Reduced African Easterly Wave Activity with Quadrupled CO₂ in the Superparameterized CESM

WALTER M. HANNAH AND ANANTHA AIYYER

North Carolina State University, Raleigh, North Carolina

(Manuscript received 17 November 2016, in final form 2 May 2017)

ABSTRACT

African easterly wave (AEW) activity is examined in quadrupled CO₂ experiments with the superparameterized CESM (SP-CESM). The variance of 2–10-day filtered precipitation increases with warming over the West African monsoon region, suggesting increased AEW activity. The perturbation enstrophy budget is used to investigate the dynamic signature of AEW activity. The northern wave track becomes more active associated with enhanced baroclinicity, consistent with previous studies. The southern track exhibits a surprising reduction of wave activity associated with less frequent occurrence of weak waves and a slight increase in the occurrence of strong waves. These changes are connected to changes in the profile of vortex stretching and tilting that can be understood as interconnected consequences of increased static stability from the lapse rate response, weak temperature gradient balance, and the fixed anvil temperature hypothesis.

1. Introduction

Several aspects of the global climate will respond predictably to warming caused by elevated CO₂ (Boer 1993; Romps 2011), but regional responses involve complex interactions across a range of spatial and temporal scales and are therefore much harder to constrain. Regional responses to warming are subject to local factors, such as topography, as well as remote factors, such as basin-scale sea surface temperature (SST) patterns. The African Sahel is a particularly interesting region in this respect because it is home to drought-sensitive populations, and is also where a large number of Atlantic tropical cyclone (TC) precursor disturbances, known as African easterly waves (AEWs), originate during the summer monsoon season (Thorncroft and Hodges 2001; Hopsch et al. 2007). Thus, the future climate of the Sahel may have far-reaching impacts on coastal communities in the tropical Atlantic.

AEWs are westward-propagating disturbances with a 2–6-day period that occur over the Sahel region during boreal summer (Reed et al. 1977; Kiladis et al. 2006). The flow associated with AEWs helps to organize

convective systems (Mekonnen et al. 2006) and also plays a role in transporting Saharan dust (Jones et al. 2003). AEWs are generally thought to be convectively triggered (Thorncroft et al. 2008), but recent studies have revealed a role for upstream development in AEW initiation that is similar to the downstream development mechanism in midlatitude Rossby waves (Diaz and Aiyyer 2013b). AEWs amplify through hydrodynamic instability of the midtropospheric African easterly jet (AEJ; Burpee 1972; Norquist et al. 1977; Hsieh and Cook 2008; Hall et al. 2006), as well as upstream development (Diaz and Aiyyer 2013a). The instability of the AEJ is associated with the reversal of the meridional potential vorticity (PV) gradient over Africa, which is a necessary condition for combined barotropic and baroclinic instability (Charney and Stern 1962).

AEW activity is categorized into two storm tracks that are located to the north (~20°N) and south (~10°N) of the AEJ, which is centered at 15°N (Chen 2006; Thorncroft and Hodges 2001). Most TC genesis occurs downstream of the southern track (Hopsch et al. 2007), with 61% of TCs forming directly from an AEW (Russell et al. 2016). Northern track waves can also contribute to TC development (Chen et al. 2008), although the proportion of TCs attributed to northern track waves is difficult to estimate. Waves in the northern track are relatively dry and shallow, growing mostly through baroclinic energy conversions around 850 mb.

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-16-0822.s1>.

Corresponding author: Walter Hannah, walter@hannahlab.org

Waves in the southern track are deeper and grow primarily through barotropic energy conversion at the level of the jet. Although both baroclinic and barotropic energy conversions play a role in both storm tracks, convective coupling and upstream energy dispersion play a central role in the generation and intermittency of AEWs, especially in the southern track (Hall et al. 2006; Diaz and Aiyer 2013a,b).

Studies of North African climate using both observations and models have mostly focused on understanding continental-scale precipitation patterns and variability (Giannini et al. 2008; Biasutti and Giannini 2006; Biasutti et al. 2009; Biasutti 2013), which are largely influenced by low-frequency SST variability and less by the warming trend associated with historical CO₂ emissions. These precipitation patterns are generally captured in coupled model simulations of the twentieth century (Cook and Vizi 2006; Roehrig et al. 2013), but the variability is often misrepresented due to SST biases that cause precipitation to be too weak over the Sahel and too strong along the Guinea coast (Vizi and Cook 2002; Ruti and Dell'Aquila 2010). This problem complicates the task of projecting how the climate of North Africa might change in the future, as well as the confidence in how well AEWs can be represented in these models. Nonetheless, climate projections consistently indicate a shorter rainy season over West Africa with warming (Biasutti and Sobel 2009; Mohino et al. 2011; Biasutti 2013), which might have implications for future Atlantic TC activity.

Two recent studies by Martin and Thorncroft (2015) and Skinner and Diffenbaugh (2014) have explored how AEW activity might change with elevated levels of CO₂ using data from phase 5 of the Coupled Model Intercomparison Project (CMIP5). Although biases are evident in the representation of AEWs, such as weak convective coupling and propagation, the models reveal a robust increase of northern track activity with warming due to an enhanced meridional surface temperature gradient and stronger baroclinic energy conversions. The model response is heavily influenced by horizontal resolution because a coarse grid cannot adequately resolve topography, which plays a key role shaping Africa's climatological precipitation and circulation patterns. Martin and Thorncroft (2015) also found a seasonal shift in wave activity such that it maximizes later in the season.

Weak convective coupling is a common issue in models with parameterized convection, as evidenced by weak spectral signatures of convectively coupled phenomena when compared to observations (Straub et al. 2010; Skinner and Diffenbaugh 2013). This is likely due to the tendency of parameterized convection

to trigger too easily, leading to a high frequency of light rain and a low frequency of heavy rain (Holloway et al. 2012). A novel alternative to conventional parameterizations, known as superparameterization, overcomes this issue and has demonstrated improved representation of convection and convectively coupled phenomena (Khairoutdinov et al. 2008; Pritchard and Somerville 2009; DeMott et al. 2011; Andersen and Kuang 2012). Superparameterization replaces the conventional parameterizations for convection, radiation, microphysics, and turbulence with an embedded two dimensional cloud-resolving model in each grid column of the host model to explicitly simulate unresolved processes (Grabowski 2001; Khairoutdinov and Randall 2001). This explicit representation of convection allows a natural evolution of convective elements and a more realistic interaction between convective-scale processes, so it is an ideal tool for deepening our understanding of convectively coupled phenomena.

McCrary et al. (2014a) examined the West African monsoon system in simulations with a superparameterized version of the NCAR Community Climate System Model (SP-CCSM; Stan et al. 2010) and found that some of the biases were improved relative to the conventional CCSM. The SP-CCSM was also found to have more AEW activity, although the waves were much more intense than observed (McCrary et al. 2014b). The stronger waves in SP-CCSM were associated with excessive diabatic generation of eddy available potential energy associated with convection (i.e., overly strong coupling), whereas CCSM exhibited a lack of coupling between convection and AEW circulations.

The goal of the present study is to understand the equilibrium response of AEW activity to a quadrupling of CO₂ in a superparameterized climate model. To do this requires understanding how tropical atmosphere will respond to elevated CO₂ and applying this to the dynamics of AEWs. The tropical atmosphere is characterized by weak horizontal gradients of temperature and pressure due to the large Rossby radius (Charney 1963). On time scales of a day or more this leads to a balance between diabatic heating and adiabatic cooling by vertical motions:

$$\omega \frac{\partial s}{\partial p} \approx Q_1, \quad (1)$$

where ω is the vertical pressure velocity, $s = C_p T + gz$ is the dry static energy, and Q_1 is the diabatic heating. This balance is known as the weak temperature gradient balance (WTG; Sobel and Bretherton 2000; Sobel et al. 2001) or the weak pressure gradient balance (Roms 2012). For time scales for which this balance is relevant

there is a tight relationship between diabatic heating and vertical motion that is dictated by the static stability.

The vertical temperature structure of the tropical atmosphere generally follows a moist adiabat, which implies that a lower-tropospheric warming leads to a larger warming aloft, due to the Clausius–Clapeyron relation. Climate models robustly predict a large tropical upper-tropospheric warming in response to elevated CO_2 , which provides a slight negative feedback to the overall warming (Soden and Held 2006; He and Soden 2015). Satellite and radiosonde data have delivered mixed conclusions about the relevance of this mechanism in the real atmosphere, but these observations are complicated by nonclimatic artifacts (Fu et al. 2004; Sherwood et al. 2008). An indirect approach of exploiting thermal wind balance to diagnose upper tropospheric temperature trends can overcome these issues, and has shown results more in line with climate projections (Allen and Sherwood 2008).

The warming aloft associated with the lapse rate response naturally increases the stability of tropical troposphere, which will affect the WTG balance in (1) such that the vertical velocity needed to balance a diabatic heating anomaly will be weaker. The time scale of AEWs is sufficiently long that the WTG balance is relevant. Thus, this change in the relationship between diabatic heating and vertical motion may have consequences for AEW dynamics in the southern wave track to the extent that WTG balance is a reasonable approximation (see section 5).

Another projected consequence of elevated CO_2 is known as the fixed anvil temperature (FAT) hypothesis, in which a reduction of longwave cooling and tropospheric warming leads to a higher terminal altitude of deep convection (Hartmann and Larson 2002; Kuang and Hartmann 2007; Ingram 2010; Zelinka and Hartmann 2010; Li et al. 2012). The FAT hypothesis and its variants nicely explain the increase in cloud height that is robust across climate model projections. Given how prevalent deep convection is over the West African monsoon region, the FAT hypothesis is a useful concept for understanding how the African monsoon will respond to elevated CO_2 . This mechanism may also be relevant to AEW activity since convection is known to play a role in their dynamics (Hsieh and Cook 2007).

We hypothesize that vertical velocity anomalies will be reduced by an altered WTG balance in a warmer climate, which will act to reduce the southern track wave activity. The data and methods are outlined in section 2, followed by a description of the model climate response and precipitation variability in section 3. Section 4 focuses on AEW activity and uses the eddy enstrophy budget to understand the response. Section 5 discusses the validity of the

WTG balance in the context of the location and time scales that AEWs exist. Conclusions are presented in section 6.

2. Methods

a. Model experiments

Experiments are conducted with the National Center for Atmospheric Research (NCAR) Community Earth System Model (CESM) version 1.1.1, similar to the experiments described in Arnold et al. (2015). The atmosphere component of the model is run with a finite-volume dynamical core and 26 vertical levels. The horizontal resolution is $0.9^\circ \times 1.25^\circ$ to adequately resolve African topography, which is higher than the T42 resolution ($\sim 2.8^\circ$) used by McCrary et al. (2014a,b). The simulations are fully coupled to the ocean, land, and sea ice models. The ocean model is the Parallel Ocean Program version 2 (POP2) run at approximately 1° resolution.

The superparameterized CESM (SP-CESM) uses a 2D cloud-resolving model (CRM) oriented in the north–south direction to replace the conventional parameterizations for convection, microphysics, radiation, and turbulence. Microphysics, radiation, and turbulence are still parameterized on the CRM grid instead of the GCM grid, so the CRM is responsible for estimating these tendencies and sending them to the host GCM. Note that since the CRM is 2D, convective momentum transport (CMT) cannot be fully represented and communicated back to the GCM grid, although Tulich (2015) has recently shown a promising method for representing CMT in a 2D superparameterized model.

Because of the high computational cost of running SP-CESM it cannot run long enough to establish an equilibrium. Instead, we follow the approach of Arnold et al. (2015) and use the conventional (non-SP) CESM to run two simulations to a steady state with CO_2 concentrations of 285 ppm for 60 years and 1140 ppm for 340 years. New SP-CESM and CESM simulations were branched from these spinup runs and integrated for 11 years with 6-hourly output, and 3-hourly output for select variables such as precipitation and surface fluxes. The SP-CESM simulations exhibited very minimal climate adjustment or drift, similar to Arnold et al. (2015). Thus, the last 10 years are used for analysis and results are not sensitive to omitting more data. The SP simulations will be referred to as SP1x and SP4x. Similarly, the standard CESM simulations that use the Zhang and McFarlane (1995) deep cumulus scheme are referred to as ZM1x and ZM4x.

b. Observation data

The ECMWF interim reanalysis (ERA-Interim, hereinafter ERAi; Dee et al. 2011) is used here for model comparison and was obtained from the ECMWF

data server. ERAi data are on a $1^\circ \times 1^\circ$ grid with 19 vertical levels from 1000 to 70 hPa. For precipitation, we make use of the TRMM 3B42 3-hourly high-resolution ($0.25^\circ \times 0.25^\circ$) merged satellite rainfall product (Huffman et al. 2007). TRMM data are interpolated to the coarser model resolution for comparison. TRMM and ERAi data are considered over the 10-yr period 2000–09 to make comparisons with the 10 years of simulation data.

c. Perturbation enstrophy budget

Our primary tool for investigating AEW dynamics will be the budget of perturbation enstrophy (Lau and Lau 1992). Kinetic energy is defined as the square of velocity divided by 2, and similarly enstrophy (ξ) is a measure of rotational kinetic energy defined by the square of the absolute vorticity (ζ) divided by 2:

$$\xi = \frac{\zeta^2}{2}. \quad (2)$$

The perturbation enstrophy is similarly defined as the square of the perturbation vorticity, and is essentially the variance of vorticity when averaged in time. Perturbations are defined here using a 2–10-day bandpass filter and indicated by a prime. Low-frequency, or “background,” quantities are defined with a 10-day low-pass filter and indicated by an overbar. We focus on enstrophy instead of vorticity because we are primarily interested in understanding processes that control the climatology of AEW activity, rather than the dynamics of individual waves, so a positive definite quantity is preferred. Many previous studies have used perturbation kinetic energy to study AEWs (Maloney and Dickinson 2003; Leroux et al. 2010; Diaz and Aiyyer 2013a; Alaka and Maloney 2014), which requires analyzing the perturbation available potential energy budget to infer the influence of convection on AEW dynamics. Since we are focused on the role of convection and vertical velocity, the vortex stretching terms in the budget of perturbation enstrophy below provide a more direct diagnostic tool.

The approximate budget of the vertical component of perturbation vorticity can be written as

$$\begin{aligned} \frac{\partial \zeta'_z}{\partial t} = & -\bar{\mathbf{u}} \cdot \nabla \zeta'_z - \mathbf{u}' \cdot \nabla \bar{\zeta}_z - (\mathbf{u}' \cdot \nabla \zeta'_z)' - \left(\hat{\mathbf{k}} \cdot \nabla \omega \times \frac{\partial \mathbf{u}}{\partial p} \right)' \\ & + \zeta'_z \frac{\partial \bar{\omega}}{\partial p} + \bar{\zeta}_z \frac{\partial \omega'}{\partial p} + \left(\zeta'_z \frac{\partial \omega'}{\partial p} \right)' + \frac{\partial F'_y}{\partial x} - \frac{\partial F'_x}{\partial y}, \end{aligned} \quad (3)$$

where $\mathbf{u}$ is the horizontal wind vector, ω is the vertical velocity, and $\mathbf{F} = \hat{i}F_x + \hat{j}F_y$ represents mechanical forcing

from unresolved processes such as friction, gravity waves, and convective momentum transport. Note that nonlinear terms have been retained, but their magnitudes are generally small.

The budget of the perturbation enstrophy can be obtained by multiplying (3) by ζ'_z .

$$\begin{aligned} \frac{\partial \xi'_z}{\partial t} = & -\bar{\mathbf{u}} \cdot \nabla \zeta'_z - \zeta'_z \mathbf{u}' \cdot \nabla \bar{\zeta}_z - \zeta'_z (\mathbf{u}' \cdot \nabla \zeta'_z)' \\ & - \zeta'_z \left(\hat{\mathbf{k}} \cdot \nabla \omega \times \frac{\partial \mathbf{u}}{\partial p} \right)' + 2\zeta'_z \frac{\partial \bar{\omega}}{\partial p} + \zeta'_z \bar{\zeta}_z \frac{\partial \omega'}{\partial p} \\ & + \zeta'_z \left(\zeta'_z \frac{\partial \omega'}{\partial p} \right)' + \zeta'_z \frac{\partial F'_y}{\partial x} - \zeta'_z \frac{\partial F'_x}{\partial y}. \end{aligned} \quad (4)$$

The first line represents advection of perturbation enstrophy by the mean flow. Terms in the second line describe the eddy flux of vorticity. The first of these terms represents the exchange of enstrophy between the mean flow and the transient eddies, similar to the barotropic energy conversion term found in the perturbation kinetic energy budget (Lau and Lau 1992). This term is positive when the eddy vorticity flux is directed down the background absolute vorticity gradient. The third line describes the effects of vortex tilting. The tilting term can be expanded as

$$\begin{aligned} \left(\hat{\mathbf{k}} \cdot \nabla \omega \times \frac{\partial \mathbf{u}}{\partial p} \right)' = & \frac{\partial v'}{\partial p} \frac{\partial \bar{\omega}}{\partial x} + \frac{\partial \bar{v}}{\partial p} \frac{\partial \omega'}{\partial x} + \left(\frac{\partial v'}{\partial p} \frac{\partial \omega'}{\partial x} \right)' \\ & - \frac{\partial u'}{\partial p} \frac{\partial \bar{\omega}}{\partial y} - \frac{\partial \bar{u}}{\partial p} \frac{\partial \omega'}{\partial y} - \left(\frac{\partial u'}{\partial p} \frac{\partial \omega'}{\partial y} \right)'. \end{aligned} \quad (5)$$

The resulting terms above are generally small, with the exception of the fifth term of (5) that involves the vertical gradient of background zonal wind, which can be large in the vicinity of a jet such as the AEJ.

The fourth line of (4) represents the generation of enstrophy by vortex stretching, which will be of central importance to the results. The final line describes enstrophy generation or destruction by mechanical forcing, such as friction and convective momentum transport. Because of the difficulty of estimating these sources and sinks from observational data, we will not be discussing them. The terms involving CMT in CESM with parameterized convection were found to be relatively small, but these may have a nonnegligible role in AEW dynamics that will be investigated in future work. Even if the mechanical forcing terms were measurable, the unfortunate reality of a data-sparse region such as North Africa is that we cannot expect a small enstrophy budget residual. The residual of model data is naturally much smaller than reanalysis, but there is not much value to be gained from analyzing it, so it will not be discussed here.

FIG. 1. Mean boreal summer surface temperature (colors) and 850-hPa wind (vectors) for (a) ERAi, (b) SP1xCO₂, and (c) SP4xCO₂.

To simplify the discussion we will use an abbreviated budget by grouping the terms in each line of (4) as

$$\frac{\partial \xi'_z}{\partial t} = \text{ADV} + \text{FLX} + \text{TLT} + \text{STR} + \text{ENF}. \quad (6)$$

As an aside, one may be tempted to use potential vorticity (PV) as the basis of a perturbation budget analysis. However, this introduces a problem when comparing the variability of two atmospheres with substantially different static stabilities. To illustrate, if we consider the Ertel PV ($P_z = -\zeta_z \partial \theta / \partial p$) then we can expand the vertical component of perturbation PV as

$$P'_z = -g(f + \bar{\zeta}_z) \frac{\partial \theta'}{\partial p} - g \zeta'_z \frac{\partial \bar{\theta}}{\partial p} - g \left(\zeta'_z \frac{\partial \theta'}{\partial p} \right)'.$$

Tropical static stability increases with warming due to the moist adiabatic lapse rate response so the magnitude of the second term in the above equation, which dominates the overall magnitude, will increase. This can give the misleading impression that perturbation PV variance has dramatically increased with warming even if the variances of ζ'_z and θ' have stayed the same. For this reason we have focused our analysis on the perturbation enstrophy budget.

3. Simulation results

a. Climate comparison

Mean July–September (JAS) surface temperature and 850-hPa wind vectors are shown in Fig. 1 for ERAi, SP1x, and SP4x. Results for ZM1x and ZM4x are similar (see Fig. S1 in the online supplemental material). We expect the surface temperature in SP1x to be colder than ERAi because a preindustrial level of CO₂ was specified. However, the fact that SP4x has a similar surface temperature magnitude to the present-day ERAi data shows that model suffers from the same North African cold bias

identified in previous studies (McCrary et al. 2014a; Arnold et al. 2015). There are several contributing factors to this cold bias, including biases in surface evaporation and albedo. The temperature bias is particularly strong over the Sahara because of higher than observed moisture and the presence of low clouds in both models (not shown). The Saharan cold bias is slightly reduced in the SP-CESM relative to CESM, but in both cases the temperature lapse rate over the Sahara does not resemble the deep, well-mixed boundary layer dominated by dry convection that is found in observations.

The model SST pattern also exhibits a bias such that the equatorial cold tongue is absent. Biases in the SST pattern likely contribute to the exaggerated low-level westerly monsoon flow centered around 10°N. These temperature biases also affect the meridional temperature gradients that control the position and strength of the midtropospheric jet (Thorncroft and Blackburn 1999) and thus likely affect the monsoon variability. In spite of this issue, the model response to elevated CO₂ shows many of the expected features, such as intense warming of the Arctic, Sahara, and tropical upper troposphere (see Figs. S2 and S3) and thus can still be informative for understanding how the real African monsoon system will respond.

Figure 2 shows maps of JAS mean tropical precipitation for TRMM and the two control simulations (Figs. 2a,b,d) along with the difference between corresponding 4xCO₂ and 1xCO₂ experiments (Figs. 2c,e). Many regional biases can be found in both models, including a double intertropical convergence zone (ITCZ) in the central Pacific and overly weak precipitation in the west and east Pacific. ITCZ precipitation is weak in all basins for SP1x. Along the south coast of West Africa SP1x and ZM1x show respectively more and less precipitation compared to TRMM data, consistent with McCrary et al. (2014a). Comparing the 1xCO₂ and 4xCO₂ simulations, there is increased precipitation near the equator and a decrease off the

FIG. 2. (a),(b),(d) Mean boreal summer precipitation and (c),(e) precipitation difference between 4xCO₂ and 1xCO₂ experiments overlaid with contours of mean precipitation from the corresponding 1xCO₂ experiment with a contour interval of 1 mm day⁻¹. Units are mm day⁻¹.

equator in both models, suggesting an equatorial contraction of the ITCZ. Precipitation in SP4x also becomes slightly more spatially diffuse in the Atlantic ITCZ.

The midtropospheric African easterly jet is an important feature for weather variability over West Africa.

The existence of the AEJ relies on heating by moist convection in the ITCZ and dry convection over the Sahara (Thorncroft and Blackburn 1999). Figure 3 shows zonal wind and apparent diabatic heating (Q_1 ; Yanai et al. 1973) zonally averaged over the red box

FIG. 3. Zonal wind (contours) and diabatic heating (Q_1 ; colors) zonally averaged over -20° – 10° E. Units are m s^{-1} for wind and K day^{-1} for Q_1 . Thicker contours indicate zonal wind values greater or equal to 10 m s^{-1} . Wind contours with magnitudes less than 5 m s^{-1} are omitted for clarity.

(-20° – 10° E) in Fig. 2 for ERAi, SP1x, and ZM1x. Here we define Q_1 as the residual of the dry static energy budget:

$$\frac{\partial s}{\partial t} + \mathbf{u} \cdot \nabla s + \omega \frac{\partial s}{\partial p} = Q_1,$$

where $s = c_p T + gz$ is the dry static energy.

There are several important differences between the datasets in Fig. 3. The AEJ is centered at 600 hPa and 12° N in ERAi data, but both models show the jet center farther north around 16° N and at a higher altitude around 500 hPa. Also notice the strong monsoon westerlies at the surface in both models, which is not found in observations.

The pattern of Q_1 in both models is also different from ERAi data (Fig. 3). The ITCZ heating profile ($\sim 10^{\circ}$ N) is trimodal for ERAi, unimodal for SP1x, and bimodal for ZM1x. Over the Sahara (20° – 30° N) there is a net cooling in both models, whereas ERAi shows a net warming due to dry convective heat fluxes. The bias in the location of

the AEJ is related to the biases in Q_1 because these processes have a large influence on the midtropospheric meridional temperature gradient which sets the jet location (Thorncroft and Blackburn 1999).

Figure 4 shows the difference of Q_1 (colors) between the $4\times\text{CO}_2$ and $1\times\text{CO}_2$ simulations against the $1\times\text{CO}_2$ Q_1 (contours). Stippling indicates that the difference is statistically significant at the 95% confidence level using the conventional Student's t test. A prominent feature of the difference field is the increased heating at upper levels centered at 300 hPa, indicating that the deepest convection has become deeper on average, consistent with the FAT hypothesis. The reduced heating in the midtroposphere around 10° N in both models is consistent with an equatorial contraction of the ITCZ mentioned previously. The ITCZ position is influenced by several factors including the ocean circulation and extratropical clouds (Frierson and Hwang 2012; Frierson et al. 2013; Schneider et al. 2014), so a shift in the ITCZ with warming is a reasonable expectation, and is likely influenced by the global-scale biases in the models. The

FIG. 4. The difference of zonally averaged Q_1 (colors) overlaid with contours of Q_1 from the corresponding $1\times\text{CO}_2$ simulation. Units are K day^{-1} . Thicker contours indicate values greater or equal to 2 K day^{-1} . Stippling indicates statistically significant differences at the 95% confidence level.

changes in diabatic heating do not impart a substantial change to the main features of the zonal mean zonal wind pattern (see Fig. S4), but there is a slight increase in vertical shear that can affect the vortex tilting terms of the perturbation enstrophy budget (see section 4).

b. Precipitation variability

The annual cycle of monthly mean rainfall averaged over -10° – 20° N is shown in Fig. 5, with a dashed line at 5° N to indicate the approximate location of the southern coast of West Africa. The “monsoon jump” is a key feature of the African monsoon that can be seen in TRMM data (Fig. 5a). The jump describes when the maximum of precipitation shifts abruptly northward from the Guinea coast to the Sahel region around 10° N in late June or early July (Cook 2015). The existence of the monsoon jump involves several mechanisms, including the position of the Saharan heat low (Ramel et al. 2006), a northward shift of the AEJ (Sultan and Janicot 2000; Gu and Adler 2004), development of the Atlantic cold tongue (Nguyen et al. 2011), moisture transport (Thorncroft et al. 2011; Pu and Cook 2012; Flaounas et al. 2012), and inertial instability (Hagos and Cook 2007; Cook 2015).

McCrary et al. (2014a) found that the coarser-resolution SP-CCSM and CCSM did not exhibit a distinct jump of precipitation, but the SP-CCSM was more realistic in that it reproduced the movement of precipitation onto the continent during the monsoon, whereas the simulations here exhibit a different behavior. The boreal spring (March–May) precipitation is much farther south than observations (Fig. 5) associated with the SST biases discussed above. Precipitation in ZM1x and ZM4x is closer to observations in the sense that there is a distinct end of a southern precipitation maximum followed by a new precipitation maximum over the continent. The SP1x and SP4x simulations exhibit a more gradual northward shift.

Comparing the $1xCO_2$ and $4xCO_2$ simulations we find a general increase of precipitation intensity during the whole rainy season (June–September). This result seems to be at odds with previous studies of coupled model ensembles that find a delayed and shorter rainy season (e.g., Biasutti et al. 2009; Biasutti 2013), but a fair comparison of the timing of the rainy season would require further analysis that carefully considers the model seasonal biases in circulation and surface temperature, which is outside the scope of this study.

Figure 6 shows the frequency distribution of 3-hourly precipitation for TRMM and model data over a region that includes the Sahel, Guinea coast, and Atlantic ITCZ. The distribution is calculated

FIG. 5. Climatological monthly seasonal cycle of precipitation averaged zonally over -10° – 20° N. Units are $mm\ day^{-1}$.

FIG. 6. Boreal summer (JAS) rain frequency distribution (%) over 0° – 20° N and -40° – 40° E. Bins are logarithmically spaced starting at 0.03 with a width of 0.01 and increasing the width by 20% (see text).

using logarithmically spaced bins similar to Pendergrass and Hartmann (2014b), except that each successive bin is 20% wider instead of 7%. Parameterized convection is notorious for producing too much light rain (Sun et al. 2006), and a high occurrence of rain rates around 1 mm day^{-1} in Fig. 6 shows that the SP-CESM is not exempt from this issue. There is also less occurrence of heavy and extreme rainfall events (20 mm day^{-1} and up) in both models compared to TRMM.

The response of SP-CESM to CO_2 reveals a decrease in moderate rain events (2 – 20 mm day^{-1}) and an increase in light and heavy to extreme events (red and blue lines in Fig. 6). The CESM response is similar, but there

is much less increase in the occurrence of heavy events, and a broader range of moderate precipitation rates that exhibit a decrease in frequency. This pattern of change in this regional and seasonally restricted distribution is consistent with other studies using global CMIP5 data (Sun et al. 2007; Lau et al. 2013; Pendergrass and Hartmann 2014a).

The global mean precipitation increase from warming by CO_2 is expected from energetic constraints (Allen and Ingram 2002; Held and Soden 2006; Pendergrass and Hartmann 2014a), but the increase in heavy and extreme precipitation that accompanies the increase in variance is expected to be driven by changes in moisture convergence (Trenberth 1999). Figure 7 shows the distribution of moisture convergence similar to Fig. 6, except that the bins are extended to included negative values. The nature of logarithmically spaced bins does not allow the distribution to be continuous across the range of negative and positive bins. Similar to the precipitation distribution, there is an increase in the occurrence of extreme moisture convergence events (greater than 20 mm day^{-1}) with warming, as well as an increase of extreme negative events.

The change of the precipitation distribution toward more heavy and light rain events also means that there is an increase in the overall precipitation variance. A substantial portion of the precipitation variance over Africa is organized by easterly waves with periods of 2–10 days and synoptic wavenumbers 4–20. To examine this we use a frequency–wavenumber filtering method to isolate these scales for westward propagating signals,

FIG. 7. Frequency distribution of column moisture convergence (%) for boreal summer (JAS) over 0° – 20° N and -40° – 40° E. Positive and negative bins are logarithmically spaced starting at 1 and -1 with a width of 0.4 and increasing the width by 20% (see text).

FIG. 8. Zonal profiles of TD filtered precipitation (see text) for boreal summer (JAS) meridionally averaged over 0° – 20° N. Black bars along the abscissa indicate the approximate location of equatorial landmasses.

which is often referred to as the tropical depression (TD) band (Kiladis et al. 2006).

Figure 8 shows the variance of TD filtered precipitation for TRMM and model data averaged meridionally over 0° – 20° N for all longitudes. This view of precipitation variability highlights three regions of activity: the east Pacific, North Africa, and the west Pacific. TD band variance is largest in the west Pacific and reasonably captured by SP1x, but is weak in both ZM1x and ZM4x. Both models produce too little variance in the east Pacific around 100° – 120° E. Over the Atlantic and West Africa the TD variance peaks just off the coast near 30° W in observations. ZM1x and ZM4x reproduce this pattern well, albeit with less amplitude. SP1x and SP4x produce a peak variance farther east around -20° W. TD variance near Africa generally increases for both models with warming, but the response is much larger in SP4x and very subtle in ZM4x. This suggests that there may be more African easterly wave activity in a warmer climate. However, as we will show in the next section, a dynamical perspective leads to the opposite conclusion of reduced activity so the increase

in TD variance may not be a valid indicator of the change in wave activity. This discrepancy will be explored in the next section.

4. African easterly waves

a. Climatology

To examine the activity of African easterly waves we will use the perturbation enstrophy (PEN) to understand the model response to warming from CO_2 . Figure 9 shows the time mean PEN zonally averaged over -20° – 10° E for ERAi, SP1x, and ZM1x. Both storm tracks are evident in ERAi data centered at 10° and 18° N, and the northern track is much more intense than the southern track. The storm track intensity is somewhat sensitive to the area of averaging, with the southern track becoming stronger farther to the west over the Atlantic, but the current averaging region captures the main AEW growth region. A similar picture can be obtained using eddy kinetic energy (EKE) as a tracer of AEW activity.

The two storm tracks are clearly visible in both models, but the northern track is much weaker than observations and the southern track is much stronger (Figs. 9b,c). The weaker northern track is due to a lack of baroclinic conversion associated with the Saharan cold bias (McCrary et al. 2014b). The southern track is not only more intense, but is also at a higher altitude than observed due to the altitude bias of the AEJ.

Figure 10 shows the PEN difference between the $4\times\text{CO}_2$ and $1\times\text{CO}_2$ simulations against PEN of the corresponding $1\times\text{CO}_2$ simulation in contours. Stippling indicates statistically significant difference at the 95% confidence level. Both models exhibit a reduction of AEW activity in the southern track centered

FIG. 9. Boreal summer mean perturbation enstrophy (colors) and zonal wind (contours) zonally averaged over -20° to 10° E. Thicker contours indicate zonal wind values greater or equal to 10 m s^{-1} . Zonal wind contours with magnitudes less than 5 m s^{-1} are omitted for clarity.

FIG. 10. Boreal summer mean difference of zonal mean perturbation enstrophy (shading) against the zonal mean zonal wind (contours). Stippling indicates statistically significant differences at the 95% confidence level. Thicker contours indicate zonal wind values greater or equal to 10 m s^{-1} . Wind contours with magnitudes less than 5 m s^{-1} are omitted for clarity.

at 10°N . The northern track in SP4x shows a clear increase in AEW activity around 20°N (Fig. 10a), consistent with previous studies (Skinner and Diffenbaugh 2014; Martin and Thorncroft 2015). The northern track response in ZM4x is less clear (Fig. 10b), but there is a notable increase in PEN around 23°N at 900 hPa and a decrease around 18°N , suggesting a northward shift of the northern track as the Sahara warms.

From Fig. 10 it is unclear whether the changes of AEW activity are due to changes in AEW number or intensity. To investigate this we can examine the

distribution of AEW event intensity using a simple 2–10-day eddy vorticity index. The index is calculated by applying a 2–10-day Lanczos filter to vorticity at 500 hPa for model data and 600 hPa for ERAi data. The filtered vorticity is then averaged over select continental and maritime regions along the southern track. All local maxima of the time series that occur during the summer months (JAS) are identified and used to calculate the distribution, which is then is scaled to represent the average number of events per year as shown in Fig. 11. Filled circular markers

FIG. 11. Distribution of positive perturbation vorticity events (see text). The perturbation vorticity was averaged over (a) $5^\circ\text{--}15^\circ\text{N}$, $-25^\circ\text{--}-15^\circ\text{E}$ and (b) $5^\circ\text{--}15^\circ\text{N}$, $-5^\circ\text{--}5^\circ\text{E}$. Units are $\text{s}^{-1} \times 10^{-6}$. Filled markers denote a statistically significant difference between the binned averages of the corresponding SP-CESM or CESM data at the 95% confidence level.

FIG. 12. ERAi boreal summer mean terms of the abbreviated perturbation enstrophy budget [(6)] in colors overlaid with contours of zonal wind. The color scale is exponential in the negative and positive direction increasing by powers of 2 from 2^2 to 2^9 . Thicker contours indicate zonal wind values greater or equal to 10 m s^{-1} . Zonal wind contours with magnitudes less than 5 m s^{-1} are omitted for clarity.

indicate when the difference between corresponding $4\times\text{CO}_2$ and $1\times\text{CO}_2$ simulations are statistically significant at the 95% confidence level.

In both regions SP4x exhibits a notable decrease in the occurrence of weak events and a slight increase in the occurrence of strong events (Fig. 11). The increase in the frequency of strong events may be heavily influenced by model bias and not systematic, as discussed in section 4c. ZM4x also shows a similar shift, albeit weaker. Results for the northern track AEW distribution are mixed and do not indicate a coherent shift (not shown), suggesting that the mechanism responsible for increased wave activity in Martin and Thorncroft (2015) does not favor a particular scale. If we consider the total number of AEWs there is a statistically significant reduction of 1–2 events per year in both models and both regions. The change in the average intensity per event

increases slightly, but the change is not statistically significantly.

b. The perturbation enstrophy budget

The reduced occurrence of weak AEWs appears to explain the reduction of perturbation enstrophy in the southern track (Fig. 10), but this begs the question of why weak AEWs are less frequent. To answer this, we will employ the perturbation enstrophy budget in (4). Figure 12 shows the four main groupings of terms from (6) for ERAi data averaged over July–September and -20° – 10°E . Because of the large range of magnitudes in these terms an exponential color scale is used with values of the form 2^n with $n \in \{2, \dots, 9\}$. The vortex stretching term is clearly the dominate process in both storm tracks, but there is a notable contribution from the eddy flux and tilting terms. The pattern of advective tendencies is strongly influenced by the vertical component of

FIG. 13. As in Fig. 12, but for SP1x.

advection, which moves low-level PEN upward in both tracks. Detailed analysis of the stretching terms revealed that the magnitude of these terms may be exaggerated in ERAi data due to issues associated with the topography of the Ahaggar and Adrar mountain ranges (not shown). Simply masking out data that lie underground does not remedy the issue, so it is likely rooted in the lack of observations in these regions.

Figure 13 shows the budget of perturbation enstrophy for SP1x similar to Fig. 12. The balance of PEN tendencies in SP1x is very different from ERAi, but there are many qualitative similarities. The northern track waves are driven by the eddy flux and vortex stretching terms, similar to ERAi. The southern track indicates a role for all terms, but there is less influence from vortex stretching and a larger influence of tilting and eddy flux terms compared to ERAi. The balance of terms in ZM1x is similar to SP1x (not shown).

Figure 14 shows the difference of perturbation enstrophy terms between SP4x and SP1x. The change in

advection is mostly due to a reduction of vertical advection but the overall magnitude of the difference is small (Fig. 14a). The change in eddy flux terms is also small, and mostly due to a reduction in the meridional eddy flux (Fig. 14b). The stretching and tilting terms (Figs. 14c,d) are the biggest factors that explain the change in the southern track PEN in Fig. 10a. The stretching tendency magnitudes are strongly reduced near the surface around 10°N where the enstrophy generation in SP1x is largest, as well as aloft near 400 hPa and 5°N where PEN destruction occurs (see Fig. 13c). The reduction of PEN generation near the surface is relevant to the midtropospheric wave track because of the vertical advective tendency that communicates this to the midlevels (Fig. 13a). The tilting terms also indicate a reduction of enstrophy generation around 10°–15°N and 650–800 hPa that is related to the contraction of the ITCZ (Fig. 4) as well as the WTG mechanism, which both act to reduce the meridional gradient of perturbation vertical velocity in

FIG. 14. Difference of zonal mean perturbation enstrophy budget terms between SP4x and SP1x.

the dominant tilting term in (5). The differences between ZM4x and ZM1x are similar but smaller in magnitude (not shown).

In spite of the significant model biases, the changes to the vortex stretching terms in Fig. 14c are relevant for how observed AEWs will respond to a warmer climate. We can directly test the hypothesized WTG mechanism by considering the profile of STR for a fixed amount of column integrated diabatic heating, which is roughly equal to the precipitation rate. According to the WTG balance condition in (1), a fixed amount of heating and increased stratification will result in reduced vertical velocity. This will reduce STR through a reduction of the vertical gradient of perturbation vertical velocity. Figure 15 illustrates this by conditionally averaging profiles of STR and ω by the precipitation rate over the West African ITCZ region (0° – 20° N, -20° – 10° E). Comparing SP1x and SP4x there is a clear reduction of the magnitude of both terms with increased CO_2 and warming. A similar plot of Q_1 is virtually unchanged is

virtually unchanged except for a small upward shift consistent with the FAT hypothesis (not shown), which is expected since the vertically integrated diabatic heating should be approximately equal to the precipitation. This supports the hypothesized consequence of WTG, since a smaller vertical velocity is needed to balance any given diabatic heating with increased stratification from the lapse rate response. In the next section we will discuss how this mechanism manifests in composites of individual AEWs.

c. AEW composite analysis

To further illustrate the processes driving the AEW distribution shift we calculate AEW composites using an index of 500-hPa perturbation vorticity averaged over the red boxes on the inset maps in Fig. 11, which are 10° wide boxes centered at -20° and 0° E. To compare weak and strong waves, events are selected using ranges of 500-hPa perturbation vorticity of $4 < (\zeta'_z \times 10^6) \leq 12$ for weak waves and $12 < (\zeta'_z \times 10^6) \leq 20$ for strong waves.

FIG. 15. Profiles of (top) STR and (bottom) ω conditionally averaged against precipitation rate in mm day^{-1} .

Figure 16 shows weak and strong AEW composites for the region centered at 0°E of 2–10-day filtered Q_1 (shading) and 2–10-day meridional wind (contours) averaged meridionally over $5^\circ\text{--}15^\circ\text{N}$ for SP1x and SP4x. Both weak and strong waves in SP4x are associated with stronger peak diabatic heating relative to SP1x, but the circulation near the surface is notably weaker (Figs. 16b,d). The weaker circulation is most prominent in the northerlies to the west, ahead of the vorticity center, where the 2–10-day meridional wind is reduced by roughly 1 m s^{-1} . There is also a noticeable upward shift in the peak diabatic heating, consistent with the FAT hypothesis. Figure 17 shows similar composites of vertical pressure velocity (ω). The composite ω structure shows a similar upward shift in SP4x, but unlike diabatic heating ω is reduced in magnitude consistent with Fig. 15. Composites profiles over the Atlantic region are similar (not shown).

To simplify the comparison of composite perturbation enstrophy budget profiles of the dominant terms (FLX,

TLT, STR) and their sum are averaged over the two regions used to define the composite indices in Figs. 18 and 19. Profiles are dashed for SP1x and solid for SP4x. Focusing first on the total tendency in the black lines for weak waves (Fig. 18a) there is a substantial reduction of the enstrophy tendency at low levels and a slight increase at 400 hPa. The strong wave composite shows a stronger total tendency around 400–500 hPa (Fig. 18b). All three terms contribute substantially to the changes in the total tendency, but there is a lack of consistency across weak and strong waves. There is also some cancellation that complicates the interpretation, but the change in the stretching tendency for weak waves is consistent with the expected change due to WTG theory (Fig. 15).

A more consistent picture can be seen for AEWs that have moved off the coast of West Africa (Fig. 11a). Figure 19 shows profiles of composite perturbation enstrophy budget terms for the maritime region centered

FIG. 16. AEW composite of Q_1 and 2–10-day filtered meridional wind meridionally averaged over 5° – 15° N for (top) weak and (bottom) strong AEW events (see text). Units are K day^{-1} for Q_1 and m s^{-1} for meridional wind.

on -20°E . The total tendency for both weak and strong waves is somewhat similar to Fig. 18, but both exhibit a clear reduction of the low-level enstrophy generation due to the change in the vortex stretching term. The midlevel negative stretching tendency around 500–800 hPa is also reduced in magnitude, consistent with the hypothesized WTG mechanism (Fig. 15).

The increase in magnitude of the composite tilting component appears to be responsible for the increased frequency of strong waves shown in Fig. 11. However, further analysis failed to uncover a clear explanation or mechanism for this increase. The tilting terms are strongly dependent on gradients of the wind field, which in turn are sensitive to the surface temperature gradients that are influenced by model biases. This suggests that the changes to the tilting terms are likely influenced by the model biases and not systematic. So although the changes in the tilting term can be quite large (Fig. 19b), they are less useful for understanding the real world climate response to CO_2 as long as these biases are large.

In spite of the altered balance of the perturbation enstrophy budget the general AEW structure is mostly unchanged with warming, except for a weaker low-level circulation that might be important for tropical cyclogenesis. Lag composite analysis shows that AEW propagation speed is also unchanged (not shown). The contrast of increased diabatic heating with altered dynamic wave forcing explains the seemingly contradictory results that AEW precipitation increases (Figs. 8 and 16) while the dynamic signature of the southern AEW track is reduced (Fig. 10).

5. Is WTG balance valid for AEWs?

Figures 15–17 are consistent with the WTG mechanism described here, but we have not discussed the validity of applying WTG balance to AEWs. The WTG approximation requires a sufficiently long time scale and weak planetary vorticity to ensure that horizontal gradients of temperature and pressure are relatively small. AEWs

FIG. 17. As in Fig. 16, but for 2–10-day filter ω . Units are Pa s^{-1} for Q_1 and m s^{-1} for meridional wind.

occur away from the equator, on synoptic time scales, and in a region known for strong meridional surface temperature gradients. All of these factors complicate the justification of invoking the WTG approximation.

Let us return to the budget of dry static energy, rewritten such that it will be approximately zero when WTG balance is valid:

$$0 \approx \left(\frac{\partial s}{\partial t} + \mathbf{u} \cdot \nabla s \right) = \left(Q_1 - \omega \frac{\partial s}{\partial p} \right). \quad (7)$$

Note that the right-hand side equality above is exact since Q_1 is defined as a residual. We will refer to the value of either side of (7) as the WTG residual.

To investigate the present conundrum we need to assess whether vertical advection and diabatic heating are “similar enough” in magnitude to justify whether WTG balance is a valid approximation. Wolding et al. (2016) considered this question using data near the equator and concluded that the WTG approximation is valid at intra-seasonal and longer time scales. However, their analysis only considered a single subjective threshold for defining

when the contribution of horizontal advection and storage is “small.” Here we will consider a slightly different approach to characterize the value of (7).

We start by filtering each side of (7) with a low-pass Lanczos filter and taking the root-mean-square (RMS) to obtain a characteristic value of the WTG residual:

$$0 \approx \left\{ \overline{\left[\frac{\partial s'}{\partial t} + (\mathbf{u} \cdot \nabla s)' \right]^2} \right\}^{1/2} = \left\{ \overline{\left[Q_1' - \left(\omega \frac{\partial s}{\partial p} \right)' \right]^2} \right\}^{1/2}, \quad (8)$$

where primes denote filtered quantities and overbars represent a time average. A low-pass filter is used because WTG balance is known to be valid on intra-seasonal and longer time scales, so as the cutoff frequency is increased we expect the value of (8) to steadily increase and reveal a time scale where the WTG approximation becomes invalid. We also expect this characteristic value to become steadily larger with latitude. Additionally, we can normalize the characteristic WTG residual in (8) by the RMS value of filtered Q_1 ,

FIG. 18. Composite profiles of select perturbation enstrophy budget terms and their sum averaged over (5° – 15° N, -5° – 5° E) for (a) weak and (b) strong AEW events. SP1x data are shown as dashed lines and SP4x data are shown as solid. Units are $s^{-2} \times 10^{-12}$.

allowing us to interpret the WTG residual in proportion to the characteristic diabatic heating:

$$0 \approx \frac{\left\{ \left[\frac{\partial s'}{\partial t} + (\mathbf{u} \cdot \nabla s)' \right]^2 \right\}^{1/2}}{\{[\overline{Q_1'}]^2\}^{1/2}} = \frac{\left\{ \left[Q_1' - \left(\omega \frac{\partial s'}{\partial p} \right)' \right]^2 \right\}^{1/2}}{\{[\overline{Q_1'}]^2\}^{1/2}}. \quad (9)$$

Estimates of the right-hand side of (8) and (9) are shown in Figs. 20a and 20b as a function of the low-pass filter cutoff frequency and latitude using 5 years of 500-hPa

ERAi data (2000–04) from 30° W to 30° E. Results for other pressure levels are similar. The hatched boxes show the approximate time scales (2–6 days) and latitudes (5° – 15° N) associated with the convective signal of AEWs in the southern track (Mekonnen et al. 2006). From Fig. 20a we can see that the WTG residual has a characteristic value of 0.5–1.0 J, but it is not clear whether this can be considered small. The normalized WTG residual shown in Fig. 20b indicates that the RMS difference is roughly 20%–30% of the characteristic diabatic heating rate south of 10° N and increases rapidly to the north. To the extent that 20%–30% is considered a small proportion, we can conclude that

FIG. 19. As in Fig. 18, but for the region maritime region (5° – 15° N, -25° – 15° E).

FIG. 20. Characteristic value of (a) the WTG residual and (b) the normalized WTG residual (see text) as a function of low-pass filter cutoff frequency and latitude using ERAi data at 500 hPa over -30° – 30° E.

WTG balance is a reasonable approximation for the southern portion of waves in the southern track. This seems like a reasonable conclusion given that normalized WTG residual is approximately 20% for longer time scales (10–15 days) near the equator. However, this result also shows that the WTG approximation is certainly not valid north of 10° N. This mix of dynamical regimes may be why some previous studies have had been able to invoke quasigeostrophic theory to investigate AEW dynamics (Kiladis et al. 2006).

6. Conclusions

The goal of this paper was to explore the response of variability in the West African monsoon (WAM) system, and specifically African easterly waves (AEWs), to a climate warmed by a quadrupling of atmospheric CO_2 . The superparameterized CESM (SP-CESM) was employed to overcome the deficient convective coupling exhibited by models with conventional convective parameterizations. The climatology of the West African region in both CESM and SP-CESM exhibits some notable biases, such as substantial differences in the Atlantic SST pattern and land surface temperatures that are colder than observed. In spite of these biases the SP-CESM produces robust AEW activity.

The results indicate an overall higher variance of precipitation, with more occurrence of extreme rain events associated with an increase in extreme moisture convergence events. The variance of precipitation filtered to

represent the time scale of AEW also increases, suggesting more AEW activity south of the African easterly jet (AEJ) in response to warming.

An investigation of AEW activity with the perturbation enstrophy budget suggests a more active northern AEW track, as well as the counterintuitive result of less activity in the southern track. The increase in the northern track activity is due to increased baroclinic energy conversion (not shown), similar to previous studies (Skinner and Diffenbaugh 2014; Martin and Thorncroft 2015).

The response of the southern track is not explained by any single term of the perturbation enstrophy budget, but the biggest changes are found in the terms that describe vertical vortex stretching and a vortex tilting term that involves the mean zonal wind and perturbation vertical velocity. The common factor between the changes of these two terms with warming can be understood as a consequence of weak temperature gradient balance (WTG). Warming induced by elevated CO_2 increases the temperature stratification and leads to a reduction of the vertical velocity required to balance any diabatic heating with adiabatic cooling. This mechanism leads to weaker vertical velocities that reduce the magnitude of certain perturbation enstrophy sources. An analysis of the WTG balance approximation suggests that this mechanism can be applied to the portion of waves that lie south of 10° N where much of the convective activity occurs, but the northern portion of the waves is likely governed by different dynamical constraints.

A preliminary examination of easterly wave activity in the west and east Pacific regions (Fig. 8) suggests a consistent response of increased diabatic heating and reduced wave circulations associated with weaker vortex stretching. This mechanism may also be relevant for understanding the response of equatorial Rossby waves or tropical intrusions of midlatitude Rossby waves. These questions are planned for the subject of future work.

The results here support the notion that precipitation events over West Africa that are currently considered extreme can be expected to become more frequent in a climate warmed by CO₂. The results also suggest that the WTG mechanism that reduces low-level vortex stretching may have a significant impact on the Atlantic tropical cyclone (TC) frequency by impeding the ability to develop a coherent low-level circulation. A caveat to this speculation is that the application of WTG may breakdown on the time scale that TC genesis occurs. Other changes to the climate of the Atlantic TC development region, such as SST or vertical shear, may overwhelm the impact of this mechanism. The uncertain balance of these competing effects means that it is still difficult to speculate on exactly how Atlantic TC activity will respond to warming. Additionally, there is the unresolved question of whether AEWs are even necessary for explaining the Atlantic TC season statistics (Emanuel 2008; Knutson et al. 2010; Russell et al. 2016).

Acknowledgments. This work was supported by a National Science Foundation (NSF) Atmospheric and Geospace Sciences postdoctoral research fellowship under Grant 1433343, as well as NSF Grant 1433763.

REFERENCES

- Alaka, G. J., and E. D. Maloney, 2014: The intraseasonal variability of African easterly wave energetics. *J. Climate*, **27**, 6559–6580, doi:10.1175/JCLI-D-14-00146.1.
- Allen, M. R., and W. J. Ingram, 2002: Constraints on future changes in climate and the hydrologic cycle. *Nature*, **419**, 224–232, doi:10.1038/nature01092.
- Allen, R. J., and S. C. Sherwood, 2008: Warming maximum in the tropical upper troposphere deduced from thermal winds. *Nat. Geosci.*, **1**, 399–403, doi:10.1038/ngeo208.
- Andersen, J. A., and Z. Kuang, 2012: Moist static energy budget of MJO-like disturbances in the atmosphere of a zonally symmetric aquaplanet. *J. Climate*, **25**, 2782–2804, doi:10.1175/JCLI-D-11-00168.1.
- Arnold, N. P., M. Branson, Z. Kuang, D. A. Randall, and E. Tziperman, 2015: MJO intensification with warming in the superparameterized CESM. *J. Climate*, **28**, 2706–2724, doi:10.1175/JCLI-D-14-00494.1.
- Biasutti, M., 2013: Forced Sahel rainfall trends in the CMIP5 archive. *J. Geophys. Res. Atmos.*, **118**, 1613–1623, doi:10.1002/jgrd.50206.
- , and A. Giannini, 2006: Robust Sahel drying in response to late 20th century forcings. *Geophys. Res. Lett.*, **33**, L11706, doi:10.1029/2006GL026067.
- , and A. H. Sobel, 2009: Delayed Sahel rainfall and global seasonal cycle in a warmer climate. *Geophys. Res. Lett.*, **36**, L23707, doi:10.1029/2009GL041303.
- , H. Sobel, and S. J. Camargo, 2009: The role of the Sahara low in summertime Sahel rainfall variability and change in the CMIP3 models. *J. Climate*, **22**, 5755–5771, doi:10.1175/2009JCLI2969.1.
- Boer, G., 1993: Climate change and the regulation of the surface moisture and energy budgets. *Climate Dyn.*, **8**, 225–239, doi:10.1007/BF00198617.
- Burpee, R., 1972: The origin and structure of easterly waves in the lower troposphere of North Africa. *J. Atmos. Sci.*, **29**, 77–90, doi:10.1175/1520-0469(1972)029<0077:TOASOE>2.0.CO;2.
- Charney, J., 1963: A note of large-scale motions in the tropics. *J. Atmos. Sci.*, **20**, 607–609, doi:10.1175/1520-0469(1963)020<0607:ANOLSM>2.0.CO;2.
- , and M. Stern, 1962: On the stability of internal baroclinic jets in a rotating atmosphere. *J. Atmos. Sci.*, **19**, 159–172, doi:10.1175/1520-0469(1962)019<0159:OTSOIB>2.0.CO;2.
- Chen, T.-C., 2006: Characteristics of African easterly waves depicted by ECMWF reanalyses for 1991–2000. *Mon. Wea. Rev.*, **134**, 3539–3566, doi:10.1175/MWR3259.1.
- , S.-Y. Wang, and A. J. Clark, 2008: North Atlantic hurricanes contributed by African easterly waves north and south of the African easterly jet. *J. Climate*, **21**, 6767–6776, doi:10.1175/2008JCLI2523.1.
- Cook, K. H., 2015: Role of inertial instability in the West African monsoon jump. *J. Geophys. Res. Atmos.*, **120**, 3085–3102, doi:10.1002/2014JD022579.
- , and E. K. Vizy, 2006: Coupled model simulations of the West African monsoon system: Twentieth- and twenty-first-century simulations. *J. Climate*, **19**, 3681–3703, doi:10.1175/JCLI3814.1.
- Dee, D., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, doi:10.1002/qj.828.
- DeMott, C. A., C. Stan, D. A. Randall, J. L. Kinter, and M. Khairoutdinov, 2011: The Asian monsoon in the superparameterized CCSM and its relationship to tropical wave activity. *J. Climate*, **24**, 5134–5156, doi:10.1175/2011JCLI4202.1.
- Diaz, M., and A. Ayyer, 2013a: Energy dispersion in African easterly waves. *J. Atmos. Sci.*, **70**, 130–145, doi:10.1175/JAS-D-12-019.1.
- , and —, 2013b: The genesis of African easterly waves by upstream development. *J. Atmos. Sci.*, **70**, 3492–3512, doi:10.1175/JAS-D-12-0342.1.
- Emanuel, K., 2008: The hurricane–climate connection. *Bull. Amer. Meteor. Soc.*, **89**, ES10–ES20, doi:10.1175/BAMS-89-5-Emanuel.
- Flaounas, E., S. Janicot, S. Bastin, and R. Roca, 2012: The West African monsoon onset in 2006: Sensitivity to surface albedo, orography, SST and synoptic scale dry-air intrusions using WRF. *Climate Dyn.*, **38**, 685–708, doi:10.1007/s00382-011-1255-2.
- Frierson, D. M. W., and Y.-T. Hwang, 2012: Extratropical influence on ITCZ shifts in slab ocean simulations of global warming. *J. Climate*, **25**, 720–733, doi:10.1175/JCLI-D-11-00116.1.
- , and Coauthors, 2013: Contribution of ocean overturning circulation to tropical rainfall peak in the Northern Hemisphere. *Nat. Geosci.*, **6**, 940–944, doi:10.1038/ngeo1987.
- Fu, Q., C. M. Johanson, S. G. Warren, and D. J. Seidel, 2004: Contribution of stratospheric cooling to satellite-inferred

- tropospheric temperature trends. *Nature*, **429**, 55–58, doi:[10.1038/nature02524](https://doi.org/10.1038/nature02524).
- Giannini, A., M. Biasutti, I. M. Held, and A. H. Sobel, 2008: A global perspective on African climate. *Climatic Change*, **90**, 359–383, doi:[10.1007/s10584-008-9396-y](https://doi.org/10.1007/s10584-008-9396-y).
- Grabowski, W. W., 2001: Coupling cloud processes with the large-scale dynamics using the cloud-resolving convection parameterization (CRCP). *J. Atmos. Sci.*, **58**, 978–997, doi:[10.1175/1520-0469\(2001\)058<0978:CCPWTL>2.0.CO;2](https://doi.org/10.1175/1520-0469(2001)058<0978:CCPWTL>2.0.CO;2).
- Gu, G., and R. F. Adler, 2004: Seasonal evolution and variability associated with the West African monsoon system. *J. Climate*, **17**, 3364–3377, doi:[10.1175/1520-0442\(2004\)017<3364:SEAVAW>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<3364:SEAVAW>2.0.CO;2).
- Hagos, S. M., and K. H. Cook, 2007: Dynamics of the West African monsoon jump. *J. Climate*, **20**, 5264–5284, doi:[10.1175/2007JCLI1533.1](https://doi.org/10.1175/2007JCLI1533.1).
- Hall, N. M. J., G. N. Kiladis, and C. D. Thorncroft, 2006: Three-dimensional structure and dynamics of African easterly waves. Part II: Dynamical modes. *J. Atmos. Sci.*, **63**, 2231–2245, doi:[10.1175/JAS3742.1](https://doi.org/10.1175/JAS3742.1).
- Hartmann, D. L., and K. Larson, 2002: An important constraint on tropical cloud–climate feedback. *Geophys. Res. Lett.*, **29**, 1951, doi:[10.1029/2002GL015835](https://doi.org/10.1029/2002GL015835).
- He, J., and B. J. Soden, 2015: Anthropogenic weakening of the tropical circulation: The relative roles of direct CO₂ forcing and sea surface temperature change. *J. Climate*, **28**, 8728–8742, doi:[10.1175/JCLI-D-15-0205.1](https://doi.org/10.1175/JCLI-D-15-0205.1).
- Held, I., and B. Soden, 2006: Robust responses of the hydrological cycle to global warming. *J. Climate*, **19**, 5686–5699, doi:[10.1175/JCLI3990.1](https://doi.org/10.1175/JCLI3990.1).
- Holloway, C. E., S. J. Woolnough, and G. M. S. Lister, 2012: Precipitation distributions for explicit versus parametrized convection in a large-domain high-resolution tropical case study. *Quart. J. Roy. Meteor. Soc.*, **138**, 1692–1708, doi:[10.1002/qj.1903](https://doi.org/10.1002/qj.1903).
- Hopsch, S. B., C. D. Thorncroft, K. Hodges, and A. Aiyer, 2007: West African storm tracks and their relationship to Atlantic tropical cyclones. *J. Climate*, **20**, 2468–2483, doi:[10.1175/JCLI4139.1](https://doi.org/10.1175/JCLI4139.1).
- Hsieh, J.-S., and K. H. Cook, 2007: A study of the energetics of African easterly waves using a regional climate model. *J. Atmos. Sci.*, **64**, 421–440, doi:[10.1175/JAS3851.1](https://doi.org/10.1175/JAS3851.1).
- , and —, 2008: On the instability of the African easterly jet and the generation of African waves: Reversals of the potential vorticity gradient. *J. Atmos. Sci.*, **65**, 2130–2151, doi:[10.1175/2007JAS2552.1](https://doi.org/10.1175/2007JAS2552.1).
- Huffman, G., and Coauthors, 2007: The TRMM Multisatellite Precipitation Analysis (TMPA): Quasi-global, multiyear, combined-sensor precipitation estimates at fine scales. *J. Hydrometeorol.*, **8**, 38–55, doi:[10.1175/JHM560.1](https://doi.org/10.1175/JHM560.1).
- Ingram, W., 2010: A very simple model for the water vapour feedback on climate change. *Quart. J. Roy. Meteor. Soc.*, **136**, 30–40, doi:[10.1002/qj.546](https://doi.org/10.1002/qj.546).
- Jones, C., N. Mahowald, and C. Luo, 2003: The role of easterly waves on African desert dust transport. *J. Climate*, **16**, 3617–3628, doi:[10.1175/1520-0442\(2003\)016<3617:TROEWO>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<3617:TROEWO>2.0.CO;2).
- Khairoutdinov, M., and D. Randall, 2001: A cloud resolving model as a cloud parameterization in the NCAR Community Climate System Model: Preliminary results. *Geophys. Res. Lett.*, **28**, 3617–3620, doi:[10.1029/2001GL013552](https://doi.org/10.1029/2001GL013552).
- , C. DeMott, and D. Randall, 2008: Evaluation of the simulated interannual and subseasonal variability in an AMIP-style simulation using the CSU multiscale modeling framework. *J. Climate*, **21**, 413–431, doi:[10.1175/2007JCLI1630.1](https://doi.org/10.1175/2007JCLI1630.1).
- Kiladis, G. N., C. D. Thorncroft, and N. M. J. Hall, 2006: Three-dimensional structure and dynamics of African easterly waves. Part I: Observations. *J. Atmos. Sci.*, **63**, 2212–2230, doi:[10.1175/JAS3741.1](https://doi.org/10.1175/JAS3741.1).
- Knutson, T. R., and Coauthors, 2010: Tropical cyclones and climate change. *Nat. Geosci.*, **3**, 157–163, doi:[10.1038/ngeo779](https://doi.org/10.1038/ngeo779).
- Kuang, Z., and D. L. Hartmann, 2007: Testing the fixed anvil temperature hypothesis in a cloud-resolving model. *J. Climate*, **20**, 2051–2057, doi:[10.1175/JCLI4124.1](https://doi.org/10.1175/JCLI4124.1).
- Lau, K.-H., and N.-C. Lau, 1992: The energetics and propagation dynamics of tropical summertime synoptic-scale disturbances. *Mon. Wea. Rev.*, **120**, 2523–2539, doi:[10.1175/1520-0493\(1992\)120<2523:TEAPDO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1992)120<2523:TEAPDO>2.0.CO;2).
- Lau, W. K.-M., H.-T. Wu, and K.-M. Kim, 2013: A canonical response of precipitation characteristics to global warming from CMIP5 models. *Geophys. Res. Lett.*, **40**, 3163–3169, doi:[10.1002/grl.50420](https://doi.org/10.1002/grl.50420).
- Leroux, S., N. M. J. Hall, and G. N. Kiladis, 2010: A climatological study of transient–mean-flow interactions over West Africa. *Quart. J. Roy. Meteor. Soc.*, **136**, 397–410, doi:[10.1002/qj.474](https://doi.org/10.1002/qj.474).
- Li, Y., P. Yang, G. R. North, and A. Dessler, 2012: Test of the fixed anvil temperature hypothesis. *J. Atmos. Sci.*, **69**, 2317–2328, doi:[10.1175/JAS-D-11-0158.1](https://doi.org/10.1175/JAS-D-11-0158.1).
- Maloney, E. D., and M. J. Dickinson, 2003: The intraseasonal oscillation and the energetics of summertime tropical western North Pacific synoptic-scale disturbances. *J. Atmos. Sci.*, **60**, 2153–2168, doi:[10.1175/1520-0469\(2003\)060<2153:TIOATE>2.0.CO;2](https://doi.org/10.1175/1520-0469(2003)060<2153:TIOATE>2.0.CO;2).
- Martin, E. R., and C. Thorncroft, 2015: Representation of African easterly waves in CMIP5 models. *J. Climate*, **28**, 7702–7715, doi:[10.1175/JCLI-D-15-0145.1](https://doi.org/10.1175/JCLI-D-15-0145.1).
- McCrary, R. R., D. A. Randall, and C. Stan, 2014a: Simulations of the West African monsoon with a superparameterized climate model. Part I: The seasonal cycle. *J. Climate*, **27**, 8303–8323, doi:[10.1175/JCLI-D-13-00676.1](https://doi.org/10.1175/JCLI-D-13-00676.1).
- , —, and —, 2014b: Simulations of the West African monsoon with a superparameterized climate model. Part II: African easterly waves. *J. Climate*, **27**, 8323–8341, doi:[10.1175/JCLI-D-13-00677.1](https://doi.org/10.1175/JCLI-D-13-00677.1).
- Mekonnen, A., C. D. Thorncroft, and A. R. Aiyer, 2006: Analysis of convection and its association with African easterly waves. *J. Climate*, **19**, 5405–5421, doi:[10.1175/JCLI3920.1](https://doi.org/10.1175/JCLI3920.1).
- Mohino, E., S. Janicot, and J. Bader, 2011: Sahel rainfall and decadal to multi-decadal sea surface temperature variability. *Climate Dyn.*, **37**, 419–440, doi:[10.1007/s00382-010-0867-2](https://doi.org/10.1007/s00382-010-0867-2).
- Nguyen, H., C. D. Thorncroft, and C. Zhang, 2011: Guinean coastal rainfall of the West African monsoon. *Quart. J. Roy. Meteor. Soc.*, **137**, 1828–1840, doi:[10.1002/qj.867](https://doi.org/10.1002/qj.867).
- Norquist, D., E. Recker, and R. Reed, 1977: Energetics of African wave disturbances as observed during phase III of GATE. *Mon. Wea. Rev.*, **105**, 334–342, doi:[10.1175/1520-0493\(1977\)105<0334:TEOAWD>2.0.CO;2](https://doi.org/10.1175/1520-0493(1977)105<0334:TEOAWD>2.0.CO;2).
- Pendergrass, A. G., and D. L. Hartmann, 2014a: The atmospheric energy constraint on global-mean precipitation change. *J. Climate*, **27**, 757–768, doi:[10.1175/JCLI-D-13-00163.1](https://doi.org/10.1175/JCLI-D-13-00163.1).
- , and —, 2014b: Changes in the distribution of rain frequency and intensity in response to global warming. *J. Climate*, **27**, 8372–8383, doi:[10.1175/JCLI-D-14-00183.1](https://doi.org/10.1175/JCLI-D-14-00183.1).
- Pritchard, M. S., and R. C. J. Somerville, 2009: Assessing the diurnal cycle of precipitation in a multi-scale climate model. *J. Adv. Model. Earth Syst.*, **1**, 12, doi:[10.3894/JAMES.2009.1.12](https://doi.org/10.3894/JAMES.2009.1.12).
- Pu, B., and K. H. Cook, 2012: Role of the West African westerly jet in Sahel rainfall variations. *J. Climate*, **25**, 2880–2896, doi:[10.1175/JCLI-D-11-00394.1](https://doi.org/10.1175/JCLI-D-11-00394.1).

- Ramel, R., H. Gallée, and C. Messenger, 2006: On the northward shift of the West African monsoon. *Climate Dyn.*, **26**, 429–440, doi:[10.1007/s00382-005-0093-5](https://doi.org/10.1007/s00382-005-0093-5).
- Reed, R., D. Norquist, and E. Recker, 1977: The structure and properties of African wave disturbances as observed during phase III of GATE. *Mon. Wea. Rev.*, **105**, 317–333, doi:[10.1175/1520-0493\(1977\)105<0317:TSAPOA>2.0.CO;2](https://doi.org/10.1175/1520-0493(1977)105<0317:TSAPOA>2.0.CO;2).
- Roehrig, R., D. Bouniol, F. Guichard, F. Hourdin, and J. Redelsperger, 2013: The present and future of the West African monsoon: A process-oriented assessment of CMIP5 simulations along the AMMA transect. *J. Climate*, **26**, 6471–6505, doi:[10.1175/JCLI-D-12-00505.1](https://doi.org/10.1175/JCLI-D-12-00505.1).
- Romps, D. M., 2011: Response of tropical precipitation to global warming. *J. Atmos. Sci.*, **68**, 123–138, doi:[10.1175/2010JAS3542.1](https://doi.org/10.1175/2010JAS3542.1).
- , 2012: Weak pressure gradient approximation and its analytical solutions. *J. Atmos. Sci.*, **69**, 2835–2845, doi:[10.1175/JAS-D-11-0336.1](https://doi.org/10.1175/JAS-D-11-0336.1).
- Russell, J. O., A. Aiyer, J. D. White, and W. Hannah, 2016: Revisiting the connection between African easterly waves and Atlantic tropical cyclogenesis. *Geophys. Res. Lett.*, **44**, 587–595, doi:[10.1002/2016GL071236](https://doi.org/10.1002/2016GL071236).
- Ruti, P. M., and A. Dell'Aquila, 2010: The twentieth century African easterly waves in reanalysis systems and IPCC simulations, from intra-seasonal to inter-annual variability. *Climate Dyn.*, **35**, 1099–1117, doi:[10.1007/s00382-010-0894-z](https://doi.org/10.1007/s00382-010-0894-z).
- Schneider, T., T. Bischoff, and G. H. Haug, 2014: Migrations and dynamics of the intertropical convergence zone. *Nature*, **513**, 45–53, doi:[10.1038/nature13636](https://doi.org/10.1038/nature13636).
- Sherwood, S. C., C. L. Meyer, R. J. Allen, and H. A. Titchner, 2008: Robust tropospheric warming revealed by iteratively homogenized radiosonde data. *J. Climate*, **21**, 5336–5352, doi:[10.1175/2008JCLI2320.1](https://doi.org/10.1175/2008JCLI2320.1).
- Skinner, C. B., and N. S. Diffenbaugh, 2013: The contribution of African easterly waves to monsoon precipitation in the CMIP3 ensemble. *J. Geophys. Res. Atmos.*, **118**, 3590–3609, doi:[10.1002/jgrd.50363](https://doi.org/10.1002/jgrd.50363).
- , and —, 2014: Projected changes in African easterly wave intensity and track in response to greenhouse forcing. *Proc. Natl. Acad. Sci. USA*, **111**, 6882–6887, doi:[10.1073/pnas.1319597111](https://doi.org/10.1073/pnas.1319597111).
- Sobel, A. H., and C. S. Bretherton, 2000: Modeling tropical precipitation in a single column. *J. Climate*, **13**, 4378–4392, doi:[10.1175/1520-0442\(2000\)013<4378:MTPIAS>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<4378:MTPIAS>2.0.CO;2).
- , J. Nilsson, and L. M. Polvani, 2001: The weak temperature gradient approximation and balanced tropical moisture waves. *J. Atmos. Sci.*, **58**, 3650–3665, doi:[10.1175/1520-0469\(2001\)058<3650:TWTGAA>2.0.CO;2](https://doi.org/10.1175/1520-0469(2001)058<3650:TWTGAA>2.0.CO;2).
- Soden, B. J., and I. M. Held, 2006: An assessment of climate feedbacks in coupled ocean atmosphere models. *J. Climate*, **19**, 3354–3360, doi:[10.1175/JCLI3799.1](https://doi.org/10.1175/JCLI3799.1).
- Stan, C., M. Khairoutdinov, C. A. DeMott, V. Krishnamurthy, D. M. Straus, D. A. Randall, J. L. Kinter, and J. Shukla, 2010: An ocean–atmosphere climate simulation with an embedded cloud resolving model. *Geophys. Res. Lett.*, **37**, L01702, doi:[10.1029/2009GL040822](https://doi.org/10.1029/2009GL040822).
- Straub, K. H., P. T. Haertel, and G. N. Kiladis, 2010: An analysis of convectively coupled Kelvin waves in 20 WCRP CMIP3 global coupled climate models. *J. Climate*, **23**, 3031–3056, doi:[10.1175/2009JCLI3422.1](https://doi.org/10.1175/2009JCLI3422.1).
- Sultan, B., and S. Janicot, 2000: Abrupt shift of the ITCZ over West Africa and intra-seasonal variability. *Geophys. Res. Lett.*, **27**, 3353–3356, doi:[10.1029/1999GL011285](https://doi.org/10.1029/1999GL011285).
- Sun, Y., S. Solomon, A. Dai, and R. W. Portmann, 2006: How often does it rain? *J. Climate*, **19**, 916–934, doi:[10.1175/JCLI3672.1](https://doi.org/10.1175/JCLI3672.1).
- , —, —, and —, 2007: How often will it rain? *J. Climate*, **20**, 4801–4818, doi:[10.1175/JCLI4263.1](https://doi.org/10.1175/JCLI4263.1).
- Thorncroft, C. D., and M. Blackburn, 1999: Maintenance of the African easterly jet. *Quart. J. Roy. Meteor. Soc.*, **125**, 763–786, doi:[10.1002/qj.49712555502](https://doi.org/10.1002/qj.49712555502).
- , and K. Hodges, 2001: African easterly wave variability and its relationship to Atlantic tropical cyclone activity. *J. Climate*, **14**, 1166–1179, doi:[10.1175/1520-0442\(2001\)014<1166:AEWVAI>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<1166:AEWVAI>2.0.CO;2).
- , N. M. J. Hall, and G. N. Kiladis, 2008: Three-dimensional structure and dynamics of African easterly waves. Part III: Genesis. *J. Atmos. Sci.*, **65**, 3596–3607, doi:[10.1175/2008JAS2575.1](https://doi.org/10.1175/2008JAS2575.1).
- , H. Nguyen, C. Zhang, and P. Peyrillé, 2011: Annual cycle of the West African monsoon: Regional circulations and associated water vapour transport. *Quart. J. Roy. Meteor. Soc.*, **137**, 129–147, doi:[10.1002/qj.728](https://doi.org/10.1002/qj.728).
- Trenberth, K. E., 1999: Conceptual framework for changes of extremes of the hydrological cycle with climate change. *Climatic Change*, **42**, 327–339, doi:[10.1023/A:1005488920935](https://doi.org/10.1023/A:1005488920935).
- Tulich, S. N., 2015: A strategy for representing the effects of convective momentum transport in multiscale models: Evaluation using a new superparameterized version of the Weather Research and Forecast model (SP-WRF). *J. Adv. Model. Earth Syst.*, **7**, 938–962, doi:[10.1002/2014MS000417](https://doi.org/10.1002/2014MS000417).
- Vizy, E. K., and K. H. Cook, 2002: Development and application of a mesoscale climate model for the tropics: Influence of sea surface temperature anomalies on the West African monsoon. *J. Geophys. Res.*, **107**, 4023, doi:[10.1029/2001JD000686](https://doi.org/10.1029/2001JD000686).
- Wolding, B. O., E. D. Maloney, and M. Branson, 2016: Vertically resolved weak temperature gradient analysis of the Madden–Julian Oscillation in SP-CESM. *J. Adv. Model. Earth Syst.*, **8**, 1586–1619, doi:[10.1002/2016MS000724](https://doi.org/10.1002/2016MS000724).
- Yanai, M., S. Esbensen, and J. Chu, 1973: Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. *J. Atmos. Sci.*, **30**, 611–627, doi:[10.1175/1520-0469\(1973\)030<0611:DOBPOT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1973)030<0611:DOBPOT>2.0.CO;2).
- Zelinka, M. D., and D. L. Hartmann, 2010: Why is longwave cloud feedback positive? *J. Geophys. Res.*, **115**, D16117, doi:[10.1029/2010JD013817](https://doi.org/10.1029/2010JD013817).
- Zhang, G. J., and N. A. McFarlane, 1995: Sensitivity of climate simulations to the parameterization of cumulus convection in the Canadian Climate Centre General Circulation Model. *Atmos.–Ocean*, **33**, 407–446, doi:[10.1080/07055900.1995.9649539](https://doi.org/10.1080/07055900.1995.9649539).