsoonal precipitation, changes in the annual cycle can also have important effects. An intensification of the annual range of precipitation on both global and regional scales has been observed for the recent decades and projected for the twenty-first century (Chou and Lan 2012; Chou et al. 2013; Huang et al. 2013; Dwyer et al. 2014). There is also high consistency among CMIP5 models in a projected phase delay of the annual cycle of tropical precipitation dominated by changes in the seasonality of circulation (Dwyer et al. 2014). From an energetics perspective, under global warming, more moist static energy is required to be added to the atmosphere in spring to reach summertime peak levels, as required by the Clausius–Clapeyron relation. As a result, the winter-to-summer transition in each hemisphere takes longer time, leading to a seasonal delay in the migration of tropical precipitation (Song et al. 2018). For the monsoon regions, specifically, the amplified annual range of precipitation for the African and Asian–Australian monsoon regions, and redistribution of rainfall from the early to late rainy season for the American and African monsoons have also been reported in multimodel projections (Lee and Wang 2014; Seth et al. 2011, 2013). The latter is due to the enhanced convective barrier in the transition season under a warmer climate, which is related to the low-level moisture content (Seth et al. 2011, 2013).

However, limited attention has been paid to other aspects of the water cycle in the monsoon regions. The water cycle refers to the movements of water within the whole Earth system. It incorporates other components including atmospheric water storage [i.e., precipitable water (PW)] and moisture flux for the atmospheric water budget, and subsurface water storage (including soil moisture and groundwater) and runoff for the land water budget (Trenberth and Fasullo 2013). A comprehensive understanding of the responses of the water cycle is critical to anticipate impacts on water availability, agriculture, and ecosystems, as well as hydrological extremes.

The specific changes in the water cycle at some target warming levels are of particular concern to decision makers and the public. The 2015 Paris Agreement has set an ambition of “holding the increase in the global-average temperature to well below 2°C above preindustrial levels and pursuing efforts to limit the temperature increase to 1.5°C, recognizing that this would significantly reduce the risks and impacts of climate change” (UNFCCC 2015). Significant efforts since then have been devoted to understanding different climatic impacts between the two warming levels, covering weather and climate extremes, water availability, agricultural yields and sea level rise, etc. (Sedláček and Knutti 2014; Schleussner et al. 2016a,b; Mitchell et al. 2016; Hulme 2016; Huang et al. 2017; King et al. 2017; Li et al. 2018; Nangombe et al. 2018; Zhang et al. 2018). Our current knowledge in the hydroclimatic impacts at these target warming thresholds is still limited, particularly for the densely populated GLM region.

This study aims to provide a comprehensive picture as well as a physical understanding of the future water cycle changes for the GLM region. Specifically, we address the following scientific questions:

- 1) How does the annual-mean water cycle respond to GHG-induced global warming in the GLM region?
- 2) How does the seasonality of the water cycle change?
- 3) What are the specific changes in the various aspects (P , E , $P - E$, runoff, and soil moisture) of the water cycle at a global warming of 1.5°C and higher thresholds?

These issues have significant implications for water stress, agriculture, ecosystems, and flood and drought risks management.

The data and methods are described in section 2. The response of the annual-mean water cycle in the GLM region to global warming is investigated in section 3. Section 4 describes the changes in the annual cycle of the water cycle. Changes at specific warming levels are compared in section 5, followed by a discussion and the conclusions in section 6.

2. Data and methods

a. Observational and reanalysis datasets

To assess the model ability in simulating hydroclimate in the monsoon regions, the model-simulated P and $P - E$ are validated against observations or reanalysis datasets. To account for uncertainties in observations, two observational global monthly precipitation datasets are employed, including 1) the Global Precipitation Climatology Project with a spatial resolution of $2.5^\circ \times 2.5^\circ$ (GPCP;

Adler et al. 2003) and 2) Full Data Product from the Global Precipitation Climatology Centre with a spatial resolution of $1^\circ \times 1^\circ$ (GPCC V7; Schneider et al. 2014). These two datasets are considered more reliable in the precipitation variations in the recent decades than other precipitation observations including the Climatic Research Unit (CRU; Harris et al. 2014), University of Delaware (UDEL; Willmott and Matsuura 2001), and NOAA's Precipitation Reconstruction over Land (PREC/L; Chen et al. 2002), due to the large declines in the rain gauge numbers since the 1990s in the latter datasets (Dai 2011; Dai and Zhao 2017; Sun et al. 2018).

Particularly, $P - E$ is estimated from the moisture convergence in reanalysis datasets for verification, as it is considered more reliable and accurate than that obtained with individual values of P and E in reanalyses (Trenberth et al. 2011; Sebastian et al. 2016). We acknowledge that while the $P - E$ and moisture convergence are generally balanced in terms of annual mean, they do not have to exactly equal each other on the seasonal time scale, because the atmospheric moisture storage tendency term (related to the progress of the seasonal cycle) can be significant. The importance of seasonal variations can be seen in section 4 of this paper as well as in previous works (e.g., Song et al. 2018). Four reanalysis datasets are used: 1) National Centers for Environmental Prediction–U.S. Department of Energy AMIP-II reanalysis (NCEP2; Kanamitsu et al. 2002); 2) European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011); 3) Japanese 55-yr Reanalysis Project (JRA-55; Kobayashi et al. 2015); and 4) Modern-Era Retrospective Analysis for Research and Applications (MERRA) conducted by the National Aeronautics and Space Administration (NASA) Goddard Space Flight Center (GSFC) (Rienecker et al. 2011). The present-day climatology of 1986–2005 is evaluated.

b. CMIP5 model simulations

We employ multimodel simulations from the CMIP5 archive (Table 1; Taylor et al. 2012). Based on monthly data availability, including the near-surface air temperature, precipitation, evaporation, specific humidity, wind fields, total runoff, and surface and total soil moisture, the historical simulations and projections under the high-end emission scenario of representative concentration pathway 8.5 (RCP8.5) from 27 models are analyzed. Only the first realization from each model is used. Historical simulations of 1986–2005 are used as a common baseline. All model output is interpolated onto a common $2.5^\circ \times 2.5^\circ$ resolution.

Specifically, the changes in $P - E$ are investigated from model simulations of P and E . For the model

TABLE 1. List of the CMIP5 models used and the timings of the 1.5°, 2°, 3°, and 4°C global warming under the RCP8.5 emission scenario. A dash (—) indicates that the warming level will not be achieved by 2100.

Model	1.5°C	2°C	3°C	4°C
ACCESS1.0	2025	2042	2061	2081
ACCESS1.3	2031	2042	2061	2081
BCC_CSM1.1(m)	2012	2028	2061	2086
BCC_CSM1.1	2021	2036	2060	2085
BNU-ESM	2010	2023	2047	2065
CanESM2	2013	2027	2049	2068
CCSM4	2015	2029	2058	2077
CESM1(BGC)	2017	2034	2059	2081
CESM1(CAM5)	2029	2040	2057	2078
CNRM-CM5	2029	2043	2067	2087
CSIRO Mk3.6.0	2035	2043	2065	2082
FGOALS-g2	2031	2047	2076	—
GFDL CM3	2025	2036	2055	2071
GFDL-ESM2G	2039	2056	2081	—
GFDL-ESM2M	2037	2051	2081	—
GISS-E2-H	2020	2037	2066	—
GISS-E2-R	2032	2052	2086	—
INM-CM4.0	2045	2058	2085	—
IPSL-CM5A-LR	2012	2028	2047	2066
IPSL-CM5A-MR	2016	2032	2051	2068
IPSL-CM5B-LR	2024	2038	2061	2086
MIROC5	2035	2050	2073	—
MIROC-ESM-CHEM	2019	2029	2049	2067
MIROC-ESM	2021	2030	2053	2070
MRI-CGCM3	2041	2053	2077	—
NorESM1-ME	2032	2047	2069	2092
NorESM1-M	2033	2049	2073	—

evaluation, both $P - E$ computed directly and from moisture convergence are compared, to be consistent with reanalysis results (i.e., Fig. 1).

c. Definition of global land monsoon domain

The global monsoon domain is defined based on precipitation characteristics, as the area where the local summer-minus-winter precipitation rate exceeds 2.0 mm day^{-1} and the local summer precipitation exceeds 55% of the annual total (Wang et al. 2012). Here, the local summer refers to May–September for the Northern Hemisphere and November–March for the Southern Hemisphere, and vice versa. In this study, the global land monsoon region is determined based on the 1986–2005 climatological precipitation from GPCP (Adler et al. 2003; see Fig. 4). For regional divisions, the equator separates the Northern Hemisphere (NH) from the Southern Hemisphere (SH) monsoon region. The NH monsoon domain is separated into the NH African, South Asian, East Asian, and NH American monsoon domains; while the SH monsoon domain is divided into the SH African, Australian, and SH American monsoon domains. We also separate the Eastern Hemisphere (EH)

from the Western Hemisphere (WH) monsoon region at 30°W and 180° . We set the boundary at 30°W instead of 0° to avoid artificially split the NH African monsoon region in two (see map in Fig. 4 for regional divisions).

d. Atmospheric and surface water budget

The atmospheric water budget can be expressed as (Seager et al. 2010; Chou and Lan 2012; Trenberth and Fasullo 2013)

$$P - E = -\frac{\partial \text{PW}}{\partial t} - \langle \nabla \cdot \mathbf{V}q \rangle, \quad (1)$$

$$P - E = -\frac{\partial \text{PW}}{\partial t} - \langle \omega \partial_p q \rangle - \langle \mathbf{V} \cdot \nabla q \rangle, \quad (2)$$

where P is precipitation, E is evaporation, PW is precipitable water (i.e., vertically integrated atmospheric moisture), $\mathbf{V}$ is wind vector, q is specific humidity, and the angle brackets denote a vertical integration through the entire troposphere. Thus, $-\langle \nabla \cdot \mathbf{V}q \rangle$ represents the total moisture convergence, which can be divided into two parts: the vertical moisture advection $-\langle \omega \partial_p q \rangle$ and horizontal moisture advection $-\langle \mathbf{V} \cdot \nabla q \rangle$. The former is related to vertical motion or low-level convergence while the latter is associated with horizontal velocity.

Changes in precipitation are balanced by those in evaporation and moisture convergence, neglecting the small changes in the tendency term. The changes in the moisture convergence, which is shown to be dominated by its vertical component, can be further separated into two parts:

$$P' - E' = -\langle \omega \partial_p q' \rangle - \langle \mathbf{V} \cdot \nabla q' \rangle, \quad (3)$$

$$-\langle \omega \partial_p q' \rangle \approx -\langle \overline{\omega} \partial_p \overline{q}' \rangle - \langle \omega' \partial_p \overline{q}' \rangle, \quad (4)$$

where the overbar represents the baseline climatology and the prime represents the departure from the climatology. Thus the changes in the vertical moisture advection are contributed by a thermodynamic (TH) term, which is related to changes in water vapor mainly induced by temperature changes, and a dynamic (DY) term, which is associated with changes in circulation. Here we neglect the nonlinear term $-\langle \omega' \partial_p \overline{q}' \rangle$.

The surface water budget is (Trenberth and Fasullo 2013)

$$\frac{\partial S}{\partial t} = P - E - R, \quad (5)$$

where S is the subsurface storage of water substance, which is referred to as total soil moisture in our analysis since groundwater is not available in climate models, and R is the total runoff.

FIG. 1. The 1986–2005 climatological annual cycle of precipitation P and precipitation minus evaporation ($P - E$) in (a) the NH land monsoon region and (b) the SH land monsoon region for multiple observations and reanalysis datasets (colored lines) and multimodels. Black lines indicate multimodel medians while dark (light) shading indicates the 25th–75th (10th–90th)-percentile model ranges. The value of $P - E$ is estimated from the atmospheric moisture budget in reanalysis datasets. In the models, both $P - E$ computed from separate simulations of P and E (Model $P - E$; dashed black curves) and from moisture convergence (Model CONV; solid black curves) are shown. The model uncertainty ranges are similar for these two variables and only those for Model CONV are shown.

e. Response of the water cycle to global warming

To investigate the sensitivity of the water cycle to global warming, specifically its reservoirs (e.g., soil moisture) and fluxes (e.g., P , E , $P - E$, and runoff), the water cycle components and global-mean surface air temperature are first averaged over decadal periods to eliminate interannual variability. Projections starting from 2006 are averaged in decades with an overlapping of 5 years, that is, the averages of 2006–15, 2011–20, up to 2091–2100. A linear regression between the smoothed water cycle components and temperature is referred to as the response rate to global warming. In this manuscript, the changes in water cycle components are expressed as changes per degree of global warming (i.e., Figs. 2–7 and 9).

f. Annual range

The intensity of annual cycle is measured by annual range, defined as the difference between the maximum and minimum monthly value in a year (Chou and Lan 2012). By this definition, the months with maximum and minimum value vary from year to year and from region to region. It identifies the exact range within a year, regardless of the potential shifts of wet and dry seasons under global warming.

g. Definition of specific warmer worlds

To account for future deviations from the current climate, a common reference period of 1986–2005 is referred to as the baseline scenario. The timings of 1.5°, 2°, 3°, and 4°C warming above the preindustrial level (1861–90) are determined using the 9-yr running-mean, global-mean surface air temperature for each model separately (Table 1). The specific warming scenarios are represented by the 20-yr windows centered on the years when respective warming occurs. The climate changes in specific warmer worlds are calculated for each model separately before deriving multimodel ensembles.

3. Response of the annual-mean P , E , $P - E$, runoff, and soil moisture to global warming

To assess the ability of CMIP5 models in simulating the current hydroclimate, the climatological annual cycles of P and $P - E$ are compared between multiple observations and reanalyses and models (Fig. 1). These two variables are selected for verification because they are fundamental to the integrated water cycle in both the atmosphere and the surface, and that global observations are more reliable compared to other water cycle components. The CMIP5 models basically reproduce

FIG. 2. Multimodel medians of annual-mean responses of (a) P , (b) E , (c) $P - E$, (d) total runoff, (e) surface soil moisture, and (f) total soil moisture to global-mean surface air temperature increases under RCP8.5. To calculate the responses, projections are averaged over decadal periods starting in 2006 and overlapped by 5 years (i.e., 2006–15, 2011–20, up to 2091–2100). The dots indicate where at least two-thirds of the models agree on the sign of response. Red lines denote the global land monsoon region.

the annual cycles of P and $P - E$ for both the NH and SH monsoon regions, with wet local summer and dry local winter (Fig. 1). However, there are slightly dry biases (approximately -1 mm day^{-1}) in simulated P and $P - E$ in the wet season for the NH monsoon region (Fig. 1a). The computation methods of $P - E$, that is, computing directly or from moisture convergence, do not affect the model results prominently (cf. solid and dashed black curves in the bottom panels of Fig. 1), as climatologically the atmospheric moisture tendency term on the monthly time scale is relatively small compared to $P - E$ and moisture convergence. The annual cycles of P and $P - E$ over individual monsoon regions are also generally reproduced (figure not shown). The annual cycles of P and $P - E$ in the NH and SH monsoon regions revealed here are similar to those in the NH and SH land regions, as depicted in Trenberth et al. (2007). A slight difference is that the annual cycle of $P - E$ in the NH land region is weaker compared to that in the NH monsoon region, related to the differences in the low and mid- to high latitudes (Figs. 5 and 7c in Trenberth et al. 2007). The ability of the CMIP5 ensemble to reasonably simulate the water cycle in the monsoon regions increases the reliability of future projections. Thus, we use all available models for further investigation of water cycle changes, in accord with previous studies in this regard (e.g., Seager et al.

2013; Collins et al. 2013; Sedláček and Knutti 2014; Zhao and Dai 2015, 2017).

The sensitivities of water cycle components including P , E , $P - E$, total runoff, and soil moisture to global-mean surface air temperature increase under RCP8.5 are given in Fig. 2. The general patterns of the responses revealed here are consistent with previous studies depicting changes between the future and present-day periods (e.g., Collins et al. 2013; Zhao and Dai 2015, 2017), except that we scale the changes with temperature, which accounts for different global warming rates in model projections. Annual-mean precipitation exhibits robust increases with warming in the South and East Asian and NH African monsoon regions at the rates of $5.16\% \text{ K}^{-1}$ ($3.86\% \text{ K}^{-1}$ – $5.66\% \text{ K}^{-1}$ for the interquartile range), $3.14\% \text{ K}^{-1}$ ($2.53\% \text{ K}^{-1}$ – $4.69\% \text{ K}^{-1}$), and $1.46\% \text{ K}^{-1}$ ($0.82\% \text{ K}^{-1}$ – $2.72\% \text{ K}^{-1}$), respectively, while decreases robustly in the NH American monsoon region by $-2.96\% \text{ K}^{-1}$ (-4.72% to $-1.02\% \text{ K}^{-1}$). Only moderate changes are projected for the SH monsoon region with low model consistency.

The responses of E , $P - E$, and total runoff generally follow that of P (Figs. 2b–d). Total soil moisture also features a similar response pattern, although the sensitivity is much lower (Fig. 2f). In comparison, top-10-cm soil moisture shows larger decreases in a wider spatial range, directly induced by the increasing evaporative demand

Units: kg/m² for storage and mm/yr for exchanges

FIG. 3. The annual-mean water cycle for the GLM region. Black numbers denote the 1986–2005 climatology, while red numbers denote responses to global-mean surface temperature increase under RCP8.5 ($\% \text{ K}^{-1}$). Both the multimodel medians and 25th–75th-percentile intervals are shown. Changes with two-thirds of model agreement are underlined. Changes of moisture budget components are normalized by climatological precipitation.

in a warmer climate (Fig. 2e). Note that the changes in runoff and soil moisture show more regional characteristics in spite of the overall similarity with those in P and $P - E$, because local processes such as vegetation, wind, and snowmelt also have an effect. In addition, the uncertainty in model projections of total soil moisture is higher than other components, due to the diverse soil schemes and vegetation parameterizations in CMIP5 models (Berg et al. 2017).

We specify the changes for the GLM region in Fig. 3. The projected intensification of water cycle with global warming in the GLM region is reflected in robust increases in annual-mean precipitation (0.81 K^{-1} , $0.27\% \text{ K}^{-1}$ – $1.69\% \text{ K}^{-1}$), evaporation (0.57 K^{-1} , $0.23\% \text{ K}^{-1}$ – $0.77\% \text{ K}^{-1}$), and runoff (2.08 K^{-1} , $1.20\% \text{ K}^{-1}$ – $4.45\% \text{ K}^{-1}$). Both surface (-0.83 K^{-1} , -1.60% to $-0.19\% \text{ K}^{-1}$) and total soil moisture (-0.26 K^{-1} , -0.70% – $0.15\% \text{ K}^{-1}$) decrease as a response to the increasing evaporative demand.

The mechanism for the increasing annual-mean precipitation in the GLM region is investigated by the moisture budget analysis [Eqs. (3) and (4); Fig. 3]. The moisture budget components are normalized by the climatological precipitation and scaled by the global-mean temperature increase, thus expressed in units of percent per kelvin for easy comparison with the response in precipitation. The increase in the moisture convergence (i.e., vertical moisture advection) dominates that in the annual-mean precipitation, at a rate of 2.50 K^{-1} ($1.72\% \text{ K}^{-1}$ – $3.28\% \text{ K}^{-1}$). Specifically, the thermodynamic term due to a moister atmosphere in a warmer climate has a contribution of 3.21 K^{-1} (2.60% – $3.49\% \text{ K}^{-1}$), while it is offset partly by the dynamic term related to the weakening of tropical circulation as the climate warms by -0.66 K^{-1} (-0.95% to $0.10\% \text{ K}^{-1}$; Vecchi and Soden 2007). The horizontal moisture advection suppresses precipitation by -0.83 K^{-1} (-1.08% to $-0.64\% \text{ K}^{-1}$). The dominant roles revealed by the moisture budget analysis are

FIG. 4. Responses of water cycle components to global-mean surface air temperature changes under RCP8.5 for submonsoon regions, including P , E , $P - E$, total runoff R , and surface (SMs) and total soil moisture (SMt). Results for the annual mean and wet and dry seasons are shown. The surface and total soil moisture content are scaled by $1/10$ and $1/200$, respectively, for display. The unit is $\text{kg m}^{-2} \text{K}^{-1}$ for soil moisture and $\text{mm day}^{-1} \text{K}^{-1}$ for others. The wet season refers to May–September for the NH and November–March for the SH, and vice versa for the dry season. The histograms denote the multimodel medians while error bars denote the 25th–75th-percentile model intervals. The vertical axis has the same range for submonsoon regions except for the global monsoon region. For regional divisions, the equator separates the NH from the SH monsoon region, 30°W and 180° separate the EH from the WH monsoon region, 60°E separates the NH African from the South Asian monsoon region, and 20°N and 100°E separate the South Asian monsoon region from the East Asian monsoon region. All the regional domains are within 40°S – 45°N .

consistent with previous studies, although they have not been scaled by temperature increases before (Hsu et al. 2012, 2013; Kitoh et al. 2013). Such processes are consistently seen in most of the regional monsoons except for the NH American monsoon region. The annual and monsoon precipitation is projected to decrease in the NH American monsoon region. Different from other submonsoon regions, no significant increase in the vertical moisture advection is seen in this region, because the negative contribution from the dynamic term almost cancels out the increases induced by the thermodynamic term (figure not shown). The negative contribution from the dynamic term, that is, weakened convection, in this region is demonstrated to be related to the projected sea surface warming patterns (Pascale et al. 2017).

In terms of the regional characteristics, there is a distinct hemispheric contrast in the water cycle changes, featuring an intensified water cycle in the NH monsoon region while no robust changes in the SH monsoon

region. Meanwhile, the EH monsoon region exhibits an intensification while the WH monsoon region shows little or slight weakening in the water cycle. This is supported by coherent changes in P , $P - E$, and total runoff (Fig. 4). Specifically, the NH African, South and East Asian monsoon regions would experience an intensified water cycle in terms of P , $P - E$, and runoff, while the NH American monsoon region would experience a weakened water cycle. The SH American, SH African, and Australian monsoon regions show moderate changes with low model consistency except for the drying soil, partly due to their opposing responses within the region (Fig. 2).

We also note that the annual-mean changes in total runoff exceed that in $P - E$ for the global monsoon region as a whole (Fig. 4). In the annual long-term mean, $P - E$ should be balanced with total runoff according to the surface water budget, regardless of changes in the terrestrial water storage [dS/dt ; Eq. (5)]. The imbalanced

FIG. 5. (a) Atmospheric moisture budget terms concerning P , E , $P - E$, vertically integrated moisture convergence (CONV), tendency of atmospheric precipitable water (dPW/dt), and precipitable water (PW) for the climatology of 1986–2005 for the Northern Hemisphere monsoon region. In (a), the right plot shows annual means (bars) and the left plot shows departures from annual means. The shadings denote the 25th–75th-percentile model intervals. PW is scaled by 1/20 for display. The unit is kg m^{-2} for PW and mm day^{-1} for others. (b) Responses to global-mean temperature changes under RCP8.5 for the NH monsoon region. (c),(d) As in (a) and (b), respectively, but for the SH monsoon region.

changes of $P - E$ and total runoff projected by the CMIP5 ensemble may be related to problems with spinup or with water balance closure in the models, the biases of which also change over time (Liepert and Previdi 2012; Liepert and Lo 2013). On the other hand, regional terrestrial water storage may change under a rapidly warming climate, through glaciers melting, lakes growing, etc. Such changes vary regionally, and have already been observed by the Gravity Recovery and Climate Experiment (GRACE) satellites in the recent decades (Rodell et al. 2018). Nevertheless, how the terrestrial water storage would change in the global monsoon region under a changing climate deserves dedicated research, which is beyond the scope of the current study.

4. Enhanced seasonality in P , $P - E$, and runoff

As precipitation is the fundamental and determining factor in the water cycle, which exhibits a distinct wet–dry season contrast in the monsoon region, it is necessary to examine the water cycle changes in the wet and dry seasons separately. Figure 4 also displays the projected responses of water cycle components (viz., P , E , $P - E$, runoff, and soil moisture) to global warming for the wet

and dry seasons. Annual-mean responses are dominated by those of the wet season. This is understandable because for the monsoon regions, monsoon rainfall dominates the annual total (Fig. 1). On the other hand, the responses in the dry season can be quite different from those in the wet season in some cases, for example, the changes of $P - E$ in the Asian, Australian and SH African monsoon regions (Fig. 4). In the next we examine the changes in the annual cycle to give further insights into the water cycle, which can exert greater impacts than annual mean.

a. Atmospheric water budget

We first examine the annual cycle of atmospheric moisture budget components following Eq. (1). Climatologically, the distinct annual cycle of P affects the annual cycles in E and $P - E$, the latter of which is closely associated with total moisture convergence (Figs. 5a,c). The atmospheric moisture storage tendency features a positive value from the late winter to mid-summer because moisture convergence and evaporation keep moistening the troposphere. In the remaining of the year (i.e., the transition from the wet season to the dry season), there is a drying tendency in atmospheric moisture. The annual mean of moisture tendency is zero

FIG. 6. As in Fig. 5, but for the annual cycle of the surface water cycle, concerning P , $P - E$, and total runoff (Runoff).

by design. Generally the atmospheric moisture budget is reasonably closed in the GLM region in the models. The annual-mean residual is 5%–9% with respect to annual precipitation.

The responses of the annual cycle to global warming are shown in Figs. 5b and 5d. The changes in P and $P - E$ feature an enhanced seasonality. Specifically, P and $P - E$ increase prominently in the wet season for both hemispheric monsoons. On the other hand, P and $P - E$ change little for the NH monsoon region in the dry season, while they decrease in the transition from the dry to wet season for the SH monsoon region (Figs. 5b,d). Such seasonal contrast is dominated by the thermodynamic contribution of the vertical moisture advection related to increasing atmospheric moisture, which enhances precipitation in the wet season and suppresses it in the dry season (Chou and Lan 2012). The seasonality of P and $P - E$ changes indicates that the intensification of water cycle in the monsoon regions mainly occurs in the wet season, while that in the dry season changes little or slightly weakens. We also note that the increase in monsoon precipitation is most prominent in October in the NH and February in the SH toward the end of the monsoon season, which indicates a phase delay in the annual cycle and a delay in monsoon retreat. This result is consistent with the projected seasonal delay of tropical rainfall (Song et al. 2018). The persistence of the rain belt in the NH monsoon regions in September–October directly leads to the sharpest rainfall reduction in the SH in

September–October, as they are inherently tied to the meridional movement of the intertropical convergence zone (Wang et al. 2017). Changes in the $P - E$ annual cycle are again associated with the similar variation in moisture convergence (light blue curves in Fig. 5). Atmospheric precipitable water increases throughout the year, with a peak in the monsoon and a valley in the dry season (black curves in Figs. 5b and 5d), consistent with changes in its tendency term (yellow curves in Figs. 5b and 5d). This indicates that the atmospheric water storage would experience an enhanced wet–dry season contrast. However, the changes in evaporation are similar throughout the year. Hence the atmospheric water cycle including P and $P - E$, the latter of which is closely associated with total moisture convergence, exhibits a more distinct wet–dry season contrast with a phase delay in the global monsoon region as the climate warms.

b. Surface water budget

The land water cycle components are investigated based on Eq. (5) including P , E , $P - E$, runoff, and soil moisture, in Figs. 6 and 7. The surface water budget is well closed in models with negligible annual-mean residual (see bars in Figs. 6a,c). Climatologically, the seasonal variation of total runoff follows that of P and $P - E$ (Figs. 6a,c). Following the enhanced annual cycles of P and $P - E$, the changes of total runoff also show a peak in the monsoon season and a valley in the dry

FIG. 7. As in Fig. 5, but for the annual cycle of the surface water cycle, concerning P , $P - E$, SMs, SMt, and tendency of total soil moisture content $dSMt/dt$. The surface and total soil moisture content are scaled by 1/5 and 1/50, respectively, for display. The unit is kg m^{-2} for soil moisture and mm day^{-1} for others.

season (Figs. 6b,d). The enhanced annual cycles in P and runoff indicate an increase in the likelihood of flooding in the monsoon season.

As soil moisture is an important proxy for drought and relates directly with agriculture (Dai 2013; Trenberth et al. 2014), Fig. 7 further depicts the annual cycle of soil moisture. Climatologically, the annual cycle of soil moisture follows that of P and $P - E$, but with several months of lag, reflecting the influence of precipitation and evaporation on soil moisture (Eltahir 1998; Seneviratne et al. 2010; Zhao and Dai 2015). We examined the lead-lag relationship between soil moisture and P and $P - E$. The results show that surface soil moisture is influenced by P and $P - E$ with a maximum lag of 2 months (Fig. 8a). The highest correlation occurs simultaneously, as the surface layer of soil moisture responds immediately to precipitation anomalies. For the total soil moisture, however, the influence of P and $P - E$ reaches farther to 9 months. The highest response occurs 1 month after precipitation anomalies, due to the gradual infiltration of precipitation into the deep soil layer (Fig. 8b). Hence in climatology, the total soil moisture content (SMt) peaks toward the end of the monsoon season, associated with a positive tendency ($dSMt/dt > 0$) during the entire monsoon season driven by the accumulation of monsoon rainfall and net infiltration into the soil (Figs. 7a,c). It reaches its valley in March–April for NH

and September–October for SH, when the dry season ends and the monsoon season is about to start. The excess of evaporation over precipitation in the preceding dry season forces the soil to lose moisture (Figs. 7a,c).

The annual cycle of soil moisture changes is weak (Figs. 7b,d). The surface soil moisture decreases steadily throughout the year for both hemispheric monsoon regions. The total soil moisture decreases robustly in the SH monsoon region, but changes insignificantly in the NH monsoon region due to large model uncertainty (Figs. 7b,d). Note that soil water deficiency is an indicator of agricultural droughts, which can lead to crop yields reductions.

We also note that for the SH monsoon region, the annual-mean changes in water cycle components including P , $P - E$, and runoff are small in magnitude with low model consistency (Fig. 4). However, the monthly variations are quite consistent among models, resulting in more robust projections, and thus confirming the necessity and importance of investigating the annual cycle.

Finally, we quantify the enhanced annual cycle using annual range as a measure (Fig. 9). Over the GLM region, the annual range of P , $P - E$, and total runoff would intensify by 3%–5% per 1 K of global warming (Fig. 9b). This is dominated by the larger increases in the maximum month (wet season) and negligible or

FIG. 8. Lagged correlation coefficients between (a) SMs and (b) SMt and P and $P - E$ over the GLM region. The dashed lines denote the 5% significance level. The shading indicates the 25th–75th-percentile model intervals.

weak decreases in the minimum month (dry season) (Figs. 9c,d). The evaporation shows a slight increase in annual range of around $1\% \text{ K}^{-1}$, influenced by that in precipitation. The changes in the annual range of soil moisture exhibit considerable uncertainty, although the multimodel medians indicate an enlarged annual range, primarily due to larger decreases in the dry season than the wet season.

To conclude, we have identified an intensified annual cycle of precipitation, which would further contribute to an enhanced seasonality in different aspects of the water cycle including $P - E$ and total runoff, and have far-reaching implications.

5. Changes in P , E , $P - E$, runoff, and soil moisture at 1.5°C and higher warming levels

Water cycle changes at specific warming levels are of great concern to both decision-makers and the public, particularly regarding the 1.5°C low warming target and higher thresholds. We examine different aspects of the water cycle in the GLM region, including P , E , $P - E$, total runoff, and soil moisture, at the 1.5° , 2° , 3° , and 4°C warming levels above the preindustrial level for the

annual mean and wet and dry seasons (Fig. 10). For the low signal-to-noise ratio case of a 1.5°C warming, multimodels project moderate changes in annual-mean precipitation, evaporation, $P - E$, and total runoff relative to the present-day baseline with low model agreement, except for the robust decreases in surface soil moisture (Fig. 10a). However, the wet–dry season contrast in water fluxes changes already emerges (Figs. 10b,c). The seasonal-mean P ($P - E$) is projected to increase by 1.22% (1.56%) in the wet season and decrease by -1.80% (-3.80%) in the dry season by the best estimate at the 1.5°C warming future. The surface soil moisture would decrease by -0.79% annually.

Higher warming levels generally amplify the changes in a 1.5°C warmer world. A 2° , 3° , and 4°C warming would increase P ($P - E$) in the wet season by 1.28%, 2.70%, and 4.56% (2.05%, 3.94%, and 5.33%), respectively, and decrease P ($P - E$) in the dry season by -2.23% , -2.67% , and -2.74% (-4.83% , -7.52% , and -9.49%), respectively, relative to the present-day level. Annual-mean soil moisture in the surface layer would decrease by -1.56% , -2.20% , and -2.93% , respectively, at the 2° , 3° , and 4°C warming levels. Thus, further warming above 1.5°C would lead to robust additional changes in water cycle in the GLM region, posing extra flood risks in the wet season and agricultural drought risks throughout the year.

6. Conclusions

We have investigated the response of the water cycle, specifically its reservoirs (e.g., atmospheric water storage, soil moisture) and fluxes (e.g., P , E , $P - E$, and total runoff), in the global monsoon regions to global warming using CMIP5 multimodel projections under the high-emission scenario of RCP8.5. On the annual time scale, the global land monsoon region as a whole, as well as the NH African and South and East Asian monsoon regions, would experience an intensified water cycle, manifested by coherent increases in P , $P - E$, and runoff, while the NH American monsoon region would experience a weakened water cycle. The SH American, SH African, and Australian monsoon regions show moderate changes with low model consistency except for the drying soil, partly due to their opposing responses in different parts of the region. The annual-mean responses generally follow those in the wet season, primarily due to the dominant role of monsoon precipitation in the annual total.

Changes on the monthly or seasonal time scales are more remarkable and significant than in the annual mean. An intensified annual cycle of precipitation along with a phase delay is identified. It would further

FIG. 9. Responses of the annual range of the water cycle components to global-mean temperature changes over the GLM region. (a) Climatological annual ranges of P , E , $P - E$, R , SMs , and SMt . SMt is scaled by 1/15 for display. The unit is kg m^{-2} for soil moisture and mm day^{-1} for others. The multimodel medians (histograms) and 25th–75th-percentile intervals (error bars) are shown. (b) The responses of the annual range of the water cycle components to global-mean temperature changes, shown in absolute values (blue; left axis) and percentages relative to the climatology (red; right axis). (c),(d) As in (b), but for the responses of the monthly (c) maxima and (d) minima of water cycle components to global-mean temperature changes.

contribute to an enhanced annual cycle (by around 3%–5% K^{-1}) with a phase delay from the current climatology in $P - E$ and total runoff, the former of which is closely associated with total moisture convergence. Thus the wet–dry season contrast would amplify, featuring an intensified water cycle in the wet season while little changes or slight weakening in the dry season. This has far-reaching implications. The peak increases in $P - E$ and total runoff in the monsoon season would contribute to flooding conditions and thus indicate increases in the probability of flooding. On the other hand, soil moisture content is projected to decrease throughout the year across the monsoon regions, suggesting potential agricultural drought risks all year round, causing reductions in agricultural yields, which are supposed to sustain nearly two-thirds of the world's population, and further threaten food security.

Regarding the temperature-related targets, a global warming of 1.5°C would already see an enhanced wet–dry season contrast in the water cycle (1.56% increase in $P - E$ in the wet season and –3.80% decrease in the dry season) and drier soil throughout the year (–0.79%) in the global monsoon region. Limiting further warming from above the 1.5°C threshold thus implies reduced

flood and drought risks from the further enhanced annual cycle of the water cycle. Hence, the enhanced seasonality, apart from the annual-mean intensification of water cycle, deserves dedicated attention in disaster management and adaptation strategies.

We note that this study investigates some specific aspects of the water cycle over the global monsoon regions, including its reservoirs (e.g., precipitable water, soil moisture) and fluxes (e.g., P , E , $P - E$, and total runoff). The integrated water cycle, however, refers to the movements of water within the whole Earth system. The future changes in other components of the water cycle in the monsoon regions, for example, the water movements within the oceans and on or within the landmass, deserve further study.

Acknowledgments. This work was supported by National Natural Science Foundation of China (Grant 41330423), the Strategic Priority Research Program of the Chinese Academy of Sciences (Grant XDA20060102), International Partnership Program of Chinese Academy of Sciences (Grant 134111KYSB20160031), China Postdoctoral Science Foundation (Grant 2018M641450), and Ministry of Science and Technology of China (Grant

FIG. 10. Changes in water cycle components including P , E , $P - E$, R , SMs , and SMt over the GLM region at 1.5°, 2°, 3°, and 4°C warming levels for the (a) annual mean, (b) wet season, and (c) dry season (% relative to the 1986–2005 baseline). The multimodel medians (histograms) and 25th–75th-percentile intervals (error bars) are shown. The wet season refers to May–September for the NH and November–March for the SH, and vice versa for the dry season. Only the 18 models in which a 4°C warming occurs before 2100 are used.

2018YFA0606501). We also acknowledge the support from the Jiangsu Collaborative Innovation Center for Climate Change. We acknowledge the World Climate Research Programme's Working Group on Coupled Modelling, which is responsible for CMIP, and we thank the climate modeling groups for producing and making available their model output. The authors declare that they have no conflict of interest.

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