Early Summer Response of the East Asian Summer Monsoon to Atmospheric CO$_2$ Forcing and Subsequent Sea Surface Warming

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ABSTRACT

The early summer regional climate change of the East Asian summer monsoon (EASM) is investigated in the phase 5 of the Coupled Model Intercomparison Project (CMIP5) archive. In the greenhouse gas–forced scenario, reduction of radiative cooling and increase in continental surface temperature occur much more rapidly than changes in sea surface temperatures (SSTs). Without changes in SSTs, the early summer rainfall in the monsoon region decreases (increases) over ocean (land) in most models. On longer time scales, as SSTs increase, rainfall changes are opposite. The total response to atmospheric CO$_2$ forcing and subsequent SST warming is a large (modest) increase in rainfall over ocean (land) in the EASM region. Dynamic changes, in spite of significant contributions from the thermodynamic component, play an important role in setting up the spatial pattern of precipitation changes. Early summer rainfall anomalies over east China are a direct consequence of local land–sea contrast, while changes in the large-scale oceanic rainfall band are closely associated with the displacement of the larger-scale North Pacific subtropical high (NPSH). Ad hoc numerical simulations with the AM2.1 general circulation model show that topography and SST patterns play an important role in early summer rainfall changes in the EASM region.

1. Introduction

It is well understood that the increase in global precipitation in response to greenhouse warming is energetically constrained rather than being limited by the availability of atmospheric water vapor (e.g., Mitchell et al. 1987; O’Gorman et al. 2012). Therefore, global precipitation changes less rapidly with temperature (at around 2% K$^{-1}$ in current climate) than the change in water vapor in the atmosphere at around 7.5% K$^{-1}$ from the Clausius–Clapeyron relation (Held and Soden 2006). Changes in precipitation at the regional scale are more complex, and arguably more important, than global changes, as circulation changes will affect the precipitation locally. Here, we explore regional changes in the East Asian summer monsoon (EASM) region during early summer, in response to CO$_2$ forcing. Although it has been found that the rainfall during the EASM season is projected to increase at the end of the twenty-first century (e.g., Zou and Zhou 2013; Kitoh et al. 2013), limited understanding has prevented us from robustly identifying the physical and dynamical processes contributing to the change of the EASM and from better constraining the intermodel spread of its projections. Understanding how the EASM responds to a changing climate can provide support to theories of its maintenance in present-day climate and shed light into the dynamics and responses to climate change of other subtropical convergence zones.

The mechanisms that alter regional precipitation vary at different time scales. A fast response to an increase in CO$_2$ concentration before sea surface temperatures (SSTs) change occurs at short time scales and is associated with changes in large-scale wind patterns in the atmosphere. Large uncertainties in the precipitation change are found in the fast response, particularly over tropical oceanic regions, which are identified as a primary contributor to the intermodel spread in the difference in simulated precipitation between two equilibrium climate states (Bony et al. 2013). A slow response to the subsequent increase in SSTs while maintaining the CO$_2$ concentration fixed in the atmosphere is found to resemble the climatological precipitation pattern following the “wet-get-wetter” behavior.


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(Held and Soden 2006). The influence on regional climate change by fast and slow responses was investigated in a recent study by Kamae et al. (2014), who show that a close relationship between changes in regional circulation patterns over East Asia and varying land–sea contrast exists in both observations and model simulations.

The wet-get-wetter response captures the thermodynamic response of net rainfall to SST forcing over oceanic regions. Based on the assumptions of unchanged relative humidity and circulation, increases in atmospheric water vapor in a warmer climate intensify climatological convergence of water vapor fluxes. As a result, climatological wet regions (positive net precipitation regions) will become wetter, and climatological dry regions (negative net precipitation regions) will become drier. This simplified depiction has been generally accepted in the study of the response of the hydrological cycle to climate change; however, because of its assumptions, it does not capture the complexity of the thermodynamic precipitation response at the regional scale. For example, Xie et al. (2010) found that tropical rainfall change follows a “warmer-get-wetter” pattern modulated by future SST pattern, rather than the wet-get-wetter pattern, which can only be realized if SSTs are increased uniformly.

While providing a useful starting point, the thermodynamic change due to SSTs is only one component of the total precipitation response. Dynamic changes in response to SST forcing have also been found to be important both globally and locally (e.g., Xie et al. 2010, 2009; He and Zhou 2015). Over the EASM region, dynamic changes have, for instance, been shown to be associated with changes in the North Pacific subtropical high (NPSH). Kitoh et al. (1997) found that global warming is associated with a strengthening and southward movement of the NPSH in a global climate model (GCM). Together with more El Niño–like patterns in future climates, it explains the mean sea level pressure anomalies that might be related to the delay of baiu withdrawal simulated in GCMs (Kitoh and Uchiyama 2006). The influence of tropical SST anomalies on the western NPSH has been vastly explored at the interannual time scale. It has been proposed that increases in rainfall over the tropical Indian Ocean due to the resulting warm SST anomalies from El Niño in the preceding year generate Kelvin waves emanating into the tropical western Pacific, inducing local northeasterly surface wind anomalies and resulting in an anticyclonic circulation over the western North Pacific (e.g., Yang et al. 2007; Xie et al. 2009; Wu et al. 2009). This signal can be enhanced by a cold tropical Pacific SST anomaly that generates anticyclonic Rossby waves to its northwestern region (e.g., Terao and Kubota 2005; Wu et al. 2010). The relationship between the western NPSH and the zonal SST gradient between the tropical Indian Ocean and the tropical western Pacific is examined in RCP4.5 and RCP8.5 model outputs in CMIP5 by He and Zhou (2015). They found that the zonal temperature gradient has a robust influence on simulated western NPSH anomalies, which modulate the climate change over eastern China. In addition, they performed a sensitivity test on the impact of tropical SST anomalies on the western NPSH, and they showed that both the tropical Indian Ocean and tropical western Pacific SST anomalies contribute to changes in the projected western NPSH intensity.

Mechanisms driven by changes other than just SSTs have, however, been invoked. Zhao et al. (2011a) investigated the tropical–North Pacific mode in present climate and found that this mode is closely correlated with the variability of climate over Asia and the Pacific Ocean through the Asian–Pacific Oscillation (APO). Sensitivity experiments emphasize the importance of the Asian land heating due to the Tibetan Plateau (TP) in generating summertime Asian–Pacific climate anomalies. Pacific SST forcing, seemingly important in this teleconnection, was suggested to play a much weaker role in the summertime Asian–Pacific atmospheric circulation. At interdecadal time scale, Zhao et al. (2011b) found that from a low-APO to a high-APO decade, both the upper-tropospheric South Asian high (SAH) and the lower-tropospheric low pressure system intensify over Asia. This strengthened circulation results in anomalous southerly and southwesterly winds prevailing over the Asian monsoon region and leads to a strong northward transport of moisture and enhanced rainfall over the Asian monsoon region.

In this paper, we investigate the response of the EASM to CO2 forcing at different time scales, and untangle various dynamic and thermodynamic processes that can mediate the precipitation response to changes in boundary forcing (such as land–sea contrast, topography, and SSTs) through radiation–circulation interactions. Specifically, we ask: What are the mechanisms of EASM rainfall changes and spread among CMIP5 models at different (fast and slow) time scales? To provide answers to this question, we leverage numerical experiments with state-of-the-art climate models in the CMIP5 archive. Additional experiments are performed with the Geophysical Fluid Dynamics Laboratory (GFDL) Atmospheric Model, version 2.1 (AM2.1; Anderson et al. 2004) to further investigate mechanisms. In section 2, we briefly describe the data and methods used in this study. In section 3, we present results from investigation of CMIP5 experiments. Additional experiments performed ad hoc are discussed in section 4. A discussion and conclusions are provided in section 5.

2. Data and method

We use 11 climate model single realization outputs (Table 1) with monthly means from several CMIP5
Table 1. CMIP5 models that have outputs in piControl, sstClim, sstClim4xCO2, and abrupt4xCO2. [Expansions of acronyms are available online at http://www.ametsoc.org/PubsAcronymList.]

<table>
<thead>
<tr>
<th>Model name</th>
<th>Modeling group</th>
<th>Resolution (plev × lat × lon)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BCC-CSM1.1</td>
<td>Beijing Climate Center (BCC), China Meteorological Administration</td>
<td>17 × 64 × 128</td>
</tr>
<tr>
<td>CanESM2</td>
<td>Canadian Centre for Climate Modeling and Analysis (CCCma)</td>
<td>22 × 64 × 128</td>
</tr>
<tr>
<td>CCSM4</td>
<td>National Center for Atmospheric Research (NCAR)</td>
<td>17 × 192 × 288</td>
</tr>
<tr>
<td>CSIRO-Mk3.6.0</td>
<td>CSIRO in collaboration with Queensland Climate Change Centre of Excellence (CSIRO-QCCCE)</td>
<td>18 × 96 × 192</td>
</tr>
<tr>
<td>INM-CM4</td>
<td>Institute of Numerical Mathematics (INM)</td>
<td>17 × 120 × 180</td>
</tr>
<tr>
<td>IPSL-CM5A-LR</td>
<td>L’Institut Pierre-Simon Laplace (IPSL)</td>
<td>17 × 96 × 96</td>
</tr>
<tr>
<td>MIROC5</td>
<td>L’Institut Pierre-Simon Laplace (IPSL)</td>
<td>17 × 128 × 256</td>
</tr>
<tr>
<td>MPI-ESM-LR</td>
<td>Max Planck Institute for Meteorology (MPI-M)</td>
<td>25 × 96 × 192</td>
</tr>
<tr>
<td>MPI-ESM-MR</td>
<td>Max Planck Institute for Meteorology (MPI-M)</td>
<td>25 × 96 × 192</td>
</tr>
<tr>
<td>MRI-CGCM3</td>
<td>Meteorological Research Institute (MRI)</td>
<td>23 × 160 × 320</td>
</tr>
<tr>
<td>NorESM1-M</td>
<td>Norwegian Climate Centre (NCC)</td>
<td>17 × 96 × 144</td>
</tr>
</tbody>
</table>

Experiments (Taylor et al. 2012): 30-yr atmosphere-only simulations forced by a fixed 1xCO2 or 4xCO2 concentration with prescribed SST distribution that remains unchanged in both sets of experiments (sstClim or sstClim4xCO2), 150-yr fully coupled ocean–atmosphere simulations forced by a constant 4xCO2 concentration (abrupt4xCO2), and fully coupled simulations forced by preindustrial forcing (piControl). The 30-yr sstClim and sstClim4xCO2, piControl, and the last 30 years of abrupt4xCO2 are averaged to represent the climatology of different climate states. The monthly resolution of available data does not allow for consideration of sub-monthly transient eddies in our analyses.

The fast response is computed as the difference between sstClim4xCO2 and sstClim, in which the SST distribution is prescribed based on the climatology from preindustrial simulations. In these two sets of experiments the only difference is, therefore, the atmospheric CO2 concentration. The slow response is computed as the difference between abrupt4xCO2 and sstClim4xCO2, in which the atmospheric CO2 concentration in both scenarios is essentially the same, and the only difference is the subsequent warming in SSTs in the abrupt4xCO2 scenario.

As done in several previous studies of regional climate changes (e.g., Seo et al. 2013; Qu et al. 2014), we use the moisture budget to study the hydrological change in the EASM region,

\[
\delta P + E = \nabla \cdot (vq), \tag{1}
\]

where \( v \) indicates horizontal winds, \( \omega \) is vertical pressure velocity, \( q \) is specific humidity in the atmosphere, \( P \) is precipitation, and \( E \) is evaporation; \( \langle \cdot \rangle \) indicates a mass-weighted vertical integral, while \( \langle \cdot \rangle \) indicates temporal mean. Ignoring water vapor storage in the atmosphere and vertical velocity at the surface, Eq. (1) can be written as

\[
\overline{P - E} = -\langle \nabla \cdot (vq) \rangle. \tag{2}
\]

This budget closes only if variables \( v \) and \( q \) include all temporal resolutions. Because our data are at monthly resolution, the calculated moisture flux convergence does not include the contribution from submonthly transient eddies. Hence, this contribution has to be estimated as the residual of Eq. (2). In the following, we will drop the notation \( \langle \cdot \rangle \), with all variables in following equations indicating monthly means.

To expose contributions from individual climatic variables to changes in the moisture budget, we decompose specific humidity \( q \) into the product of relative humidity \( H \) and saturation specific humidity \( q_s \), as done by previous studies. One caveat is that by using monthly averages of relative humidity, we ignore its covariance with temperature (through the saturation specific humidity) on frequencies higher than monthly.

The anomalous moisture budget can hence be written as

\[
\delta(P - E) = -\langle \delta \nabla \cdot (vHq_s) \rangle + \text{residual}, \tag{3}
\]

where \( \delta \) indicates the difference between sstClim4xCO2 and sstClim (abrupt4xCO2 and sstClim4xCO2) scenarios in the fast (slow) response, and the second term on the right-hand side of Eq. (3) is a residual, including submonthly transient eddies and moisture tendency in the atmosphere. The moisture flux convergence term can be further decomposed as

\[
\begin{align*}
-\langle \delta \nabla \cdot (vHq_s) \rangle &= -\langle \nabla \cdot [\delta vHq_s(T)] \rangle - \langle \nabla \cdot [v \delta Hq_s(T)] \rangle - \langle \nabla \cdot [v \delta Hq_s(T)] \rangle - \langle \nabla \cdot [v \delta Hq_s(T)] \rangle \\
&= -\langle \nabla \cdot [\delta vq_s(T) \delta H] \rangle - \langle \nabla \cdot [v \delta Hq_s(T)] \rangle - \langle \nabla \cdot [v \delta Hq_s(T)] \rangle - \langle \nabla \cdot [v \delta Hq_s(T)] \rangle. \tag{4}
\end{align*}
\]
where terms on the right-hand side represent, respectively, the change due to winds, relative humidity, saturation specific humidity and hence temperature, the covariance between relative humidity and temperature, the covariance between winds and relative humidity, the covariance between winds and temperature, and the covariance between relative humidity and temperature. Assuming no changes in winds and relative humidity anomalies due to saturation specific humidity, $-\langle \nabla \cdot \{vH\delta q_s(T)\} \rangle$, can be further decomposed into two terms, $-\langle \nabla \cdot \{vH\delta q(T)\} \rangle$ and $-\langle \nabla \cdot \{vH\delta[q_s(T) - q_s(T)]\} \rangle$, where $q_s(T)$ is $q_s$ at the surface. The former can also be written as $-\alpha \delta T(P - E)$, where $\alpha = L_v/RT^2$, $L_v$ is the latent heat of evaporation, and $R$ is the gas constant for water vapor. The term $-\alpha \delta T(P - E)$ has been described in the literature as the wet-get-wetter pattern (e.g., Held and Soden 2006), by assuming fixed shape of the temperature profile under climate change (similar to the Planck response in climate sensitivity studies) and ignoring changes in transient eddy fluxes. It predicts that with warming, $\delta T$, the pattern of net precipitation $(P - E)$ will simply be enhanced: becoming more positive when it is already positive, and more negative when it is already negative. In the following, we will refer to this component as the component induced by surface warming, to emphasize that this is the response in net precipitation due to changes in surface temperature through changes in the saturation specific humidity, assuming a fixed atmospheric temperature profile. The latter arises due to lapse rate changes or changes in the shape of the temperature profile.

In addition to these CMIP5 experiments, we also perform simulations with the GFDL AM2.1. Six experiments (noTopo_control, noTopo_4xCO2, Topo_control, Topo_4xCO2, Uni4K, and CMIP5SST) are performed (see more details in Table 2). These experiments have been designed to explore the impact of different regional forcings, such as land–sea contrast, topography, and SST distribution, on the EASM response. For instance, the difference between noTopo_4xCO2 and noTopo_control is expected to show how enhanced land–sea thermal contrast influences regional precipitation without any contribution from topographic forcing. These results can be compared with their counterparts with full topography. The difference between Topo_4xCO2 and Uni4K or CMIP5SST is expected to show how SST patterns (in addition to uniform SST warming) affect the EASM. Climatologically fixed SSTs without interannual variability from monthly-mean Reynolds SST analysis are used as boundary condition (Smith et al. 1996). Each experiment ran for 25 years, and the last 14 years of the simulations are used for the analyses.

We limit our analyses of changes in the EASM precipitation and circulation to the month of June, when most models well capture the EASM rainfall band. While we explore circulation changes over the entire North Pacific and their possible links to precipitation changes, in this study, we primarily focus on the mei-yu–baiu (MB) front, the most prominent precipitation feature of the EASM in early summer (e.g., Chen and Bordoni 2014). Notwithstanding the intermodel spread in its simulated climatology, the rainfall band is well defined only in June across all models. In this regard, we take June as the month most representative of the mature MB season. Consequently, we ignore possible changes in the EASM seasonality and only focus on seasonal mean changes in rainfall intensity and position as defined by the June long-term mean.

3. Rainfall anomalies

With quadruple CO2 forcing, rainfall increases over the EASM region, particularly over the oceanic regions on the southern flank of the rainfall band (Fig. 1a). Most of the precipitation increase only happens when SST starts to warm. With CO2 forcing alone, rainfall decreases over oceanic regions, while it increases over east China (Fig. 1b). The decrease in precipitation is collocated with the rainfall band, indicating that this response is not simply a result of model artifact but a robust signal in changes in the EASM precipitation. The slow response shows a pattern opposite to the fast response: rainfall decreases over east China while it significantly increases over the oceanic regions (Fig. 1d). The difference in EASM precipitation between coupled and uncoupled simulations is fairly small [Fig. 1c; the spatial pattern and magnitude is consistent with a recent study by Song and Zhou (2014, their Fig. 2c)] compared to that in either fast or slow response, allowing us to safely conclude that air–sea interaction can be ignored and that the signal in Fig. 1d comes from the SST forcing in the MMM.

a. Fast response

The fast response of the EASM rainfall band to elevated CO2 concentration with fixed SSTs features a decrease (increase) of precipitation over oceanic (land) regions (Fig. 1b). This precipitation response is robust in most models (not shown).

Anomalies in net precipitation (Fig. 2a) largely explain the pattern of precipitation change in the EASM (Fig. 1b), with changes in evaporation being important only over oceanic regions: here, the contribution by evaporation decreases along regions of large climatological evaporation (Fig. 2b). The spatial pattern of net precipitation change is consistent with changes in mean
moisture flux convergence (Fig. 2c), although transient eddy flux anomalies, calculated as the residual of the moisture budget, are not negligible (Fig. 2d). Changes in mean moisture flux convergence are mainly captured by those due to winds (Fig. 2e). Contributions from changes in temperature (Fig. 2g), relative humidity (Fig. 2f), and their covariances (Figs. 2j–l) play a less important role. This confirms that in the absence of SST changes, the precipitation response is primarily dominated by changes in circulation, as seen in other tropical and subtropical regions (Bony et al. 2013).

b. Slow response

At a first glance, changes in the slow response appear to follow the wet-get-wetter pattern. However, important deviations from the simple thermodynamic change exist (Fig. 3a): While the response is characterized by a well-organized positive change in net precipitation, this is located to the south of, rather than being collocated with, its climatological location. The net precipitation change over east China is negative, counteracting its positive change in the fast response. Surface evaporation increases, particularly over oceanic regions where large evaporation reductions occur in the fast response (Fig. 3b). This increase in surface evaporation might be due to the experiment configuration: in sstClim4xCO2, SSTs are prescribed and surface evaporation is strongly limited; in abrupt4xCO2, SSTs are interactive, and local SST changes might contribute to the narrow band of enhanced evaporation.
The mean flux convergence, $\langle \nabla \cdot (\mathbf{v} q_s) \rangle$, captures the overall spatial pattern of the net precipitation change in Fig. 3a, with strong moisture convergence on the southern flank of the rainfall band (Fig. 3c). Transient eddies contribute significantly to the balance (Fig. 3d). Because of the monthly resolution of the CMIP5 data, the transient eddy contribution is estimated from the moisture budget residual, which prevents a more careful mechanistic understanding of this response. Changes due to winds (Fig. 3e) and temperature (Fig. 3g) are both important, with circulation changes dominating the overall spatial pattern, and temperature changes increasing moisture convergence over the climatological convergence zone. Contributions from relative humidity changes are nontrivial, but their magnitude and spatial extent are smaller than those from wind and temperature changes (Fig. 3f). As discussed in section 2, changes due to temperature can be decomposed into components induced by surface warming (Fig. 3h) and lapse rate changes (Fig. 3i). The component induced by surface warming relates the climatological net precipitation, weighted by the surface warming, to changes in net precipitation, or the so-called wet-get-wetter pattern. This component dominates the total response due to temperature, in both magnitude and spatial pattern. Weak signals over some land and oceanic regions are due to nearly zero climatological net precipitation, where local precipitation is primarily balanced by evaporation (cf. Fig. 7a in Chen and Bordoni 2014). The coupling between temperature (saturation specific humidity) and wind changes (Fig. 3l) is dominant among the covariance terms (Figs. 3j–l) and resembles the dynamic change due to only winds (Fig. 3e). The reasoning is as follows: since temperature increases everywhere, the sign in the response is due to changes in winds, with specific humidity $q_s(T)$ and specific humidity changes $\delta q_s(T)$ acting as scaling factors.\(^1\)

\(^1\) A comparison between Figs. 3e and 3l shows that $q_s(T)$ and $\delta q_s(T)$ are of similar magnitude. This is due to the nonlinear dependence of $q_s(T)$ on temperature, which gives rises to big changes in $q_s(T)$ even for small changes in $T$. For instance, the water vapor saturation pressure is 3523 Pa at 300 K and 4701 Pa at 305 K, which implies that a 5-K temperature difference results in a 33% difference in water vapor saturation pressure.
In both fast and slow responses, changes in circulation are significant and dominate the spatial pattern of the precipitation anomalies. Changes in thermodynamic quantities, such as temperature and relative humidity, play a less important role. Hence, we focus primarily on analyzing the local circulation changes and infer possible mechanisms through which fundamental forcings, such as land–sea contrast, topography, and atmospheric CO₂, affect local circulations directly or indirectly through larger-scale atmospheric circulation changes such as those of the NPSH.

4. Dynamic contribution to rainfall changes

Figure 4 shows changes in geopotential height, precipitation, and moisture flux due to changes in winds (i.e., the dynamic component). Specifically, to clearly link geopotential height to circulation changes, in Fig. 4 we show differences in the local geopotential relative to the maximum value in the NPSH. This is because, through geostrophic balance, winds are linked to gradients in geopotential height rather than its magnitude. Additionally, geopotential heights tend to systematically shift upward under global warming.

On the larger scale, changes in the location and the strength of the NPSH in the fast response are within one standard deviation of the intermodel spread and therefore not significant (Fig. 4, top). In the slow response, instead, the NPSH moves southward and weakens significantly (Fig. 4, bottom). This suggests that changes in winds over the EASM region are mostly local responses in the fast response, while resulting from a combination of local and remote responses, mediated by the NPSH, in the slow response.

The dynamic moisture flux convergence anomalies \[-(\nabla \cdot (q_0 \delta v))\], can be further decomposed into a wind convergence component \[-(q_0 \nabla \cdot \delta v)\] and an advection component \[-(\delta v \cdot \nabla q_0)\]. In the following, we will analyze separately the contribution to precipitation changes by these two terms and we will discuss possible mechanisms responsible for these changes.

a. Changes in wind convergence

The wind convergence component, \[-(q_0 \nabla \cdot \delta v)\], can be expressed in terms of the vertical advection using continuity, \[-(\delta \omega \partial_z q_0)\]. The change in this term is largely explained by changes in vertical velocity at 500 mb (i.e., \(\delta \omega_{500}\), Fig. 5).

The vertical velocity is directly associated with remote forcing (i.e., energy advection), local radiative and surface fluxes, and stability. According to the MSE budget (e.g., Chen and Bordoni 2014), vertical velocity can be approximated as the fraction between energy input and moist static stability. Here, we define a proxy for vertical velocity at 500 mb based on the MSE budget,
\[
\alpha_{\text{sst}0\text{apprx}} = \frac{-\langle v \cdot \nabla E \rangle + F_{\text{net}}}{-\alpha \langle \partial_p h \rangle},
\]
where \(F_{\text{net}} = S_t^i - S_s^i + S_s^i - R_t^i + R_s^i - R_s^i + SH + LH\) is the net energy flux into the atmosphere, with the subscript \(t\) and \(s\) denoting the top of atmosphere and surface, and \(S\) and \(R\) indicating the shortwave and longwave radiative fluxes, respectively.\(^2\) Also, \(h = c_p T + g z + L_v q\) is the MSE and \(E = c_p T + L_v q\) is the atmospheric moist enthalpy; \(\alpha\) is a coefficient added to account for the coupling between vertical velocity and MSE stratification. Transient eddies are ignored and the coupling coefficient \(\alpha\) is assumed to be homogeneous for simplicity. Figure 6 shows changes in vertical velocity as diagnosed from the model directly and from the approximation in Eq. (A1) (i.e., \(\delta w\) and assuming \(\alpha = 1\)).

At the first order, changes in vertical velocity can be partitioned into changes in energy input and changes in stability (see the appendix). Contributions from changes in energy input (mostly from horizontal advection of moist enthalpy) are significantly larger than those from changes in stability in both fast and slow responses (Fig. 7). In the fast response, anomalous positive moist enthalpy advection over northeast China and negative moist enthalpy advection over the climatological rainfall band are closely associated with changes in vertical velocity. In the slow response, anomalies in moist enthalpy advection change sign, with anomalous positive

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\(^2\) The vertical integration of moist static energy stratification is from 700 to 100 mb to account for the steepest slope for stability.
moist enthalpy advection over ocean and negative advection over land. Contributions from local stability are considerably smaller, however, with a destabilizing effect over land in the fast response, and over oceanic regions in the slow response. Anomalies in moist enthalpy advection are due to both dry enthalpy and latent energy advection, with similar spatial pattern (not shown) because of the close relationship between temperature and water vapor changes via the Clausius–Clapeyron relationship.

**b. Changes in advection**

Changes in the advection term \(-\langle \delta \mathbf{v} \cdot \nabla q_0 \rangle\) are a direct result from (mostly geostrophic) wind anomalies. In the fast and slow responses, changes in local precipitation over east China and adjacent oceanic regions are strongly associated with meridional wind anomalies (Figs. 8b,d). Intensified (weakened) meridional wind enhances (reduces) moisture transport, resulting in higher (lower) rainfall. In addition, the meridional component of the geostrophic flow on a \(\beta\) plane can induce convergent flow, which reinforces local precipitation in addition to positive advective anomalies. Changes in meridional wind at 850 mb are largely geostrophic, a consequence from changes in surface pressure gradient through geopotential height (\(Z_{850}\)) gradient anomalies. For simplicity, ignoring subtle influences from changes in the atmospheric temperature between the surface and 850-mb pressure level, \(Z_{850}\) is only dependent on \(\ln(p_s)\), where \(p_s\) indicates surface pressure. Anomalies in locational differences in surface pressure—that is, \(\delta \ln(p_{s1}/p_{s2})\), where \(p_{s1}\) and \(p_{s2}\) indicate surface pressures in two separate regions (please refer to Fig. 9 for details)—change the \(Z_{850}\) gradient, and thereafter create wind anomalies, \(\delta v_{850}\). In the fast response, enhanced land–sea contrast is manifest in an increased surface pressure gradient, with lower pressure over land and higher pressure over ocean. The meridional wind is subsequently enhanced. In the slow response, however, land–sea contrast is weakened, and the meridional wind is reduced. This relationship is well observed among different model simulations (Fig. 9).3

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3 The robustness of the relationship between \(\delta \ln(p_{s1}/p_{s2})\) and \(\delta v_{850}\) is insensitive to the width of the region we choose (with the eastern boundary varying from 130° to 140°E).
Changes in precipitation over the oceanic rainfall band, however, are largely due to changes in zonal wind, particularly in the slow response (Fig. 8c). Enhanced lower-level westerly wind might be related to a southward displacement of the NPSH. In the fast response, the NPSH does not feature significant changes in its spatial pattern, which might explain why contributions from anomalous advection of climatological moisture are limited.
c. Summary

We have diagnosed precipitation changes in the EASM region in both fast and slow responses. Some robust conclusions emerging from this diagnosis include the following:

- Spatial changes in net precipitation are associated with changes in the moisture flux convergence, which is dominated by the dynamic component (i.e., by changes in circulation).
- The wind convergence term in the dynamic component is directly linked to changes in vertical velocity through mass conservation.
- These changes in vertical velocity are found to be mostly related to changes in moist enthalpy advection, with changes in vertical stability playing a secondary role.
- Changes in horizontal moisture advection over east China are dominated by changes in the meridional wind, which is a consequence of changes in land–sea contrast. The zonal component dominates the slow response over the oceanic regions, as a possible consequence of the southward displacement of the NPSH.

5. Numerical simulations

We use the GFDL AM2.1 to investigate the impact of land warming and SST patterns on the EASM response to CO₂ forcing. We have previously shown that topography plays an essential role in the rainfall band formation (Chen and Bordoni 2014). As discussed there, the presence of topography reinforces the land–sea thermal contrast, in addition to mechanically interacting with the prevailing flow. However, in a changing climate, can enhanced land–sea thermal contrast due to land warming alone provide a large enough forcing to cause changes in precipitation? In addition, how do changes in SST patterns affect the response of the EASM?

a. Land warming

To expose impacts of land warming alone on the EASM, we design two experiments with changing CO₂ concentration in the absence of global topography: noTopo_control and noTopo_4xCO₂. Figure 10 shows...
the simulated precipitation change in May, June, and July\textsuperscript{4} with and without topography. Consistent with Chen and Bordoni (2014), in the presence of topography, a well-organized rainfall band is simulated during the EASM season, while the rainfall band disappears when topography is removed. In the absence of topography (Fig. 10, right), there are no significant changes in precipitation until July, when the rainfall band dissipates even in the control experiment (with topography). The difference highlights the limited impact of land warming alone on the EASM rainfall, and emphasizes the importance of topography not only in its climatology but also in its response to CO\textsubscript{2} forcing.

As discussed by several previous studies (Sampe and Xie 2010; Molnar et al. 2010; Chen and Bordoni 2014), the TP can affect the EASM formation by enhancing the existing land–sea thermal contrast, which can reinforce the southeasterly flow on the western flank of the NPSH, and/or through mechanical interaction with the upper-level zonal flow, which also has been shown to reinforce the western flank of the NPSH (Molnar et al. 2010; Chen and Bordoni 2014). Our experiments show that topography is essential to project changes in the EASM rainfall under CO\textsubscript{2} forcing. In the fast response, the presence of topography increases the land–sea thermal contrast. However, this enhancement alone cannot generate a response similar to the one obtained when topography is retained. If we assume linearity and additivity, these results imply that changes in land–sea

\textsuperscript{4}In the AM2.1 simulations, the MB season is anticipated by one month. This motivates our inclusion of the month of May in the analyses presented in this section.
thermal contrast alone are not sufficient for rainfall changes that would otherwise be simulated if topography is retained and that other dynamical processes, such as the interaction between the larger-scale zonal flow and topography, might be playing an important role. There also exists the possibility that linearity and additivity do not hold, suggesting nonlinear interactions in the precipitation response. More detailed investigation of the exact mechanisms through which topography affects the EASM response to warming will be the focus of future studies.

b. SST patterns

Previous literature (e.g., Xie et al. 2010) has discussed the importance of SST patterns in regional precipitation changes, arguing that the wet-get-wetter response can hold only for uniform SST changes. We illustrate the impact of spatially varying SST patterns on the projected EASM rainfall by comparing the Uni4K and CMIP5SSST experiments (Fig. 11). With spatially varying SSTs (Fig. 12a), rainfall increases from east China through the northwestern Pacific; rainfall instead decreases over Japan and part of the northwestern Pacific in the case of uniform SST warming (Fig. 12b). In the latter case, changes in regional net precipitation do not follow the pattern of the climatological net precipitation, as would be expected from the wet-get-wetter response.

Differences in rainfall projection are largely due to the dynamic component (Fig. 13) in the moisture budget. With spatially varying SSTs, the NPSH weakens and moves southward, which is associated with a southward displacement of the westerly jet (Fig. 12a). The weakening of the NPSH, together with its spatial displacement, creates an anomalous westerly wind to the southeast of Japan, resembling the MMM response in CMIP5 simulations. With uniform SST warming, the NPSH intensifies and there is little evidence of any southward displacement. As a consequence, the prevailing wind to the southeast of Japan is northeasterly, which results in a reduction in precipitation.
6. Summary and discussion

In this study, we analyzed the response of the EASM early summer rainfall band to atmospheric CO2 forcing and subsequent SST warming within the context of the moisture budget. The spatial pattern of net precipitation changes is dominated by changes in mean moisture flux convergence, which in turn is primarily explained by changes in circulation. The thermodynamic component, however, is nonnegligible; it mimics the net precipitation climatology and contributes significantly to rainfall changes under warming. Surface pressure anomalies, as a consequence of land–sea contrast due to CO2 forcing, create an anomalous meridional flow over east China and adjacent oceans, which affects the moisture advection. The NPSH weakens and moves significantly southward in the slow response, creating an anomalous westerly flow to the south of the climatological rainfall band and subsequently increasing moisture advection. In addition to contributions from anomalous moisture advection due to winds, anomalous wind convergence also contributes to rainfall changes. The spatial pattern is collocated with that of vertical velocity anomalies at 500 mb, which can be thought of as a response to a remote forcing, provided by anomalous horizontal moist enthalpy advection. The schematics in Fig. 14 summarizes the underlying mechanisms: in the fast response to CO2 forcing, enhanced land–sea thermal contrast reinforces the meridional wind, which results in an increase of rainfall over east China; in the slow response, the land–sea thermal contrast is weakened because of sea surface warming, and the rainfall over east China decreases, resulting from a southward displacement of the NPSH.

Numerical simulations without topography show that enhanced land–sea contrast due to land warming alone cannot induce similar precipitation changes. This result implies that the land warming is not a sufficient condition for the EASM rainfall changes. It also emphasizes the important role of topography in the EASM response in terms of its climatology and climate change. In addition, spatially varying SST changes are shown to play a key role in rainfall changes in the

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**Fig. 13.** GFDL AM2.1 simulations of (a)–(c) net precipitation, (d)–(f) evaporation, (g)–(i) wind component, (j)–(l) relative humidity component, and (m)–(o) temperature component as in Fig. 2 for (left) fast response (Topo_4xCO2 – Topo_Control), (middle) slow response (CMIP5SST – Topo_4xCO2), and (right) slow response with uniform 4-K increase in SSTs (Uni4K – Topo_4xCO2) with full topography in June. Line contours (between 3 and 9 mm day$^{-1}$ with interval of 1 mm day$^{-1}$) indicate climatological precipitation in each comparison.
oceanic regions through associated changes in the NPSH. One caveat worth emphasizing is that, because of the spread in model performances, our analyses are limited to the month of June, when all models well capture the EASM rainfall band. Hence, the results and mechanisms discussed in this work might not be directly applicable to the late summer response (i.e., August) of the EASM.

The fast and slow responses of the EASM to CO₂ forcing show an opposite pattern, implying a compensating effect in transient climate change. This result is consistent with recent work by Shaw and Voigt (2015), who highlight how changes in land–sea contrast in response to the direct radiative forcing and the indirect SST warming have an opposite impact on global circulation changes. This might explain why, over the North Pacific, the NPSH shows little change under transient climate change such as the representative concentration pathway scenarios (He and Zhou 2015) because of the combined and contrasting influences of the fast and slow responses. Speaking of the EASM specifically, we acknowledge the importance of land–sea contrast but emphasize the role of topography rather than that of land warming alone. Our simulations are based on GCM experiments with realistic continents, rather than the more idealized study by Shaw and Voigt (2015), who prescribe SSTs to artificially introduce land–sea contrast in their aquaplanet simulations.

Results emerging from this work have important implications for improving EASM projections in GCMs. The dynamic component due to circulation changes, although highly model dependent and hard to constrain, can disclose mechanisms through which different forcing agents influence the EASM. The thermodynamic component mimics the climatology. Therefore, a better representation of the climatological precipitation will be the first necessary step to reduce spread in regional precipitation projections. In addition, analysis of the results from the fast response highlights how dramatic changes in rainfall can occur even as a direct response to CO₂ forcing, without any SST warming. These changes can have a tremendous societal impact on heavily populated monsoon regions. This confirms how geo-engineering schemes that have been proposed as climate mitigation strategies and that only aim at reducing surface warming without CO₂ sequestration might have unexpected implications for the global and regional hydrological cycle (Bony et al. 2013; O’Gorman et al. 2012).

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APPENDIX

Decomposition of the Approximated Vertical Velocity

We use \( a, s, \) and \( w \) to represent energy input \( - (\mathbf{v} \cdot \nabla E) + \mathbf{F}^{\text{net}} \), stability \( (\delta_p h) \), and approximated vertical velocity at 500 mb \( (\omega_{500,\text{apprx}}) \). Therefore, Eq. (5) can be expressed symbolically as \( \omega = (a/\omega) \). Fractional changes in vertical velocity \( (\delta \omega) \) can be expressed as

\[
\frac{\delta \omega}{\omega_0} = \frac{\delta a}{a_0} - \frac{\delta s}{s_0}. \tag{A1}
\]
where the subscript 0 indicates the control experiment, which is sstClim (sstClim4xCO2) in the fast (slow) response. Because $a_0$ is close to zero in some regions, we reformulate Eq. (A1) by multiplying $a_0$ on both sides of the equation:

$$\frac{a_0 \delta w}{w_0} = \delta a - \frac{a_0 \delta s}{s_0}.$$  \hspace{1cm} (A2)

One advantage of this approach is that we can avoid imposing an empirical value of $\alpha$ while still being able to diagnose respective contributions.

REFERENCES


