Simulations of the West African Monsoon with a Superparameterized Climate Model. Part II: African Easterly Waves

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ABSTRACT

The relationship between African easterly waves and convection is examined in two coupled general circulation models: the Community Climate System Model (CCSM) and the “superparameterized” CCSM (SP-CCSM). In the CCSM, the easterly waves are much weaker than observed. In the SP-CCSM, a two-dimensional cloud-resolving model replaces the conventional cloud parameterizations of CCSM. Results show that this allows for the simulation of easterly waves with realistic horizontal and vertical structures, although the model exaggerates the intensity of easterly wave activity over West Africa. The simulated waves of SP-CCSM are generated in East Africa and propagate westward at similar (although slightly slower) phase speeds to observations. The vertical structure of the waves resembles the first baroclinic mode. The coupling of the waves with convection is realistic. Evidence is provided herein that the diabatic heating associated with deep convection provides energy to the waves simulated in SP-CCSM. In contrast, horizontal and vertical structures of the weak waves in CCSM are unrealistic, and the simulated convection is decoupled from the circulation.

1. Introduction

The West African monsoon (WAM) involves many complicated interactions between the atmosphere, ocean, and land surface on a wide range of temporal and spatial scales, from individual rain events to global atmospheric dynamics. Coupled global circulation models (CGCMs), which are used to project future climate changes, have difficulty representing the mean annual cycle and variability of precipitation over West Africa (e.g., Cook and Vizy 2006; Roehrig et al. 2013). This problem is often attributed to sea surface temperature (SST) biases in the equatorial Atlantic and their influence on the intertropical convergence zone (e.g., Vizy and Cook 2002). Even when forced with observed SSTs, however, the models struggle to capture key rain-making weather systems such as African easterly waves (AEWs) (Sander and Jones 2008; Ruti and Dell’Aquila 2010).

AEWs are synoptic-scale disturbances with periods of approximately 3–6 days, wavelengths in the range of 2000–6000 km, and westward phase speeds of 7–9 m s\(^{-1}\) (e.g., Burpee 1972, 1974; Reed et al. 1977). AEWs are the dominant mode of variability over West Africa during the summer monsoon season [June–September (JJAS)] and are strongly linked to rainfall and convection (Reed et al. 1977; Duvel 1990: Mathon et al. 2002; Fink and Reiner 2003; Gu et al. 2004; Kiladis et al. 2006; Mekonnen...
et al. 2006). It is well established that AEWs act as seed disturbances for Atlantic hurricanes (Carlson 1969; Duvel 1990; Avila and Pasch 1992; Thorncroft and Hodges 2001; Hopsch et al. 2010).

Despite decades of research, a complete understanding of the initiation, development, and maintenance of African easterly waves continues to elude us. In the past, AEWs were thought to develop solely due to the combined barotropic/baroclinic instability of the midtropospheric African easterly jet (AEJ) (e.g., Burpee 1972; Thorncroft and Hoskins 1994a,b). Current theory suggests, however, that AEW initiation and growth cannot be accounted for without an additional energy source (Hall et al. 2006; Thorncroft et al. 2008; Hsieh and Cook 2005, 2007, 2008). In particular, it has been argued that diabatic heating associated with convection is needed to account for the dynamics of AEWs. Both theory (Thorncroft et al. 2008) and observational evidence (Berry and Thorncroft 2005; Mekonnen et al. 2006; Kiladis et al. 2006) suggest that convective heating in the vicinity of the Darfur mountains acts as a finite-amplitude perturbation to the regional circulation, triggering AEWs. The waves then propagate westward, drawing additional energy from the barotropic/baroclinic instability of the AEJ. It has also been suggested that AEWs can be initiated by the intense convection associated with the ITCZ (Hsieh and Cook 2005). Convective heating within the ITCZ creates potential vorticity (Schubert et al. 1991) and the reversal of the meridional potential vorticity gradient triggers the Charney–Stern relationship, which provides energy to the waves (Hsieh and Cook 2005; Berry and Thorncroft 2012). Berry and Thorncroft (2012) showed that the organized deep convection embedded within the AEWs themselves is critically important for the overall energetics of the waves and that without the convective heat source AEWs cannot persist over Africa for as long as is observed.

A number of studies have examined AEW activity in CGCMs (Ruti and Dell’Aquila 2010; Daloz et al. 2012; Skinner and Diffenbaugh 2013), uncoupled GCMs (Fyfe 1999; Chauvin et al. 2005; Reale et al. 2007; Pohl and Douville 2011), and regional climate models (Seo et al. 2008; Sylla et al. 2013). Both Ruti and Dell’Aquila (2010) and Skinner and Diffenbaugh (2013) demonstrate that the CGCMs used in phase 3 of the Coupled Model Intercomparison Project (CMIP3) exhibit disparate wave activity, with some showing overly robust wave activity and others showing little to no wave activity. Skinner and Diffenbaugh (2013) go on to highlight that synoptic-scale precipitation variability is strongly correlated with AEW activity. Uncoupled GCM and regional climate model experiments suggest that increased horizontal resolution may improve the simulation of AEWs (e.g., Seo et al. 2008). One limitation of these studies is that they have focused primarily on the kinematic properties of AEWs, neglecting the role that convection and diabatic heating play in AEW development and maintenance. As discussed above, it is now believed that convection is an important component of AEW dynamics and must be considered in model evaluations.

Our goal in this study is to examine and compare the role that convection plays in the simulation of AEWs in two CGCMs. The first is the standard Community Climate System Model version 3 (hereafter CCSM3) (Collins et al. 2006) and the second is the superparameterized CCSM3 (SP-CCSM) (Stan et al. 2010). With superparameterization, conventional cloud parameterizations are replaced by a simplified cloud-resolving model (CRM) embedded in each grid column (e.g., Grabowski 2001; Khairoutdinov and Randall 2001). The superparameterization does not increase the resolution of the host model; it is simply a more physically based way to parameterize heating and drying rates. Superparameterized GCMS can serve as a stepping stone between traditional GCMS and the global CRMs, which cannot yet be used in climate studies. With superparameterization, we can study the coupling between the large-scale circulation and cloud systems without the computational cost of global CRMs.

As shown in Part I of this paper (McCrary et al. 2014, hereafter Part I), the implementation of the superparameterization improves many aspects of the West African monsoon (WAM) simulated by CCSM3, including the mean position and intensity of monsoon rainfall and the meridional displacement of the ITCZ during the annual cycle. Part I shows that many of the changes to the simulated WAM system can be attributed to the improved vertical structure of atmospheric heating and moistening simulated by the SP-CCSM. By examining the 2–6-day bandpass-filtered eddy kinetic energy (EKE), Part I also shows that the SP-CCSM exaggerates the intensity of easterly wave activity over Africa, while CCSM3 exhibits little to no wave activity in the same region (see Fig. 12 in Part I). The difference between these models is not their resolution, but in the way subgrid-scale cloud processes are represented.

The purpose of this paper is to identify why the two models exhibit such different wave activity. Our emphasis is on trying to understand how changing the physics influences AEW structure. We do this by exploring the relationship between AEWs and convection in both CCSM3 and SP-CCSM. The paper is organized as follows. Section 2 discusses the models, data sources, and methodology used to examine AEWs. Section 3 presents a brief overview of the aspects of the climatology
over West Africa that are important for AEWs. The variability of the simulated AEW activity is compared to observations in section 4. Section 5 discusses the horizontal and vertical structures of the simulated waves. The energetics of the simulated AEWs is explored in section 6. The paper ends in section 7, with a summary of the results and a discussion.

2. Data and methods

a. Models, observations, and reanalysis

A detailed description of the models used in this study can be found in Part I, Stan et al. (2010), and DeMott et al. (2011, 2013). The experimental designs are summarized in Table 1. Briefly, the atmospheric models were run at T42 resolution with a semi-Lagrangian dynamical core. Each simulation is 25 years in length. The CCSM3 simulation used 26 levels, while the SP-CCSM had 30 levels. In CCSM3, deep convection is parameterized following Zhang and McFarlane (1995). In SP-CCSM, 2D CRMs with periodic lateral boundary conditions replace the conventional parameterization of subgrid-scale cloud processes. The embedded CRMs have a horizontal spacing of 4 km, 28 levels, and are oriented in the east–west direction. Since the CRMs are 2D, convective momentum feedback from the CRM to the large scale is not included. In both simulations the atmospheric model is coupled to the 3rd version of the Parallel Ocean Program (POP) ocean model (Smith and Gent 2002) and the Community Land Model version 3 (CLM3, Bonan et al. 2002). The cloud-resolving models are coupled to the land surface at the GCM scale. More information about the superparameterization can be found in Grabowski and Smolarkiewicz (1999), Grabowski (2001), Khairoutdinov and Randall (2001, 2003), and Khairoutdinov et al. (2005, 2008). The ocean coupling is described by Stan et al. (2010). Randall et al. (2014) provide an overview of the superparameterized framework highlighting the influence of superparameterization on tropical variability.

Table 1 also summarizes the observation-based datasets used in this study. Briefly, precipitation is from the 3-hourly Tropical Rainfall Measure Mission (TRMM) 3B42 precipitation dataset (Huffman et al. 2007). Outgoing longwave radiation (OLR) is from the daily mean National Oceanic and Atmospheric Administration interpolated OLR dataset (Liebmann and Smith 1996). All other meteorological fields are from the Interim European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-Interim, herein ERA-I) (Dee et al. 2011). All observational data have been averaged to daily time scales. The apparent heat source, $Q_1$, is estimated from ERA-I and both models using the approach of Lin and Johnson (1996): $Q_1$ is a measure of diabatic heating and includes the influence of both radiative heating and latent heating. Although ERA-I is anchored to observations through data assimilation, $Q_1$ is also influenced by precipitation in ERA-I, which is

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Origin/reference</th>
<th>Horizontal resolution</th>
<th>Temporal resolution</th>
<th>Vertical levels</th>
<th>Selected variables</th>
</tr>
</thead>
<tbody>
<tr>
<td>CCSM3 Model</td>
<td>(Collins et al. 2006)</td>
<td>T42 2.8° × 2.8°</td>
<td>Daily mean</td>
<td>26 levels interpolated to standard pressure levels</td>
<td>Precipitation, OLR, winds, temperature, specific humidity, geopotential height</td>
</tr>
<tr>
<td>SP-CCSM Model</td>
<td>(Stan et al. 2010)</td>
<td>T42 2.8° × 2.8° CRMs; 4 km in north–south direction</td>
<td>Daily mean</td>
<td>30 levels interpolated to standard pressure levels, CRM levels collocated with GCM levels</td>
<td>Precipitation, OLR, winds, temperature, specific humidity, geopotential height</td>
</tr>
<tr>
<td>TRMM (3B42)</td>
<td>Satellite and rain gauge (Huffman et al. 2007)</td>
<td>0.25° × 0.25°</td>
<td>3 hourly averaged to daily means from 1998–2010</td>
<td>Surface</td>
<td>Precipitation</td>
</tr>
<tr>
<td>NOAA OLR</td>
<td>Satellite (Liebmann and Smith 1996)</td>
<td>2.5° × 2.5°</td>
<td>Daily mean from 1979–2010</td>
<td>Top of atmosphere</td>
<td>OLR</td>
</tr>
<tr>
<td>ERA-Interim</td>
<td>Radiosonde, satellite, model forecast (Dee et al. 2011)</td>
<td>1.5° × 1.5°</td>
<td>4 times daily averaged to daily means from 1979–2010</td>
<td>25 levels 1000–100 hPa</td>
<td>Winds, specific humidity, temperature, geopotential height</td>
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3. Monsoon climatology

a. Precipitation, OLR, and low-level winds

The observed seasonal cycle of the WAM has been well documented in previous studies (e.g., Liebmann et al. 2012). Part I gives a detailed discussion and examination of the seasonal cycle as simulated by both CCSM3 and SP-CCSM. Here we show the key features of the summer monsoon that are related to AEW activity.

Figures 1a–c show the JJAS climatological precipitation, 925-hPa winds, and the monsoon air mass from observations and the models. The simulated climatology is based on the entire 25-yr record from each model, in observations the climatology of TRMM is from 1997–2010, and ERA-I is derived from 1979–2010. In the observed monsoon, JJAS represents the peak monsoon season when the largest rainfall amounts of the year occur over West Africa. During this season, the ITCZ has moved northward, shifting from over the ocean during boreal winter to over the continent during JJAS. Embedded within the observed monsoon are three distinct precipitation maxima: one over the Ethiopian Highlands, one near the Bight of Bonny near Cameroon, and a third near the Guinea Highlands that extends out over the Atlantic Ocean. These precipitation maxima are all colocated with regional topography. The southwesterly low-level monsoon winds cross the equator, and bring moisture onto the continent. The monsoon winds converge with the dry northeasterly harmattan winds in what is referred to as the intertropical discontinuity (indicated by the thick back line in Fig. 1), just to the north of the 1 mm day$^{-1}$ precipitation contour. Monsoon rainfall is qualitatively similar in other precipitation products such as those of the Climatic Research Unit (CRU) and Global Precipitation Climatology Project (GPCP).

When compared to CCSM3, SP-CCSM better represents both the magnitude and the spatial patterns of summer monsoon precipitation over West Africa (Fig. 1). The region of maximum precipitation is shifted from over the Gulf of Guinea in CCSM3 (not realistic) to over the continent in SP-CCSM. SP-CCSM captures the local maximum in precipitation over the Atlantic as well as the maximum over the Ethiopian Highlands. SP-CCSM does not simulate the maximum in precipitation near Cameroon nor does it capture the dry region along the Guinea Coast between the two coastal precipitation maxima. Rain rates in the SP-CCSM trend to be much larger than observed and rainfall during the peak monsoon period does not penetrate as far northward as observed. In CCSM3, rain rates over the continent are weaker than observed, and the simulated precipitation maximum occurs over the Gulf of Guinea. It is likely that the rainfall biases along the Guinea Coast in SP-CCSM and over the Gulf of Guinea in CCSM3 are linked to warm SST biases in both models and a misrepresentation of the Atlantic equatorial cold tongue (see Part I). Monsoon winds do not penetrate as far northward as observed in either model, and the intertropical discontinuity is positioned farther south than observed in both models. These differences have important implications for the energetics analysis presented in section 6.
OLR is often used as a proxy for rainfall because it is more normally distributed than precipitation and the time record for daily OLR covers a much longer period (e.g., ~30 yr for the NOAA OLR dataset compared to 13 yr for TRMM precipitation). We use OLR as a proxy for rainfall in this study, but with some misgivings, as discussed below.

Figures 1d–f show that the JJAS OLR in the observations and both models corresponds well with regions of enhanced rainfall. The observed OLR climatology is based on the 1979–2010 time period; the simulated climatology is from the 25-yr simulation. In the models, the OLR is influenced not only by the cloud parameterizations, which generate the simulated clouds, but also by the microphysics parameterizations and the way that radiation is parameterized. For example, the colder cloud tops found in CCSM3, compared to SP-CCSM, might seem to suggest that more deep convection forms over West Africa in CCSM3. In fact, however, the cold cloud tops in CCSM3 are not associated with much rainfall.

b. The African easterly jet

As discussed in the introduction, the AEJ is important for AEW dynamics. While instabilities associated with the AEJ are not sufficient to trigger AEWs (Hall et al. 2006), the jet is believed to be important for the propagation and overall development of AEWs. The jet develops during the summer season owing to the strong low-level meridional temperature and soil moisture gradients between the Gulf of Guinea and the Sahara. Figures 1g–i show the JJAS mean zonal wind at 600 hPa. In ERA-I, the peak winds reach about 12–14 m s$^{-1}$ (Fig. 1h). The winds are found at about 15°N, at the
600-hPa level, and extend from the west coast to approximately 5°E. Easterly winds also extend down to about 850 hPa, where they give way to the westerly monsoon winds (not shown). The AEJ is reasonably well positioned in both CCSM3 and SP-CCSM. The jet tends to be slightly weaker than observed in SP-CCSM and extends farther east than observed, to about 20°E. As mentioned in Part I, the AEJ may be weaker than observed in SP-CCSM because of the overly active simulated easterly wave activity. The strength of the AEJ may also be influenced by the differences in the low-level meridional temperature gradients found in each model as well as differences in the representation of the diabatically forced circulations associated with the ITCZ and the Saharan heat low (Part I).

4. African easterly waves

To examine the organization of tropical convection during boreal summer, space–time spectra of the OLR for JJAS have been calculated for the observations and both models. Figure 2 shows the average June–September signal-to-noise ratio power spectra of OLR averaged between 15°S and 15°N for disturbances that are symmetric (top) and antisymmetric across the equator (bottom). These figures were created using the methods outlined in Wheeler and Kiladis (1999). Anomalies are generated by removing the first three harmonics of the seasonal cycle, the June–September signal is calculated by subdividing each summer into two 96-day segments that overlap by 60 days, and the red-noise spectrum is based on the data from all months of the year.

In the observations, considerable power is found in the tropical depression (TD) range, which includes westward propagating waves with wavelengths 2000–7000 km and periods 2–10 days. The TD range has been shown to correspond well with AEW activity (K06). CCSM3 exhibits little to no power in ranges associated with the Madden–Julian oscillation (MJO), equatorial Rossby (ER) waves, or inertia–gravity waves. It does have substantial power in the Kelvin wave range, although at shorter wavelengths than observed. There is no coherent easterly wave power in CCSM3.

Implementation of the superparameterization alters the interactions between the large-scale dynamics and cloud processes in such a way that the simulated spectrum of tropical wave activity is more realistic. In particular,
the SP-CCSM simulates TD activity, suggesting that AEW activity should be present in this model. Easterly waves are predominantly a Northern Hemisphere phenomenon, so significant power is also found in the TD region of the antisymmetric power spectra (Figs. 2d–f). Figure 3 shows the variance of TD filtered OLR from observations and for both models. During JJAS, the observed easterly wave activity occurs broadly across most of the tropics and is strongest over West Africa, the eastern Atlantic, and the intra-American seas (Fig. 3a). Easterly waves are also observed over the central and western Pacific. Over West Africa, TD filtering captures approximately 20%–30% of the total observed OLR variance (McCrary 2012).

In SP-CCSM, the simulated easterly wave activity is greatly exaggerated over West Africa and the east Atlantic, where the variance in TD-filtered OLR is more than 4 times larger than observed (Fig. 3b). In this region, the simulated TD-filtered OLR contains 35%–45% of the total variability (McCrary 2012). SP-CCSM also simulates a southward shift in the TD-filtered OLR over Africa, compared to observations. This corresponds to the regional maximum of precipitation in SP-CCSM (Figs. 1b,e). As in the observations, easterly wave activity occurs broadly over the tropics in SP-CCSM, but it is weaker than observed over the central and western Pacific. We speculate that the exaggerated easterly wave activity over Africa may be due in part to differences in the simulated convective processes over land compared to oceans.

As expected from the space–time spectra, easterly wave activity in CCSM3 is weaker than observed (Fig. 3c). Over Africa, the only signal in the CCSM3-simulated TD-filtered variance is found over the Gulf of Guinea, where precipitation is a maximum. This variance is less than half of the observed and accounts for less than 15% of the total variance (McCrary 2012). We obtained similar results using traditional kinematic indices of AEW activity, such as the variance in 2–6-day bandpass-filtered 700-hPa meridional wind (see Part I). Again, SP-CCSM shows exaggerated AEW activity, while CCSM3 exhibits much less activity than is observed (see Part I).

5. The structure of African easterly waves

Next, we compare the horizontal and vertical structures of the simulated AEWs with observations during JJAS. To do this we use lagged linear regression techniques similar to those found in K06 and references within. These techniques are often used to study convectively coupled waves and allow us to compare the composite horizontal and vertical structure of the observed and simulated waves. Following K06, we use the TD-filtered time series of OLR at the point 10°N, 10°W (our base point) as an index of AEW activity. This base point time series of AEW variability serves as the reference time series against which all dynamic, thermodynamic, and convective variables are regressed. The TD-filtered AEW index is regressed against unfiltered daily anomalies of all circulation- and convection-related variables of interest at all grid points and levels covering the West African domain. These variables include OLR, streamfunction, winds, vertical velocity, temperature, humidity, and diabatic heating. In the following figures, the regressed value of each variable is plotted at each grid point and level for a −1 standard deviation change in the filtered AEW index. This allows us to examine the circulation associated with a typical AEW disturbance passing over the point 10°N, 10°W. Lag regressions are also calculated to examine the circulation on the days before and after the wave passes the base point. Regressions are calculated for the JJAS time period using data from all years available from each dataset and model.

The base point 10°N, 10°W was chosen because it was used in K06 and is in a region of high OLR variability. It is south of the AEJ, where barotropic energy conversions occur. Other base points have also been examined, and results are discussed by McCrary (2012).
a. Horizontal structure

The horizontal structure of AEWs is shown in Fig. 4. This figure shows the OLR and circulation (streamfunction and winds) anomalies corresponding with a $-1$ standard deviation change in the TD-filtered OLR time series at the base point $10^\circ$N, $10^\circ$W from observations (left), SP-CCSM (middle) and CCSM3 (left). With this we can examine the linear relationship between AEW activity at the base point and the corresponding spatial patterns of the circulation and convection over all of West Africa. To examine the temporal progression of the waves as they pass the base point, the regressions are calculated for lag $-2$ (top), 0 (middle), and 2 (bottom) days. For OLR and wind anomalies only values where correlations are determined to be statistically significant at the 95% confidence level using a two-tailed $t$ test are shown. The streamfunction is shown everywhere for ease of interpretation of the overall circulation. For each model, the contours are scaled by the standard deviation of the TD-filtered time series at the base point, in order to emphasize the differences in the structures of the waves rather than their magnitudes. In all cases, the magnitudes are larger than observed in SP-CCSM and smaller than observed in CCSM3.

Despite large differences in the strength of the AEWs in the observations compared to SP-CCSM, the horizontal structure of the simulated waves is fairly realistic. At lag 0 (Figs. 4b,e), there is a relatively large area of negative OLR centered near the base point, indicating...
a region of enhanced convection. This convective region is flanked on either side by regions of resolved-scale subsidence. In both cases, the convective anomalies are collocated with anomalies of the 850-hPa circulation. Enhanced convection is associated with an anomalous cyclonic circulation, and suppressed convection is associated with an anomalous anticyclonic circulation. At lag 0, convection occurs within the trough of the wave in both cases. Following the development of these waves at various lags, we see that they emanate from East Africa (Figs. 4a,d) and propagate offshore into the Atlantic basin (Figs. 4c,f), where it is observed that they sometimes act as seed disturbances for Atlantic hurricanes.

Some notable differences between the observed and simulated waves are as follows. OLR anomalies in SP-CCSM are shifted to the south of the center of each circulation pattern, which is not observed. The observed waves originate over the Darfur Mountains or as far east as the Ethiopian Highlands (K06), but in SP-CCSM the waves form farther west. As noted by K06, as the observed waves propagate westward, the OLR minimum shifts from behind the trough over East Africa, to centered within the trough over West Africa, and finally to ahead of the trough over the Atlantic. A similar analysis done with TRMM rainfall data rather than OLR shows that actual precipitation may occur ahead of the trough as AEWs propagate across West Africa (Fig. 5). In Fig. 5
the horizontal structures of the waves from observations and both models are calculated using the TD-filtered time series of precipitation, rather than OLR, at the base point. The TRMM data show that, as might be expected in midlatitude synoptic storms, precipitation tends to occur ahead of the trough axis as the waves propagate across West Africa (Figs. 5a–c). The differences between precipitation and OLR may be due to either persistence of high clouds after precipitation has ended or the coarse resolution of the NOAA OLR product. The spatial relationship between convection and the wave in SP-CCSM is different from observations, which show that convection (both OLR and precipitation) resides near the trough axis as the wave propagates across Africa (Figs. 4d,e and 5d,e) and falls behind the trough when the wave moves out over the Atlantic (Figs. 4f and 5f). Differences between the simulated and observed waves are likely due to either the coarse resolution of the global model, which spreads rainfall out spatially, or incorrect timing of convection relative to the wave. The waves simulated by SP-CCSM also propagate more slowly than observed, which may be due to the overly strong convection found in the model.

In CCSM3, the regression method does show a weak convective signal along 10°N, but there is no statistically significant circulation pattern associated with the OLR or precipitation anomalies (Figs. 4g–i and 5g–i). This indicates that convection is decoupled from the circulation on the TD time and space scales.

As mentioned in the introduction, a number of previous studies have shown that observed AEWs tend to follow two different tracks, one to the south of the AEJ and another to the north. In this section we have focused on the southern track of the observed AEWs, as the direct link with moist convection is clearer south of the AEJ. When we examined AEWs using base points to the north of the AEJ, we found that the northern track is not well captured by the SP-CCSM. Further discussion is given by McCrery (2012). As shown in the next section, diabatic heating appears to drive the waves simulated by SP-CCSM. The northern track of observed AEWs tends to be dry and not as closely associated with convection as the southern track (Reed et al. 1977). In SP-CCSM, any wave activity to the north of the AEJ appears to emanate out of the region of maximum convection, which is not observed (McCrery 2012).

b. Vertical structure

The vertical structures of the observed and simulated AEWs are depicted in Fig. 6, which shows vertical cross sections, along 10°N, of anomalies of meridional wind (Figs. 6d–f), omega (Figs. 6g–i), temperature (Figs. 6j–l), specific humidity (Q, Figs. 6m–o), and the apparent heat source (Q1, Figs. 6p–r) that correspond with a ± 1 standard deviation change in the TD-filtered OLR time series at the base point 10°N, 10°W at lag 0 for the observations and both models. The thick black line in each panel highlights the position of peak convection, which can also be found in the OLR lag-regression plots at the top of the panel (Figs. 6a–c). As usual, we expect that the strongest convective activity roughly corresponds with a minimum in OLR.

The vertical structure of the observed AEWs is that expected for the first baroclinic mode, with meridional winds changing sign at approximately 300 hPa (Fig. 6d). There is very little tilt of the meridional winds with height in the lower levels, indicating that barotropic energy conversion is important for AEWs in this region. The peak in convection occurs near the trough axis, where the winds change from southerly (positive) on the east side of the trough to northerly (negative) on the west side. As expected, the region of peak convection is associated with the strongest upward motion (Fig. 6g).

As pointed out by K06, there are two distinct maxima in the vertical profile of omega, at 700 and 400 hPa. These are thought to correspond with two different cloud populations, namely shallow convection at low levels and deep convection aloft. Vertical profiles of specific humidity (Fig. 6m) and diabatic heating Q1 (Fig. 6p) show peaks similar to the two peaks in omega. The relative minima in omega, specific humidity, and Q1 found at 600 hPa are near the freezing level.

Ahead of the peak in convective activity, the lower troposphere is anomalously warm and moist (Figs. 6j,m), most likely due to the effects of shallow cumulus convection ahead of the peak in deep convection (K06), although warm advection from northerlies ahead of the trough may influence temperatures as well. Where convective activity is strongest, the layer between 850 and 500 hPa is cool and moist. These cool, moist conditions also trail the peak in convection by about 10° longitude. The anomalously cool temperatures behind the peak in convection may be due to adiabatic cooling associated with the peak in upward motion found in the same layer, or they may be due to the evaporation of falling rain. A second cool region is found near the surface. This low-level cool layer is likely due to convective downdrafts and cold air advection from the Gulf of Guinea. The secondary peak in vertical motion found in the upper troposphere between 200 and 400 hPa closely corresponds with anomalously warm temperatures. This warming is thought to be due to the diabatic heating associated with deep convection. The positive correlation between temperature and Q1 at the same level suggests that heating acts to strengthen the disturbance. Behind the main convective signal, at upper
Fig. 6. Zonal cross section of (a)–(c) OLR (W m\(^{-2}\)) and zonal–height cross sections of (d)–(f) anomalous meridional wind (m s\(^{-1}\)), (g)–(i) omega (Pa s\(^{-1}\)), (j)–(l) temperature (K), (m)–(o) specific humidity (g kg\(^{-1}\)), and (p)–(r) the apparent heat source \(Q_1\) (K day\(^{-1}\)) along 10\(^\circ\)N from (left) observations, (middle) SP-CCSM, and (right) CCSM at lag day 0. The position of peak convection is highlighted with the thick black line.
levels, anomalously moist conditions persist, likely associated with trailing stratiform convection. This picture of AEWs closely follows what we might expect from any convectively coupled equatorial wave (e.g., Kiladis et al. 2009).

The vertical structure of AEWs in SP-CCSM is markedly similar to that observed. The meridional winds have a “first baroclinic mode” structure, changing sign at approximately 300 hPa (Fig. 6e). At low levels, the slight eastward tilt of the meridional winds suggests that baroclinic energy conversions may be important for AEWs near 10°N, more in the model than in the observations (this will be examined further in section 6). As in ERA-I, the region of maximum convection is associated with enhanced upward motion (Fig. 6h), although it is an order of magnitude stronger in the model than in the reanalysis. Also, there is only one maximum in the omega profile, centered at about 400 hPa. The vertical profiles of convection associated with AEWs in this model do not show the observed relative minima at 600 hPa. The vertical profiles of omega (Fig. 6h), specific humidity (Fig. 6k), and diabatic heating (Fig. 6n) also show one vertical peak. It may be that the vertical resolution of the model is too coarse to capture the finescale vertical structures associated with deep convection in the tropics. The simulated vertical structure of the temperature (Fig. 6k) is comparable to observations, with a warm signal ahead of the peak in convection, cool anomalies in the mid to lower troposphere trailing the peak in convection, and warming between 200 and 400 hPa. There is also a weak warming and moistening ahead of the deep convection, which may be associated with shallow convection. The diabatic heating associated with the waves in SP-CCSM is a factor of 4 stronger than calculated from ERA-I. As discussed in the next section, diabatic heating appears to play a critical and perhaps exaggerated role in generating AEWs in SP-CCSM.

In CCSM3, the simulated vertical profiles of the meridional wind (Fig. 6f), omega (Fig. 6i), temperature (Fig. 6l), humidity (Fig. 6o), and Q1 (Fig. 6r) are quite different from those observed. Convection in CCSM3 always occurs ahead of the trough axis in a region of anomalous northerlies. As expected from the maps of AEWs, the meridional wind anomalies associated with convection are much weaker than observed (Fig. 6). At lag 0 there is noticeable eastward tilt with height in the meridional winds below 800 hPa. The profiles of omega are very top-heavy, with the maximum uplift above 400 hPa. The vertical profiles of diabatic heating are also top-heavy, with positive values of Q1 only above 600 hPa. While the signature of shallow convection at low levels ahead of the peak in deep convection is strong in the humidity and temperature plots, the peak in convection is associated with weak positive anomalies of moisture and temperature at the upper levels, suggesting top-heavy convection. This top-heavy convective heating profile may help to explain the deeper minima of the OLR in CCSM3, relative to SP-CCSM.

6. The energetics of AEWs

We now examine the energy sources and transformations that are important for the AEWs. The Lorenz energy diagram provided in Fig. 7 shows the processes that can potentially contribute to the eddy kinetic energy (EKE) associated with AEWs. Three processes dominate the energy cycles of the waves: barotropic conversion, baroclinic conversion, and diabatic heating. In this section, primes represent departures from a 10-day running average, so can be interpreted as synoptic-scale anomalies. This method does not separate “wave” from “no-wave” disturbances and we will actually be focusing on the energy associated with all synoptic-scale disturbances. The current analysis in terms of temporal anomalies is comparable to that of Hsieh and Cook (2007), who analyzed AEWs in terms of departures from a zonal average.

Figure 8 shows meridional cross sections of the barotropic and baroclinic sources of EKE for ERA-I and the two models, averaged between 20°W and 10°E. This range of longitudes is associated with amplified AEW activity in both ERA-I and SP-CCSM. In these figures, warm colors indicate regions where AEWs gain EKE, and cool colors show regions where EKE is lost.
In ERA-I and both models, there are three regions of EKE production due to barotropic conversions (Figs. 8a–c). The first, found at upper levels, is associated with the tropical easterly jet. The second is in the middle troposphere for the observations and both models. In both ERA-I and SP-CCSM3, this production region is centered at ~10°N and is south of the AEJ, near the ITCZ. Given the exaggerated strength of the AEWs in SP-CCSM, it is not surprising that the barotropic energy conversion term is larger in the model than in ERA-I.
CCSM3 also produces a source in the midtroposphere, but it is markedly weaker and positioned far to the south of the jet, over the Gulf of Guinea. The third production region is found at low levels in the observations and both models, near the location of the intertropical depression (ITD), where the southwesterly monsoon winds converge with the northeasterly harmattan winds. In ERA-I, this source region is centered at 17.5°N, whereas in both SP-CCSM and CCSM3 it is found at 15°N. Recall from section 3 that in both models the monsoon winds do not penetrate as far northward as observed, so the simulated confluence region of the ITD tends to be displaced southward relative to the observations. Hsieh and Cook (2007) found that the low-level production of EKE is partially offset by the loss of EKE due to frictional dissipation. Unfortunately, neither ERA-I nor the models have the diagnostics needed to perform a complete energy budget, so that possibility cannot be evaluated here.

Baroclinic conversion is also important for AEWs. In ERA-I and both models south of 12°N, positive conversions of EKE due to baroclinic processes are found in the lower and upper troposphere, with a region corresponding with the destruction of EKE in between (Figs. 8d–f). For ERA-I, the region of positive baroclinic energy conversion at low levels extends from 12° to 25°N and from the surface to 600 hPa. As discussed by Reed et al. (1977) and Hsieh and Cook (2007), the positive baroclinic conversions found in this region are associated with the ascent of warm dry air from the north and the descent of cool moist air from the south. In both SP-CCSM and CCSM3, the baroclinic term is stronger at low levels than in ERA-I, and extends farther to the south. In both models, the temperature perturbations are stronger than observed (McCrary 2012). That the baroclinic conversion to EKE extends farther south in SP-CCSM supports the idea that baroclinic conversions found in this region are associated with the ascent of warm dry air from the north and the descent of cool moist air from the south. In both SP-CCSM and CCSM3, the baroclinic term is stronger at low levels than in ERA-I, and extends farther to the south. In both models, the temperature perturbations are stronger than observed (McCrary 2012). That the baroclinic conversion to EKE extends farther south in SP-CCSM supports the idea that baroclinic conversions found in this region are associated with the ascent of warm dry air from the north and the descent of cool moist air from the south.

We now investigate potential sources of eddy available potential energy (EAPE). Figure 9 shows the meridional cross sections of the two other terms associated with the eddy available potential energy equation. The first term represents the conversion of mean potential energy to eddy available potential energy owing to the eddy heat flux along the horizontal mean temperature gradient (Figs. 9a–c). Given that the meridional temperature gradient in ERA-I and both models is largest over North Africa, it is not surprising that this term is positive north of 10°N at low levels. In this region, mean available potential energy is converted to EAPE and then to EKE via baroclinic processes.

The second term (Figs. 9d–f) represents the generation or destruction of EAPE due to diabatic processes. We will refer to this term as the “generation term,” denoted by $G$. In ERA-I diabatic processes generate EAPE north of 15°N at low levels. Farther south, centered at 10°N, $G$ is negative at low levels (750 hPa) but positive at upper levels (350 hPa) (Fig. 9d). The close correspondence of $G$ with the baroclinic term at 10°N suggests that the EKE consumed (generated) by the baroclinic term is in part generated by heating and cooling. What this tells us is that, at low levels, cold anomalies created by adiabatic cooling of rising air are balanced by convective heating ($Q_1 > 0$). At upper levels, warm descending is offset by evaporative cooling ($Q_1 < 0$). At upper levels, a positive covariance of temperature and heating acts to generate EAPE, supporting wave development. When integrated over the entire depth of the atmosphere, it appears that $G$ weakly tends to increase the EAPE between 15° and 25°N, especially over the ocean (Fig. 10).

Convection plays an exaggerated role in the energetics of AEWs in SP-CCSM (Fig. 9e). When averaged between 20°W and 10°E, the simulation is similar to the observations except that the overall magnitudes of the sources and sinks of EAPE are stronger than observed. Near 7°N, 750 hPa we see a large region where $G$ is negative. Above this, centered at 400 hPa, there is a region where $G$ is positive. Averaged throughout the depth of the troposphere, $G$ acts as a source of EAPE (Fig. 10b).

In CCSM3, the only substantial values of $G$ are at upper levels (Fig. 9f). Remember from section 5b that convection in this model is fairly top-heavy, so we would expect diabatic heating to be larger at upper levels.

In summary, we have shown that in ERA-I and SP-CCSM, strong baroclinic conversion plays an important role in dynamics of AEWs at low levels along 10°N (see section 5b).

In both ERA-I and SP-CCSM, strong baroclinic conversion is also found between 500 and 200 hPa centered over 10°N (Figs. 8d,e). This is most likely due to the ascent of warm air associated with the latent heat released owing to the convection embedded within the AEWs. Below 500 hPa the baroclinic term is negative ($\omega' T' < 0$), which can be due to ascending cold air or descending warm air. It has been suggested that dynamical forcing within the waves may cause ascending cold air (Yanai 1961; Diedhiou et al. 2002; Hsieh and Cook, 2007). The variances of vertical velocity and temperature in CCSM3 are much weaker than those in either ERA-I or SP-CCSM (McCrary 2012). At upper levels, positive baroclinic conversion occurs only above 350 hPa, and is shifted south, centered over the equator.
7. Summary and discussion

The primary goal of this paper was to investigate the sensitivity of AEWs to model physics and the subgrid-scale representation of convective processes. We have examined the variability, structure, and energetics of AEWs in two CGCMs: the standard CCSM3, which uses traditional convective parameterizations, and the SP-CCSM, where convective parameterizations are replaced by embedding a 2D CRM into each grid box. The

\[ -\frac{c_p}{T} \gamma \nabla_H T' \cdot \nabla_H T \]

\[ \frac{\gamma}{T} Q'T' \]

Fig. 9. Meridional–height cross sections of the (right) \(-T^{-1}c_p\gamma \nabla_H T' \cdot \nabla_H T\) and (left) \(+T^{-1}\gamma Q'T'\) conversion to eddy kinetic energy averaged between 20°W and 10°E from (a),(d) ERA-I, (b),(e) SP-CCSM, and (c),(f) CCSM. Energy terms are in m² s⁻² day⁻¹.

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The advantage of the SP-CCSM is that convective clouds and basic mesoscale dynamics are explicitly represented, without the immense cost of a global cloud-resolving model. The important thing to remember in this study is that both models are run at a coarse T42 resolution and that the significant differences found in the simulation of AEWs in each model is due to their different representations of moist physics.

Through this work we have shown that the incorporation of the superparameterization into CCSM3 greatly modifies how tropical convection is organized and enables the simulation of AEWs, although they are stronger than observed. CCSM3 produces little to no easterly wave activity over West Africa, and it appears that convection is decoupled from the circulation in that model. Diabatic heating profiles are top-heavy in CCSM3, and a large dry bias in the midtroposphere may inhibit deep convection. In the SP-CCSM, on the other hand, AEWs are a robust feature of the variability over West Africa. Although AEWs in SP-CCSM are much stronger than observed, their horizontal and vertical structures are similar to observations. The waves in SP-CCSM develop in central/eastern Africa, as observed, and propagate westward at slightly slower phase speeds than found in observations. The vertical structure of the waves has the form of the first baroclinic mode, with warm moist conditions found at low levels ahead of the region of deep convection, and some indication that stratiform convection may follow the region of deep convection with rain falling through dry conditions and cooling the air through evaporation. This structure is similar to that described in Kiladis et al. (2009), who studied the vertical profiles of convectively coupled equatorial waves.

The AEWs found in SP-CCSM are associated with intense convection, and therefore excessive diabatic heating. Diabatic heating generates eddy available potential energy, which serves as an important source of energy for the simulated AEWs. It appears that AEWs in SP-CCSM are maintained by the combined effects of the hydrodynamic instability of the AEJ and the convection imbedded within the waves. Additional work is needed to fully diagnose the processes that generate and maintain the waves in SP-CCSM. Forecasting experiments might be useful.

Results from the SP-CCSM are consistent with a number of other studies that have pointed to convection as an important component of wave initiation and growth (e.g., Hall et al. 2006). Our analysis of the ERA-I data shows that diabatic heating is only a weak energy source for AEWs. Keep in mind, however, that $Q_1$ is not directly available from ERA-I and was approximated for use in this study (section 2). In addition, $Q_1$ for ERA-I is
influenced by the model’s parameterizations. Perhaps $Q_1$ is underestimated in ERA-Interim and diabatic heating does play a more critical role in wave development. A comparison of the climatologies of $Q_1$ across multiple reanalysis products (Janiga and Thornicroft 2013) shows that they have qualitatively similar patterns, but quantitative differences that could influence the results. A second issue is that cumulus momentum transport is not included in SP-CCSM and may act in nature as a frictional influence that reduces the amplitude of the waves. We also note that Diaz and Aiyyer (2013) present evidence that upstream development associated with the energy dispersion from a previous AEW may act to trigger subsequent AEWs over East Africa, suggesting that convection is important for the amplification of AEWs but not their genesis.

It has been argued that higher-resolution simulations are needed to simulate the complex WAM system and African easterly waves. AEWs are relatively small features (wavelengths 2000–6000 km) that propagate quickly (phase speeds 7–9 m s$^{-1}$) across West Africa so it does seem likely that the simulated AEWs would benefit from increased spatial resolution. A moderate increase in resolution does not lead to more realistic physics, however. The point of our work is that the AEWs are also sensitive to model physics. We show that an improved representation of subgrid-scale cloud processes allows AEWs to develop even in a coarse-resolution CGCM. Not only do AEWs exist in the low-resolution SP-CCSM, but their horizontal and vertical structures are remarkably similar to observations.

Based on our analysis, we wonder if the varied wave activity found in the CMIP3 models identified by Ruti and Dell’Aquila (2010) and Skinner and Diffenbaugh (2013) may be a function of convective parameterization and the way that temperature and heating profiles are represented in CGCMs. We have demonstrated that moist physics is important for the simulation of AEWs and are currently investigating the role that convection plays in AEW variability in other CGCMs.

Taking our study together with others published earlier (e.g., Benedict and Randall 2009; DeMott et al. 2011), SP-CCSM has been shown to improve the simulation of tropical precipitation and variability over many regions of the tropics. In terms of West African climate, this is important because AEW activity varies on both interannual (e.g., Grist 2002) and intraseasonal (Janicot et al. 2011; Alaka and Maloney 2012) time scales. There is evidence that the Madden–Julian oscillation (Matthews 2004) and the Asian summer monsoon (Janicot et al. 2011, 2009) modulate African rainfall and AEW activity on intraseasonal time scales.

Because the SP-CCSM can simulate both the MJO (Benedict and Randall 2009) and the Asian summer monsoon (DeMott et al. 2011, 2013) with reasonable fidelity, it may be possible to use the model to examine the teleconnections between the Indio-Pacific and West Africa, thereby improving our understanding of seasonal and annual rainfall variability over Africa.

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