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West African extreme daily precipitation in observations and stretched-grid simulations by CAM-EULAG

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West African extreme daily precipitation in observations and stretched-grid simulations by CAM-EULAG

by

Abayomi Abiodun Abatan

A thesis submitted to the graduate faculty
in partial fulfillment of the requirements for the degree of

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Program of Study Committee:
William J. Gutowski Jr., Major Professor
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2011

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ABSTRACT

In this study, we evaluate the performance of a non-hydrostatic global climate model, CAM-EULAG (CEU), with grid stretching capability that uses NCAR Community Atmospheric Model (CAM) physics and EULAG dynamics, a non-hydrostatic parallel computational model for simulating all-scales geophysical flows developed by Smolarkiewicz and colleagues. The intent of the work is on assessing the model’s utility for West African climate study. First, we evaluate its ability to simulate the climatology of West Africa with emphasis on assessing the model’s capability for West African climate study. Second, we examine extreme precipitation events and their physical causes for West Africa. We compare CEU rainfall with Tropical Rainfall Measuring Mission and Global Precipitation Climatology Project precipitation, and simulated atmosphere with output from a global atmospheric reanalysis (ERA-Interim) produced by the European Centre for Medium-Range Weather Forecasts. We find that the model simulates well the mean climate over West Africa during the summer monsoon season, July–September. It reproduces the mean rainfall at the peak of the West African summer rainy season, and it captures the rainbelt associated with the ITCZ. The model simulates the core of the rainbelt consistent with the core of the deep ascent lying between the axes of the African Easterly Jet and the Tropical Easterly Jet. The model simulates fairly well the interannual and intraseasonal variability of the extreme precipitation events. Although they are not as intense as observed, the spatial scale of extreme events in the model is comparable to the observed scale. Simulated large-scale processes on extreme-event days compares well with corresponding ERA-Interim fields on observed extreme-event days. The thesis concludes with recommendations for further analysis and improvements of the model.
CHAPTER 1. GENERAL INTRODUCTION

1. Introduction

The issues of climate change and changes in extreme events have been subjects of concern receiving considerable attention recently from governments (Chen et al., 2010) and institutions. The Intergovernmental Panel on Climate Change (IPCC, 2001) defines an extreme weather event as an event that is rare within its statistical reference distribution at a particular place. Definitions of "rare" vary, but an extreme weather event would normally be as rare as or rarer than the 10th or 90th percentile. Rare events such as erosion and flooding resulting from increases in precipitation extremes can produce severe impacts on physical, biological, agricultural and socio-economic systems over many regions of the world (Nicholls and Alexander, 2007; Bonsal and Kochtubajda, 2009; Maraun et al., 2009; Roy, 2009).

Substantial changes in temperature resulting from climate change can produce changes in other atmospheric fields such as moisture and thus produce changes in precipitation characteristics including extreme behavior. In fact, there already exists an increase in the occurrence of precipitation events over most land areas in many parts of the world (Jones and Hulme, 1996; Dai et al., 1997; Hulme et al., 1998; Karl and Knight, 1998; Doherty et al., 1999), although large regional differences exist in the increase. An increase in intense precipitation event is consistent, not only with changes in overall precipitation, but also with increasing temperatures and atmospheric water vapor (Trenberth et al., 2007; Easterling et al., 2000). The increase in extreme precipitation has been most pronounced in middle and high latitudes, where total precipitation has increased. However, increases in extreme precipitation have also been documented in regions such as northern Japan (Manton, et al., 2001) where there has been a reduction in total precipitation amount (IPCC, 2007).
Because of the importance of extreme events to humans, ecosystems and socio-economic systems, considerable efforts have been devoted to research on extreme temperature and precipitation events in many regions of the United States (Aguilar et al., 2005; Gutowski et al., 2007, 2008, 2010; Peterson, 2008), the Caribbean (Peterson, 2002), Mexico, Canada (Wang and Zhang, 2008), Europe (Christensen and Christensen, 2003), Asia (Griffiths et al., 2005), Australia, regions of Africa (Aguilar, 2009; New et al., 2006), China (Chen et al., 2010; Wang and Zhou, 2005; Zhai and Pan, 2003) to mention but a few. The observational and numerical modeling results of the above mentioned studies have shown that there are remarkable increases in intensity of precipitation extremes. However, there have been relatively few observational and simulation studies of extreme precipitation events in West Africa.

2. Thesis Organization

The remaining chapters of this thesis are arranged as follows. First, a literature review is presented in section 3. A performance assessment of the model in chapter 2 shows that the model can be used for impact study over West Africa. Chapter 3 is a study of extreme daily precipitation events and its physical causes in West Africa. General conclusions and recommendations for future work are discussed in Chapter 4, followed by references.

3. Literature Review

3.1. West African Climate

West Africa is that part of Africa that lies approximately between 5° N and 20° N and occupies an area of approximately 5 million km². It is bounded on the west and south by the Atlantic Ocean and on the north by the Sahara desert. The eastern border lies on a line running from the Cameroon Mountains to Lake Chad. The region can be subdivided using mean annual rainfall into zones: Sahelo-Sahara, Sahel, Soudan, Soudano-Guinea and Guinea savannah. The climate of West Africa is
characterized by wet and dry seasons. The weather pattern is associated with the northward and southward migration of a narrow zone of reversal in the meridional wind, called the Intertropical Discontinuity (ITD). It is a region of trade-wind confluence, which produces weak horizontal pressure gradients responsible for weak winds at the surface. Another commonly used term in the literature is the Intertropical Convergence Zone (ITCZ), associated with the zone of maximum convection. Both the ITD and ITCZ exhibit seasonal migration following the seasonal movement of the overhead sun. The wind systems associated with the ITD are characterized mainly by the northeasterly and southwesterly trade winds. During the wet season, the moist southwest monsoon with its maritime characteristics from the Gulf of Guinea invades the region, bringing with it cool breezes. It is often associated with convection and cloudiness. The northeast trade wind, which characterizes the dry period, on the other hand is continental, hot, dry and dust-laden because of its long track from the Sahara desert. During this period, the harmattan wind blows southwestwards across the region, sometimes reducing visibility to less than 1000 m.

### 3.2. Rainfall variability and extreme events

The inhabitants of West Africa depend solely on rainfed agriculture for sustenance. However rainfall variability – arising from the interaction of the region's climate with large-scale atmospheric circulation – is evident through changes in extreme precipitation events such as erosion and floods. These remain a major challenge for increased agricultural production, necessary for food security.

The interannual and intraseasonal rainfall variability in West Africa, as in other African countries, has been documented by several investigators in numerous publications. There is a strong link between interannual rainfall variability in West Africa and patterns of sea surface temperature (SST) anomalies in the tropical Atlantic, Pacific and Indian Oceans (Ward et al., 1990; Folland et al., 1991; Ward, 1992; Shinoda and Kawamura, 1994; Nicholson and Grist, 2001; Rowell, 2001; Bader and Latif, 2003; Giannini et al., 2005). Compositing five Sahelian wet and dry years, Ward (1992) found
that SST forcing from all three major ocean basins may contribute to seasonal Sahelian rainfall variability. The Sahelian rainfall variability has been associated with warming in the tropical Indian Ocean SST since 1950s (Bader and Latif, 2003). The authors show that the warming trend in the Indian Ocean played a crucial role for the drying trend experienced over West Sahel in the 1950s to 1990s. The weakening of the large-scale zonal gradient of SST from the western Pacific to the eastern Indian Ocean has been suggested to enhance the likelihood of Sahel drought (Rowell, 2001).

The El Niño-Southern Oscillation (ENSO) in the tropical Pacific Ocean (e.g., Nicholson, 1997; Nicholson et al., 2000; Nicholson and Selato, 2000; Hulme et al., 2001) has been confirmed as one of the more important factors influencing rainfall variability for some regions in Africa. Hulme et al. (2001) in their detailed analysis of African climate change observed a strong ENSO relationship for equatorial east Africa (high rainfall during a warm ENSO event) and southern Africa (low rainfall during a warm ENSO event), consistent with earlier studies. Elsewhere in Africa, West Africa in particular, there has been a controversy on the influence of the ENSO on rainfall. While there is a general consensus among researchers on ENSO’s influence in some regions, for instance the Guinea coast, where it tends to increase rainfall (Nicholson, 2001), there is a controversy over its influence in the Sahel. The authors cited at the start of this paragraph found weak correlation between ENSO and Sahelian June-August drying, consistent with Ropelewski and Halpert (1987). For example, although dry conditions prevailed throughout the Sahel and most parts of West Africa in 1997, El Niño appears to have had minimal impact on the dry conditions experienced in the region (Nicholson et al., 2000). This suggests that other factors may be important. Others such as Semazzi et al. (1988), Hastenrath, (1990), and Ward (1992) to mention but a few suggest that the impact is larger. Ward (1992) further notes that ENSO’s influence appear to be greater during dry years than wet years. The different opinion among several authors is due to the complex nature of ENSO's influence in the region.
More recently, the causes and the impacts of the 2007 flooding in most parts of the sub-Saharan Africa and the Sahel in particular have been documented in literature. This flooding left thousands of people homeless and impoverished. Among the potential causes of the flood is the La Niña event that year in the tropical Pacific (Paeth et al., 2010). The authors observed that anomalous SST in the tropical Atlantic Ocean coupled with a La Niña episode in tropical Pacific favors stronger than normal precipitation, as occurred during the 2007 rainy season in West Africa.

In addition, significant changes in the intensity and location of the lower and upper tropospheric jets – the African easterly jet (AEJ) and the tropical easterly jet (TEJ) – during the summer monsoon are known to play an important role in modulating rainfall over West Africa. The meridional migration of the interaction between the dry continental air mass and the moist southwesterly air mass (monsoon) leads to a seasonal reversal of the land-sea temperature gradient with altitude and the generation of a strong thermal wind, AEJ, above 700 mb. A number of earlier studies (Kanamitsu and Krishnamurti, 1978; Newell and Kidson, 1984) showed a consistent relationship between the AEJ and West Africa rainfall variability. The strength of AEJ is linked with warming in three ocean basins – South Atlantic, Indian and Pacific - and stronger AEJ results in an equatorward shift of the zone of convection leading to reduced rainfall in the regions north of latitude 12° (Adedoyin, 1997). Using a conceptual model for Sahel rainfall variability and compositing wet (1958-1967) and dry (1968-1997) years, Nicholson and Grist (2001) showed that the location of the AEJ and the associated shear instabilities are the most important local factors controlling Sahel rainfall variability. During wet years, deep, well-developed, low-level westerlies displace the AEJ northward and increase the shear instabilities. In contrast, the dry years are characterized by poorly developed equatorial westerlies, resulting in a significant southward shift of the AEJ south of the Sahel.

The variability of West Africa rainfall is also correlated with variations in the well-organized wave disturbance, African easterly waves (AEWs). The waves originate between 20° - 30°E as a result of
barotropic and baroclinic conversions of energy from the mid-tropospheric AEJ (Charney and Stern, 1962; Carlson, 1969a, b; Burpee, 1972, 1974; Rennick, 1976; Albignat and Reed, 1980; Kwon, 1989; Laing and Fritsch, 1993; Thornicroft and Hoskins, 1994a, b; Thornicroft, 1995; Chen 2006). They are observed to propagate westward in the lower troposphere of the tropical North Atlantic. Burpee (1972, 1974) found that the westward propagating disturbances are concentrated between 5° N and 15° N. The AEW activity occurs during the summer monsoon from June to September with maximum intensity in August. AEWs are generally recognized as playing an important role in modulating daily rainfall over West Africa during the summer monsoon season (Thorncroft, 2001; Fink and Reiner, 2003).

Even though some of the work above discussed precipitation extremes, most of the work focused on mean behavior. Studies such as Alexander et al. (2006) have shown that analyzing changes in extremes requires daily data. Even at that, there have been only a few studies (Easterling et al., 2003; New et al., 2006) that have examined changes in African daily precipitation extremes. Studies on extremes using daily data has not been possible in West Africa partly because networks of observing stations with sufficient quality and quantity of data have not been available to the scientific community. This problem stems from a number of factors ranging from political instabilities to government bureaucracy on data policy issues, which, starting from the mid-1980s, plagued most of the countries in the region. Analysis of annual precipitation data in Africa shows that precipitation has increased in parts of western Africa (Easterling et al., 2003), though large areas remain unanalyzed. More recently, the Working Group on Climate Change Detection and Indices, part of the joint World Meteorological Organization Commission for Climatology (CCI)/World Climate Research Programme (WCRP) project on Climate Variability and Predictability (CLIVAR) organized regional climate change workshops to address some of the issues of data availability in many parts of the world including West Africa. Based on a workshop held in Casablanca, Morocco, from 18 to 23 February 2001 with 23 African countries in attendance, an analysis of annual precipitation using the 95th
percentile showed that extreme precipitation events had increased at some stations but decreased at others (Easterling et al., 2003). Consistent with the Casablanca workshop, the WMO/CLIVAR and START cosponsored southern Africa climate extremes workshop held in Cape Town, in June 2004, found that there has been statistically significant increase in regionally averaged daily rainfall intensity (New et al., 2006). From all the international daily data sets available, Alexander et al. (2006) found that precipitation indices show a tendency toward wetter conditions throughout the 20th century. However, none of these studies have attempted to study daily precipitation extremes and causal behavior using numerical simulation.

The aim of this study is in two-fold. First, we use a climate model, CAM-EULAG (CEU), to simulate the climatology of West Africa with emphasis on assessing the model’s capability for West African climate study. Second, we examine extreme precipitation events and physical causes over West Africa where extensive study is lacking as a result of past paucity of data.

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CHAPTER 2. Assessing the capability of CAM-EULAG to simulate rainfall and associated dynamics over West Africa

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Abayomi A. Abatan, William J. Gutowski Jr., and Babatunde J. Abiodun

Abstract

This study evaluates the performance of a non-hydrostatic global climate model with grid stretching capability (CEU) that uses NCAR Community Atmospheric Model (CAM) physics and EULAG dynamics, a non-hydrostatic, anelastic, parallel computational model for simulating all-scales geophysical flows, to simulate the spatiotemporal variability of West African climate and associated dynamics. We compare CEU rainfall with Tropical Rainfall Measuring Mission and Global Precipitation Climatology Project precipitation, and CEU winds and temperature with global atmospheric reanalyses (ERA-Interim, ERAIM) produced by the European Centre for Medium-Range weather Forecasts.

The prominent features of the West African monsoon circulations and the general pattern of rainfall in the simulation compare well with observations. The model simulates the mean rainfall at the peak of the West African summer rainy season (July–September), and it captures the rainbelt associated with the ITCZ around 12°N. Examination of the intraseasonal variability of rainfall shows that the model captures the three distinct phases of the West African monsoon circulation: onset, peak, and cessation phases. Furthermore, the meridional migration of the rainfall shows that the rainbelt propagates northward reaching its northward limit in August. The seasonal migration of the rainbelt is linked to the northward excursion and weakening of the African Easterly Jet (AEJ) and the
appearance and intensification of the Tropical Easterly Jet (TEJ). The model shows that the core of the rainbelt coincides with the core of the deep ascent lying between the axes of the AEJ and the TEJ. However, there exist notable discrepancies between the simulations and observations. The model shows a tendency to underestimate rainfall over orographic regions. Also, it simulates too little summer rainfall over the Guinean coast while it simulates too much rainfall over the Soudano-Sahel region.

1. Introduction

West African economy and food supply are highly dependent on agricultural production. Agriculture provides the major source of income and livelihoods for a large portion of the population, and about 17–50% of GDP in some West African countries (Schlenker and Lobell, 2010). However, agriculture is heavily dependent on climate, with the seasonal characteristics of rainfall (Omotosho, 1990; Sultan and Janicot, 2003a; Sultan et al., 2005) being the primary factor. The climate of West Africa is characterized by two distinct seasons, the dry and the rainy seasons. The dry season extends roughly from mid-October to mid/late March, while the rainy season starts gradually from the months of April and May, reaching its peak in the months of June, July, August, and September.

The spatial pattern of rainfall in this region has been documented (e.g., Nicholson, 1980; Rowell et al., 1995). West African rainfall exhibits a roughly zonally elongated structure, characterized by a seasonal northward-southward migration of the rainbelt. The migration is consistent with the seasonal excursion of the Intertropical Convergence Zone (ITCZ) and follows the seasonal movement of the sun. The annual rainfall decreases sharply from the Guinean coast to the northern fringe between the Sahara and the Sahel. Rainfall varies from about 1000–2000 mm year$^{-1}$ near the coast at 5°N to less than 100 mm year$^{-1}$ at the northern fringe. Rainfall in regions north of 20°N tends to occur largely in just one or two months (Nicholson, 1979; Eltahir and Gong, 1996).
West Africa rainfall depends on the West African Monsoon (WAM), a large-scale circulation that forms from the seasonal reversal of winds due to the land-sea temperature contrast between the Sahara desert and the equatorial Atlantic Ocean. The largest portion of the rainfall is from deep convective systems (e.g., squall lines and mesoscale convective systems (MCSs) and the ITCZ) embedded within the WAM system during the months of July, August, and September (Reed et al., 1977; Lavaysse et al., 2006; Fortune, 1980). The onset and end of the rainy season, the length of rainfall period and the annual rainfall amount are controlled by the northward progression of the WAM (Omotosho, 1985; Ward, 1992; Cifelli et al., 2010).

Rainfall is, of course, highly variable on spatial and temporal scales (Maloney and Shaman, 2008; Sultan, et al. 2003a), and such variability can result in unpleasant conditions in the region. In the last few decades, the region has experienced severe and protracted droughts as a consequence of significant decline in rainfall during the main rainy season, June–September. The severe droughts have had disastrous impacts on the socio-economic development of the region, with the Sahel mostly affected. Drastic agricultural loss and freshwater shortages have impacted the region, due to a 30–40% decline relative to 30-year rainfall means (Nicholson and Webster, 2007). In addition, significant long-term change in rainfall has occurred in the humid region of the Guinea coast (Eltahir and Gong, 1996; Nicholson et al., 2000).

The causes of rainfall variability in West Africa, though not fully known (Rowell et al., 1995), have been the focus of several studies. Changes in intensity and distribution of rainfall have been attributed to the complex land-atmosphere interaction in the regional climate, ocean-atmosphere interaction, land-use change, anomalous tropical atmospheric circulation and changes in mean climate resulting from anthropogenic greenhouse forcing (e.g., Folland et al., 1991; Fontaine et al., 1995; Rowell et al., 1995; Fontaine and Janicot, 1996; Wang and Eltahir, 2000).
West African rainfall variability has been linked to anomalous tropical atmospheric circulations. A number of studies have shown that the African easterly jet (AEJ), Tropical easterly jet (TEJ), African easterly waves (AEWs), squall lines, and deep convective systems are among the prominent features of African climate that significantly modulate the spatio-temporal distribution of rainfall during the boreal summer season (e.g., Carlson, 1969; Burpee, 1972; Reed et al., 1977; Kanamitsu and Krishnamurti, 1978; Houze and Bett, 1981; Bolton, 1984). A stronger AEJ results in an equatorward shift of the zone of convection leading to reduced rainfall in the regions north of latitude 12° (Fontaine et al., 1995; Adedoyin, 1997). Anomalous low Sahelian rainfall is also related to weakening of the TEJ (Newell and Kidson, 1984; Fontaine et al., 1995). In contrast to the dry years in the Sahel, the wet years are characterized by the strengthening and northward displacement of the AEJ, accompanied by rain-bearing systems (Nicholson and Grist, 2001).

The role of AEWs in modulating daily rainfall over West Africa has been acknowledged by many (e.g., Fink and Reiner, 2003; Moron et al., 2008; Nicholson, 2009). Fink and Reiner (2003) suggest that the impact of AEWs on the squall line genesis is greater at the peak of the Sahelian rainy season. The reduction of the growth rate of mid-tropospheric transient AEWs, which form as a result of instability of the AEJ, leads to a reduction in the frequency of generation of organized squall lines, which are embedded within AEWs (Adedoyin, 2000; Druyan et al., 2008). Fewer squall lines, the main rain-bearing systems in West Africa, means less rainfall.

Understanding the dynamics and spatio-temporal variability of these features and associated rainfall is crucial in assessing the potential for enhancing water resources management, improving food security, and alleviating poverty in West Africa. These have prompted a number of previous studies, using operational radiosonde observations, reanalyses and models, to examine the characteristics of West African rainfall variability (e.g., Carlson, 1969; Burpee, 1972; Rowell et al., 1995; Grist and Nicholson, 2001; Nicholson and Grist, 2003; Cifelli et al. 2010). Several studies
suggest that regional climate models (RCMs) simulate realistic interannual variability of rainfall over West Africa (for example, Afiesimama et al. 2006; Patricola and Cook, 2010a,b). Patricola and Cook (2011) constructed an ensemble of nine future climate scenarios for northern Africa at the end of the twenty-first century with one RCM, the Weather Research and Forecasting (WRF) model. They argued that the RCM produces a realistic precipitation projection over much of the northern tropical Africa, with projected midsummer drought over Guinean Coast. Although RCMs are suitable tools for simulating the WAM, and may offer the best solution to understanding climate change in West Africa because of their higher horizontal resolution and capability to resolve small scale atmospheric features (Jenkins et al., 2002), they do have limitations. In the tropics, RCMs have problems resolving inertia-gravity waves in dispersing heat from rainfall (McGregor, 1997). Similarly, RCMs can provide the more intense events, but may still miss the most extreme precipitation (Gutowski et al., 2003, 2007).

Many studies have used atmospheric Global Climate Models (GCMs) to study climate change over Africa (e.g., Hulme 1994; Held et al. 2005; Hoerling et al. 2006; Abiodun et al. 2008). GCMs typically use relatively coarse spatial resolution on the order of only a few hundred kilometers, e.g., 150–450 km (IPCC, 2001, 2007), much larger than the characteristic scales of most atmospheric features that dominate African climate. Typically, the models are unable to resolve important sub-grid scale features, such as clouds and topography, because of their coarse spatial resolution. Given these limitations, GCMs fail to adequately reproduce the historic regional climate (Giorgi and Mearns, 1999; Yates, 2003), as well as many small-scale processes. GCMs give only a simplified rendition of synoptic-scale and large-scale atmospheric features (Denis et al., 2002). Given these limitations, some now construct climate scenarios with higher resolution atmospheric GCMs (e.g, 0.5° latitude/longitude). Even at higher resolution, there are still some uncertainties in the simulated climate. For example, high resolution atmospheric GCMs may produce much stronger than observed
surface westerlies in the Northern and Southern Hemisphere winter (Manabe et al., 1979; Held and Phillips, 1993; Jones et al., 1997).

Higher resolution GCMs with stretch-grid capability are recently being employed in climate simulations (e.g., Fox-Rabinovitz et al., 2006). Stretch-grid models feature an advantage over uniformly high resolution GCMs in that highest resolution may focus on one or more regions of interest, with coarser, less computationally demanding resolution elsewhere. Grid stretching offers potentially improved climate simulation and enhanced understanding of regional climate over the selected region. For example, some studies indicate that stretch-grid models give a more realistic simulation of the monsoon system (Mo et al, 2005; Abiodun et al., 2011).

In this study, we assess the capability of CAM-EULAG, a high resolution global model with grid-stretching capability, to reproduce the interannual variability of rainfall and associated atmospheric circulation over West Africa. The assessment involves examining climatological mean fields. Although there have been several modeling studies in Africa, there has been relatively little study on the performance capability of CEU on African climate (e.g., Abiodun et al., 2011). In section 2, we present some salient details of the model, setup, and the description of other data used to validate the simulation. In section 3, we present the results, while conclusions derived from the study appear in section 4.

2. Model description and observational dataset

This study uses an atmospheric global circulation model (AGCM), CAM-EULAG (hereafter, CEU) to study West African climate. CEU is constructed using EULAG dynamics (Prusa et al., 2008) and the Community Atmospheric Model (CAM) physics (Collins et al., 2004, 2006), developed by the U.S. National Center for Atmospheric Research (NCAR).
2.1 EULAG

EULAG is a non-hydrostatic, anelastic, parallel computational model for simulating all-scales geophysical flows (Smolarkiewicz and Prusa, 2005). It features a generalized coordinate formulation enabling grid adaptation (Smolarkiewicz and Prusa, 2005), non-oscillatory forward-in-time integration algorithms (Smolarkiewicz, 2006; Prusa et al., 2008), and robust elliptic solver (Smolarkiewicz et al., 1997). EULAG’s all-scale capability has permitted its successful application at scales ranging from centimeters (Andrejczuk et al., 2004) to $10^6$ km (Elliott and Smolarkiewicz, 2002). For example, EULAG has a proven record of being successfully applied in the areas of turbulence, gravity wave dynamics, urban flows, flows past complex/moving boundaries and micrometeorology (Smolarkiewicz and Prusa, 2002, Prusa et al., 1996; Prusa and Gutowski, 2006). Further details on EULAG appear in Smolarkiewicz and Pudykiewicz (1992), Prusa et al. (1996) Prusa and Smolarkiewicz (2003), Prusa and Gutowski (2006), Abiodun et al. (2008), Prusa et al. (2008) and Abiodun et al. (2011).

2.2 CAM

The physics used in CEU come from CAM 3.1, which has been the atmospheric component of the NCAR Community Climate System Model (CCSM). The CAM physics include precipitation processes (e.g., moist deep convection, shallow convection and large-scale stable condensation), turbulence mixing, clouds and radiation, and ocean-surface and land models. Of relevance to this study, deep convection is based on a plume ensemble approach where updrafts and downdrafts are related to the interaction between cloud base mass flux and the convective available potential energy (CAPE) (Zhang and McFarlane, 1995; Collins et al., 2004). The CAM employs Hack (1994) mass flux scheme to represent shallow convection, and Sundqvist (1988) bulk microphysical parameterization scheme for evaporation of the convective precipitation, while the parameterization of non-convective clouds processes uses prognostic condensate scheme (Rasch and Kristjánsson,

2.3 CEU

Detailed description of the coupling of CAM and EULAG appears in Abiodun et al. (2008). We run CEU with a static stretched grid that concentrates grid points in the vicinity of West Africa (Fig. 2.1). The highest grid stretching centered over West Africa has latitude-longitude resolution of 0.5° X 0.5°, and extend zonally from about 0° – 40°N. The resolution elsewhere becomes as coarse as 4° (lat) and 2.8° (lon). The model uses 26 vertical levels, with model top at 30 km. Like other grid-point GCMs, the grid boxes of CEU tend to become elongated at the poles, a feature than can produce noisy behavior on small scales and contribute to numerical instability (Abiodun et al., 2008). To minimize such behavior, latitudes poleward of 75° have a sponge layer to damp the state of the atmosphere toward the zonal average. Abiodun et al. (2008) show that this damping has little influence on simulations at lower latitudes.

We ran model using observed sea surface temperatures and sea ice distributions obtained from the standard CAM package. The simulation ran from 1 January 1996 to 30 January 2008. We discarded the first two years for spin up and analyzed the 10 year period 1998-2007.

2.4 TRMM, GPCP, and ERA-Interim data sets

We used three sources of observation-based data to evaluate the simulation. For precipitation we used the gridded precipitation provided by Tropical Rainfall Measurement Mission (TRMM). Specifically, we used the multi-satellite 3B-42, version 6, product, which is on a 0.25° by 0.25° grid in a global belt extending from 50°S to 50°N. The TRMM 3B-42 product consists of rainfall
estimates derived from several satellite measurements including the TRMM Microwave Imager (TMI), Visible-Infrared Scanner (VIRS) and TRMM precipitation radar. Huffman et al. (2007) give the details of the data assimilation and their quality control. We used daily rainfall aggregated from 3-h values for 1998 to 2007, consistent with our simulation. In order to facilitate comparison of the simulation with observations, we regridded the rainfall dataset to a 0.5° X 0.5° grid. Analysis of a subset of the data showed that regridding did not alter substantially the daily spatial pattern of precipitation seen on the 0.25° X 0.25° grid.

In addition, we compared simulated precipitation with a gridded precipitation data set from the Global Precipitation Climatology Project (GPCP), specifically the Global Daily Merged Precipitation Analyses on a 1° X 1° grid (Huffman et al., 2001). GPCP precipitation estimates are derived from satellite sensors and rain gauge measurements.

We compared the simulated atmosphere with output from the ERA-Interim reanalysis (Berrisford et al., 2009; Dee et al., 2011). ERA-Interim reanalysis (ERAIM) datasets are available at a 1.5° X 1.5° resolution, with 37 pressure levels from 1000 to 1 hPa, for the period 1979 - 2010. The gridded data products available online include, among others, 6-hourly (0000, 0600, 1200, and 1800 UTC) upper-air fields covering the troposphere and stratosphere. For this study, we use data from 1998 to 2007.

3. Results and discussion

In this section, we examine and compare the mean rainfall climatology from TRMM and GPCP observations with the CEU simulation. The simulated synoptic circulation features, such as AEJ and TEJ, are compared with ERAIM dataset. We present results for July, August and September (JAS) for 10-year period, 1998 – 2007. We focus on JAS because that is the peak of the monsoon season, a critical period for the West African economy and thus a critical period to simulate well.
3.1. Climatology of West African summer rainfall and atmospheric systems

3.1.1. Rainfall climatology in JAS

Figure 2.2 shows mean rainfall (shaded) superimposed with mean surface temperature (contour) for July through September (JAS) during 1998 - 2007, from TRMM, GPCP, ERAIM (temperature) and CEU. A prominent feature in the rainfall distributions is a band of maximum rain associated with the ITCZ in the Atlantic Ocean, with maximum rainfall of about 16 mm day$^{-1}$ in both TRMM (Fig. 2.2a) and GPCP (Fig. 2.2b), and about 10 mm day$^{-1}$ in CEU (Fig. 2.2c). TRMM and GPCP places the rainbelt in a broader zonal band between 5° and 12°N, with the core located around 8°N. CEU however features a weaker core located around 12°N. Another interesting feature of the rainfall climatology over land in the TRMM data is the zone of maximum rainfall in localized highlands in Guinea, the Cameroon Mountains and the Jos Plateau. Fine scale features are obvious over land, with rainfall over the Cameroon Mountains of the same magnitude as the maximum embedded within the ITCZ over the Atlantic Ocean. The orographic related rainfall maximum found over the Cameroon Mountains extends northeast-southwest, with an anvil-like shape encompassing Nigeria on the left and Lake Chad to the right. GPCP features a pattern similar to TRMM but with lesser maximum magnitude (10 mm day$^{-1}$, Fig. 2.2b). The smoother features in GPCP could be due to the coarser resolution. The CEU (Fig. 2.2c) simulates the fine scale features in West Africa rainfall consistent with TRMM. The model captures the localized rainfall maximum over Cameroon Mountains with a northeast-southwest tilt in a narrow strait, except that it underestimates the maximum rainfall (8 mm day$^{-1}$) and misses completely the rainfall maxima over the Jos Plateau, Nigeria. However, over land between 5°W and 5°E and from 8° to 14°N, the model simulates pockets of maximum rainfall. Immediately above the rainbelt (Fig. 2.2), rainfall decreases northward as far as the southern fringe of the Sahelian region at about 16°N.
In general, CEU simulation shows some disparity in the rainfall climatology pattern compared with TRMM and GPCP. It has a local minimum rainfall over Guinea highland region rather than a local maximum. It fails to capture the rainfall maximum over Cameroon Mountains and Jos Plateau. Similarly, region of reduced rainfall over land, 8°–13°W, 6°–16°N, can be seen. However, the fine scale features associated with the westward extension of the rainbelt over 5°W–5°E, 6°–16°N are well captured by the model. Over the Atlantic Ocean, the rainband concentrated along the ITCZ are less well simulated; the model underestimates the westward extend of the ITCZ and the magnitude of the rainfall. Generally, precipitation decreases south and north of the maximum rainbelt. It is obvious that uncertainties exist in TRMM and GPCP rainfall climatology, which could be related to the higher resolution of the TRMM data. The underestimation of rainfall over highlands may be linked to the model's failure to resolve topography accurately. On the other hand, since the model captures the climatological pattern of the rainfall, the differences in simulated rainfall intensity may be due to other factors. Similar underestimates of rainfall have been observed in regional model simulations over West Africa and East Africa, for example, Jenkins (1997) and Sun et al. (1999). Afiesimama et al. (2006) suggests that the underestimated rainfall may be related to the choice of convection scheme used. Overall, the rainfall pattern is well captured by CEU.

3.1.2. Temperature climatology in JAS

Another distinctive feature of the West African climate is the temperature evolution. The West African climate is characterized by latitudinal variation in mean annual temperature, which increases from the Guinea coast to the Saharan region. As a result, a large temperature gradient exists. The largest contrast in air temperature between the Guinea coast and the Sahara is related to the different airmasses over these regions. The coastal region is under the influence of moist maritime airflow from the Gulf of Guinea, while the dry, hot continental air stream from the Sahara desert dominates the Saharan region.
Figure 2.2 depicts the spatial distribution of mean JAS surface air temperature from ERAIM and CEU simulation superimposed on the mean summer rainfall. The reanalysis dataset (Fig. 2.2a) features a pronounced increase in temperature from about 20°C in the south to about 34°C in the north. Comparisons between rainfall and temperature shows that regions between 4° and 16°N dominated by low temperatures coincides with the rainbelt over West Africa. The temperature gradient between northern and southern zones of about 14°C suggests that there is a feedback between rainfall and temperature, and the effect is to lower the surface temperatures (Nicholson and Grist, 2003). Lower values over the highlands of the Jos Plateau (24°C), Cameroon Mountains (22°C) and Fouta Djallon in Guinea (23°C) and higher values (> 34°C) over northern Sahel/central Sahara are observable. The differences in temperature over these regions between the warmer and cooler regions imply that topography plays a significant role in West African climate. The strong temperature gradient within the latitudinal band 12°–18°N separates the temperature patterns over coastal zones and the Saharan region. In addition, as can be inferred from the figure, two distinct temperature maxima are located at about 24°N. The first in the northwest peaks at about 35°C and second in the northeast (Ahaggar Mountainous regions in central Sahara) peaks at about 28°C. The distinction between temperature in the east and west suggests that the regions are dominated by different dynamic controls (Hastenrath, 1990); hence they show somewhat different patterns of temperature (Nicholson et al., 2000).

The model (Fig. 2.2b) temperature patterns agree fairly well with ERAIM (Fig. 2.2a), but the model underestimates temperature amounts by about 1–2°C over highlands, as well as over northwestern regions of West Africa. The north-south gradient in temperature is well simulated, though shifted northwards by a few degrees to the northern fringe of the Sahel. In addition, the model has temperature too low over the highlands of the eastern sector. Overall, the model performs fairly well in simulating regional temperature climatology over West Africa.
3.1.3. Zonal wind climatology in JAS

Figure 2.3 shows the spatial distribution of mean JAS zonal wind superimposed with wind vectors at 850 hPa over West Africa for ERAIM and CEU, respectively. The most prominent differences in these figures are the intensity and structure of the equatorial westerlies and the Saharan easterlies. In ERAIM (Fig. 2.3a), a narrow zone of maximum westerlies is located between 6° and 10°N. The westerlies over the Atlantic Ocean extend eastward as far as 20°E, with maximum westerly flow of 5 m s\(^{-1}\) centered at approximately 12°E over the Cameroon Mountains. Weak westerlies (1 m s\(^{-1}\)) over the continent extend northwards as far as 20°N, overlain by easterlies over the Saharan region and Atlantic Ocean. Two peaks of easterlies are found embedded between latitude 15° and 24°N. The first is located between 25° and 30°W over the oceanic area and has maximum value of about 7 m/s. The second is located over land between 12° and 17°E and is comparable in magnitude to the first.

Compared to ERAIM, the model’s pattern of westerlies and easterlies show some similarity to the reanalysis (Fig. 2.3b), although CEU winds tend to be stronger and its West African westerly jet spans more latitudes. The westerly jet is located at 12°N between 18° and 22°W with maximum speed of 10 m/s. Over land, the westerlies extend zonally between 6° and 16°N with local westerly maxima of 10 m/s concentrated at approximately 12°N. Also, the model simulates the maximum westerly flow at 12°E (compare Fig. 2.3a), but locates it farther north, along Lake Chad at 10°N. Figure 2.3b, however, shows that the model simulates monsoon that penetrates northward to about 16°N, covering the lower half of the Sahel. The difference in the northward extent of the monsoon between ERAIM and CEU may be due to the differences in the strength of the simulated westerlies, in particular, the westerly jet. In fact, studies have shown that the low-level westerly jet is an important agent for transporting moisture from the eastern tropical Atlantic Ocean into continental West Africa during boreal summer (Patricola and Cook, 2007; Pu and Cook, 2010). The northern hemisphere easterly flow stretches zonally, in a wave like manner, along the northern axis of the low level monsoon with...
peaks around 10-12 m s\textsuperscript{-1}. Compared to ERAIM, the model overestimates the easterlies located between 20° and 30°W over the ocean. The latitudinal distribution of rainfall as seen in Figure 2.2 is consistent with the core of the equatorial westerlies, suggesting the role of the enhanced southwesterly monsoon flow on rainfall.

Since the variability of the WAM rainfall is linked to the location and strength of easterly jets and convection, we examine the relationships between rainfall and easterly jets and vertical winds. Figure 4 portrays the vertical structure of the JAS wind fields as a function of latitude averaged over 10°W to 10°E for ERAIM and CEU. In ERAIM (Fig. 2.4a), the low-level westerly monsoon flow is located below 800 hPa between 2° and 17°N. The horizontal extent acts as boundary separating the shallow moist southwesterly flow from the easterly flows associated with sinking motions located between 25° and 31°N, and 3°N – 10°S. The shallow westerly monsoon flow has maximum speed of 4 m s\textsuperscript{-1} at 950 hPa between 7° and 15°N. Also evident in the upper troposphere are the two mid-latitude westerly jets. Compared to the ERAIM, the model simulates the upper-tropospheric westerlies and the low-level westerly wind (Fig. 2.4b). The model overestimates the magnitude and depth of the West African tropical westerlies. The model’s tropical westerlies extend upward as far as 700 hPa, with a maximum speed of 8 m s\textsuperscript{-1} located between 10° and 12°N. Also, the model simulates a horizontal extent of the westerlies that, at the surface is similar to the ERAIM extent in Figure 2.4a. The two upper-tropospheric westerlies are well reproduced, although the model underestimate the strength of the westerly jet at 30°N by about 4 m s\textsuperscript{-1}, while it overestimate the southern hemispheric westerly jet located at 10°S by about 6 m s\textsuperscript{-1}.

Figure 2.4 shows other prominent features of the zonal circulation over West Africa: upper-tropospheric easterly jet (TEJ), mid-tropospheric easterly jet (AEJ), and low-level easterly jet. The observed AEJ, with core speed of 12 m s\textsuperscript{-1} at 600 hPa, centered around 14°–15°N, appears immediately north of the monsoon layer between 520 hPa and 700 hPa. The observed TEJ appears at
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200 mb with core speed of 14 m s\(^{-1}\), centered at about 5°N. This upper level jet is an integral part of the Asian monsoon system (Cook, 1999; Fig. 2a of Chen, 2005). Another, but weaker low level easterly jet core in ERAIM (Fig. 2.4a) is located at 900 hPa between 27° and 30°N, with core speed of about 3 m s\(^{-1}\). Several studies indicate that this easterly flow is associated with the northern track of African easterly waves, which are less significant because they are drier and shallower (Chen, 2006) and, in addition, do not have any link to precipitation (Diedhiou et al., 2001) and contribute little to tropical cyclogenesis (Thorncroft and Hodges, 2001; Chen, 2006). The AEJ at the mid-tropospheric level has been a focus of West African climate research for several decades because of its importance to the WAM rainfall. Burpee (1972) finds that occurrence of the AEJ is a response to the surface baroclinic zone and the reversal of the temperature gradients in the middle troposphere. Cook (1999) shows that soil moisture in association with positive meridional temperature gradients plays a vital role in AEJ characteristics.

As can be seen in Figure 2.4b, the model simulates four easterly jets in contrast to ERAIM’s three. The low level easterly jet core is located at about 900 hPa, with core speed of 10 m/s between 26° and 28°N. The location of the jet is well simulated, except that CEU overestimates the magnitude by 7 m s\(^{-1}\) compared to ERAIM. At the mid-tropospheric level between 550 and 770 hPa, two jets are simulated one in the Northern Hemisphere (NH) and the other in the Southern Hemisphere (SH). The model locates the AEJ of the NH at 20°N, 600 hPa, with core speed of about 8 m s\(^{-1}\), consistent with earlier observational and modeling studies (Cook, 1999; Afiesimama, 2007). In contrast to ERAIM, CEU places the AEJ to the north of the upward extension of the monsoon. The southern hemispheric AEJ (AEJ-S; as defined by Nicholson and Grist, 2003) is located at about 700–770 hPa, with core speed of 6 m s\(^{-1}\) between 0° and 5°S. This jet was first observed by Burpee (1972). Zhang et al. (2006) observed similar feature in Luanda using sounding data. This jet is however not obvious in the reanalysis. Finally, the model simulates a stronger TEJ with core speed twice as large as ERAIM.
reanalysis in the upper levels. Comparisons of the locations of the jets with Figure 2.2 indicate that the AEJ and TEJ lie approximately north and south of the rainbelt.

The vertical cross section of the mean JAS vertical wind (omega, mb s⁻¹) also appears in Figure 2.4. Two contrasting areas of strong rising and descending motions north and south of the equator, respectively appear in both ERAIM (Fig. 2.4a) and CEU (Fig. 2.4b). The greatest contrasts are the width, strength and location of the rising motions. A prominent feature is the narrow column of rising motion in ERAIM located between 3° and 10°N, while in CEU, the broad rising motion lies between 8° and 17°N. The core of the ascending air of about 4 m s⁻¹ at 9°N in ERAIM is located at 600 hPa consistent with the core of the mid-tropospheric AEJ. Similarly, in CEU, the maximum vertical motion of 5.5 m s⁻¹ is at 13°N, 600 hPa. This column of rising motion is bounded in the north and south by the axes of AEJ and TEJ respectively. Comparison with Figure 2.2 shows that the West African rainbelt is associated with the column of the ascending motion lying between the axes of the AEJ and TEJ. Two shallower columns of rising air near the surface also appear. The first is found at 5°N in ERAIM, and at 7°N in CEU between the surface and 850 hPa. This rising air merges with the ascent found between the AEJ and TEJ. Nicholson (2009) suggests that it corresponds to frictionally induced ascent at the interface of the Gulf of Guinea and the coast of West Africa. The second shallower column of rising motion lies between 17° and 25°N in ERAIM, and between 22° and 24°N in CEU. Studies have shown that this rising air is associated with dry convection (Sultan et al., 2003a; Nicholson, 2009). Interestingly these columns of rising air are separated by the axis of the AEJ. Another prominent feature is the subsiding motion with core speed (8 mb s⁻¹) located at about 27°N in CEU, consistent with earlier observational studies (e.g., Sultan et al., 2003a).

Overall, CEU reproduces spatial patterns of the zonally extended area of summer rainfall and associated summer features of the atmospheric circulation over West Africa. It reproduces the rainband associated with the ITCZ, the heat low, the low level equatorial westerlies, the oceanic
westerly jet, the monsoon circulation, the NH and SH AEJ, TEJ and the vertical motions. The rainbelt is associated with the core of the rising motion, and embedded between the axes of AEJ and TEJ, suggesting a link between West African rainfall and the AEJ, TEJ, and vertical motion. The model result is consistent with observational studies, which have shown that West African rainfall is associated with AEJ, TEJ, and convection respectively (Janicot et al., 1996; Diedhiou et al., 1999). The model shows considerable discrepancies in terms of the location, magnitude and number of the jets (westerly and easterlies), and the monsoon with reference to ERAIM. For example, the NH AEJ is underestimated by about 4 m s\(^{-1}\) and more northerly by about 5° latitude, and the strength of the monsoon is overestimated by about 4 m s\(^{-1}\). We suggest that the intensification of the monsoon is a response to the weakened and northward displacement of the AEJ associated with the strong westerly jet over the east Atlantic bothering the coast of West Africa. There is no significant change in the latitudinal position of the TEJ (Fig. 2.4a,b), though it is stronger in the model.

3.2 Intraseasonal variability of West African climate

3.2.1. Intraseasonal variability of the mean rainfall fields

In this section, we examine the time-latitude cross sections, from 0° to 30°N, of annual cycle of rainfall averaged between 10°W and 10°E for TRMM, GPCP and CEU-SG datasets for the period of simulations, 1998 – 2007. Figure 2.5 shows the time-latitude cross section of the mean annual rainfall superimposed with mean 925 hPa horizontal wind field and the zero isoline of the zonal component of wind speed. The vectors clearly depict the moist-laden southwesterly trade winds (monsoonal winds) from the Guinea Gulf and the dust-laden northeasterly trade winds from the Sahara desert. The boundary between these two air masses of different origins is distinctly delineated by the zero isoline of the zonal wind component.
As shown in Figure 2.5, three cells of rainfall maxima associated with the latitudinal location of the ITD and ITCZ are evident in the annual cycle. The first is in May to early July in TRMM (Fig. 2.5a), early May to late June in GPCP (Fig. 2.5b) and early March to late April in CEU (Fig. 2.5c). The second is in July to September in both TRMM and GPCP (Fig. 2.5a, b) and early June to August in CEU (Fig. 2.5c). The third appears only in CEU and occurs in early October to November. The coastal regions, south of 8°N are characterized by two rainy season regimes, while regions north of 12°N are characterized by a single rainy season regime (see also Table. 2.1). Some studies have shown that the evolution of West African monsoon rainfall is characterized by three distinct phases (Le Barbé et al., 2002; Sultan and Janicot, 2003b; Zhang et al., 2006). Consistent with these studies, the features displayed in Fig. 5 can be classified into onset, intensification and cessation phases respectively, where the cessation in CEU does not occur as early as it does in the TRMM and GPCP results.

The onset phase, from both GPCP and TRMM, is from March to early June (Fig. 2.5a, b), while in CEU-SG, it is from early February/March to early May (Fig. 2.5c). The onset phase is characterized by rain band intensification along the Guinean coast. The ITCZ is located between the equator and 8°N, and the isoline of 4 mm day^{-1} is located below 10°N. During July to September in both TRMM and GPCP, and June to August in CEU, rainfall maxima which mark the second phase are evident. During this period, an abrupt northward shift (referred to as 'monsoon surge') in the rainbelt into the Sahelian region occurs, consistent with the northernmost position of the ITD/ITCZ. The rainbelt at this time is located between 9° and 12°N in both TRMM and GPCP. The CEU captures the monsoon jump and the rainbelt is well simulated, but it is located farther north between 10° and 14°N. The isoline of 4 mm day^{-1} is now at its peak at about 16°N in TRMM and GPCP, but farther north at about approximately 20°N in CEU. This transition in rainfall marked the beginning of the enhanced rainy season in the Sahel (Camberlin and Diop, 2003), with evidence of sudden termination of heavy rainfall along the Guinean coast. Previous studies such as Adedokun, (1978) and Omotosho (1988)
have shown that the reduced rainfall along the Guinean coast region, associated to the so-called little dry season from mid-July to mid-September, is not connected to lack of moisture in the atmosphere. In fact, during this period, monsoon is fully developed (Sultan and Janicot, 2003b), which suggests that there is enough moisture in the atmosphere (see Fig. 2.4). Omotosho (1988) suggests that the lower rainfall is due to the stronger subsidence associated with outflows from deep convective systems located to the north of the area. The second phase of the WAM rainfall is followed by the cessation phase beginning from September/October (Le Barbé et al., 2002; Camberlin and Diop, 2003) to late February. During this period, rainfall decreases and gradually retreats southwards to its southernmost position at the coast during boreal winter, consistent with the southernmost position of the ITD/ITCZ. During this last phase, the isoline of 4 mm day\(^{-1}\) is located south of 8°N, and the ITCZ lies between the equator and 8°N. The three phases are closely linked to (i) the location and northward movement of the ITD and ITCZ with a time lag following the solar zenith position, (ii) the peak in mesoscale convection and (iii) the southward retreat of the ITD and ITCZ following the solar zenith position.

For better assessment of the model in simulating the intraseasonal variability of the mean rainfall fields, we examine the spatial distribution of rainfall over West Africa, in association with the AEJ and TEJ, at monthly scale during the summer season, since these periods are when the monsoon is at its peak and extends over a wider area. The spatial distribution of rainfall over West Africa superimposed with mean wind vectors and the zero isoline of the zonal wind component appears in Figure 2.6, for TRMM and CEU for July (Figs. 2.6a,d), August (Figs. 2.6b,e), and September (Figs. 2.6c,f) for 1998 to 2007. We used TRMM precipitation because of its higher resolution compared to GPCP. As discussed earlier, the red line, which represents the zero isoline of the zonal wind component, outlines the boundary between the monsoon winds and the harmattan winds and shows the northernmost limit of the ITD. Three main features are obvious in this distribution: the intense rainfall associated with the ITCZ in a narrow latitudinal band which exhibits a northward and
southward excursion during the period, the pockets of maximum rainfall over orographic zones, and
the minimum rainfall regions immediately north and south of the intense narrow rainband. Rainfall
values ranges between 1 mm day$^{-1}$ and 15 mm day$^{-1}$, and the shaded areas in these figures are where
rainfall exceeds 4 mm day$^{-1}$.

In June (not shown), the rainbelt associated with the ITCZ in TRMM is located at about 6°N over
the Atlantic Ocean and extends westward over land covering the coastal zones of West Africa.
Rainfall maxima (in excess of 12 mm day$^{-1}$) are located along the Guinean coast and the orographic
regions. The isoline of 4 mm day$^{-1}$ over land is located at about 12°N, and the zero isoline of zonal
wind component is at 19°N. The model places the rainbelt (about 8 mm day$^{-1}$) at about 12°N, with the
isoline of 4 mm day$^{-1}$ over land located at about 18°N, and the zero isoline of zonal wind component
at 20°N. The model simulates lower rainfall over the coastal and orographic regions. In July, the
rainband moves northward in both the TRMM (Fig. 2.6a) and CEU (Fig. 2.6d), extending over the
Sahel. In TRMM, rainfall maxima located between 4° and 10°N are associated with highlands in the
Jos Plateau, eastern Nigeria/northwestern Cameroon and central Guinea. During this month, the ITCZ
has shifted to around 10°N, the isoline of 4 mm day$^{-1}$ over land is located around 14°N, and the zero
isoline of zonal wind component is at 21°N. The CEU however, places the ITCZ at 14°N, the isoline
of 4 mm day$^{-1}$ over land is located around 20°N, and the zero isoline of zonal wind component is at
19°N. This leaves the Gulf of Guinea, coastal zones and the orographic regions relatively drier in
CEU compared to TRMM. In August, the ITCZ remains quasi-stationary in both TRMM (Fig. 2.6b)
and CEU (Fig. 2.6e) at around 10°N and 14°N. Rainfall intensifies over land, and the coastal zones,
with the exception of the highlands, experiences lower rainfall. Both the isolines of 4 mm day$^{-1}$ and
zero zonal wind remain quasi-stable in TRMM. In CEU, a slight southward displacement of both the
isoline of 4 mm day$^{-1}$ and the isoline of zero zonal wind components in response to the strengthened
northeasterly trade winds is seen. In September, the rainband shifts southwards leaving the northern
regions relatively drier, as seen in both TRMM (Fig. 2.6c) and CEU (Fig. 2.6f). The ITCZ is now
located around 6°N and 8°N in TRMM and CEU. During this period, the coastal regions again experience intense rainfall, while the Sahel is dominated by stronger harmattan winds especially in the model. The seasonal excursion of the rainband is consistent with the meridional migration of the ITD/ITCZ, a response to the seasonal migration of the zolar maximum. The model (Figs. 2.6d-f) captures the seasonal evolution of the West African rainfall.

3.2.2. Mean winds at 600 and 200 hPa

The mean wind field at 600 hPa (Figs. 2.7a-h) over West Africa shows the location and intensity of the AEJ in June-September in ERAIM and CEU. In the ERAIM, the AEJ, in June (Fig. 2.7a), is located in a broad zonal band between 5°–17°N with core speed of 14 m s\(^{-1}\) between 10° and 13°N. In July, the core of the AEJ is located between 14° and 18°N with maximum core speed in excess of 12 m s\(^{-1}\) (Fig. 2.7b). Figure 6c shows that the jet tilts slightly northward in a northwest to southeast direction while covering a wider latitudinal band in August. The core speed of the jet located at approximately 17°N is similar in magnitude to the observed core speed in July. However in September (Fig. 2.7d), which marks the southward retreat of monsoon system and the subsequent cessation of rainfall in West Africa (Fig. 2.5), AEJ, would be seen migrating southwards. The jet core is found at about 15°N with speed exceeding 12 m s\(^{-1}\), consistent with previous months. Similarly in October (not shown), the jet core of about 8 m s\(^{-1}\) located in a narrow zonal band is found at a mean latitudinal position of about 10°N.

The seasonal evolution of AEJ latitudinal position however suggests that AEJ weakens as it migrates northwards. During the summer season, from June, when the AEJ is at its southernmost mean latitudinal position, to August, at its northernmost latitudinal position, AEJ moves northwards at a regular latitudinal interval of about 2° per month. However, from August to October, AEJ exhibits an irregular pattern with a sharp drop in latitudinal position as it shift southwards. A number
of studies have suggested that the seasonal march of the AEJ is similar in all respects to the meridional movement of the ITD/ITCZ during summer.

The model, though, simulates a somewhat weaker and more northerly AEJ (Figs. 2.7e-h), but captures fairly well the seasonal cycle of the jet. In comparison to ERAIM, the AEJ core has speeds of about 12 m s$^{-1}$ at 15°N (June), 10 m s$^{-1}$ at 18°N (July), 10 m s$^{-1}$ at 20°N (August), and 7 m s$^{-1}$ at 20°N (September). In October (not shown), the jet is located at 13°N with core speed of about 8 m s$^{-1}$. The simulation of the jet approximately 3° north of its location in ERAIM could explain why the model simulates rainfall further north than TRMM. In addition, since the position of AEJ is linked to seasonal rainfall variability, it suggests that part of the underestimate of rainfall over Guinea coast may be due to the northward extent of the jet.

The mean wind field at 200 hPa (Figs. 2.8a-h) over West Africa features the seasonal cycle of the magnitude and location of the TEJ for four summer months of June, July, August and September for ERAIM and CEU. In ERAIM (Figs. 2.8a-d), the TEJ peaks around 9 m s$^{-1}$ at 5°N in June, around 16 m s$^{-1}$ at 7°N in July, also around 16 m s$^{-1}$ at 6°N in August, while in September, it is located at 5°N with core speeds in excess of about 12 m s$^{-1}$. The TEJ attains its maximum intensity in July/August and is at its northernmost position in July. In August, the TEJ begins its southward migration as seen in Fig. 2.8c, and reaches the southernmost position in September (Fig. 2.8d) during the period of this study. The jet weakens as it moves southwards.

In comparison to ERAIM, CEU overestimates the core speed and locates the jet too far northward (Figs. 2.8e-h). CEU produces core speeds of 10 m s$^{-1}$ at 8°N (June), 20 m s$^{-1}$ at 9°N (July), 22 m s$^{-1}$ at 8°N (August), and 21 m s$^{-1}$ at 5°N (September). The existence of this jet suggests that there is a layer of warm air to the north and cool air to the south of the jet. Studies have shown that the jet is being driven by the differential heating and cooling between the Tibetan highlands and the Indian Ocean (Nicholson and Grist, 2003). The difference in latitudinal position of about 3° north in CEU versus
ERAIM could have a significant impact on the simulated rainfall over West Africa, as previous studies have suggested that TEJ plays important roles in West African rainfall (Jenkins et al., 2005; Nicholson, 2009). The latitudinal variation of these jets and their relative roles in convection over West Africa will be discussed in the following section.

3.3. Regional climatological study

A more detailed assessment of the model in simulating the regional precipitation and circulations is carried out by comparing results over specific regions with observation. The simulated and observed precipitation and associated circulations averaged over 10°W to 15°E for two selected regions, including the entire West Africa is presented in the following sections.

3.3.1. Precipitation

West Africa is classified into distinct climatic regions based primarily on mean annual rainfall. Nicholson and associates in series of studies over West Africa have identified five climatic regions: Sahelo-Sahara, Sahel, Soudan, Soudano-Guinean and Guinea coast (e.g., Nicholson et al., 2000). However, in order to classify and choose the appropriate climatic regions for this study, we examine the mean annual rainfall for every 2° of latitude over the interior of West Africa, 5°W to 5°E. The percentage contribution of individual months to the mean annual rainfall for every 2° latitudinal interval appears in Table 2.1. The regions from 4° to 8°N (the Guinea Coast) experience double maxima in the annual cycle, with the primary maximum in May and the secondary maximum in October/November. Rainfall seems to be less variable in the regions 8° to 12°N (the Soudano-Guinean) during summer. This region equally experience two wet seasons, with the first maximum rainfall being June or July, while the secondary maximum occur in September or October. The regions from 12° to 20°N (the Sahelo-Sahara, Sahel and Soudan) is characterized with a single wet season, with August being the wettest. The percentage contribution of August rainfall to the annual
mean rainfall is 25%–43%, while July and September equally contribute about 22% to 26% of the annual mean rainfall. These regions receive most of their precipitation in summer months of June to September. There are some similarities in the rainfall climatology for the regions lying approximately between 12° and 18°N (i.e. Soudan, Sahel, and Sahelo-Sahara), hence, we have in this study combined them (hereafter; Soudano-Sahel). So, contrary to Nicholson’s et al. (2000) regional classification, we tend to have, for the purposes of this study, three climatic regions: Soudano-Sahel, Soudano-Guinean and Guinea coast. The results for only the two extreme regions, Soudano-Sahel and Guinea coast, will be presented.

In order to better assess the capability of the model in simulating regional climate, we examine the intraseasonal variability of rainfall for only Guinean coast and Soudano-Sahel. For comparison, we have also included the entire West Africa. It is clear that the regions are dominated by different atmospheric mechanisms. Thus, alongside the rainfall study, we have also examined the intraseasonal variability of the associated large-scale circulations.

Figure 2.9a shows the time series of rainfall over West Africa (5°–20°N, 10°W–15°E) in the TRMM, GPCP and CEU datasets. The figure features an inverted V-shaped rainfall pattern, with peak in August greater than 5 mm day\(^{-1}\) in TRMM, GPCP and CEU datasets. The model shows a tendency to overestimate rainfall before and after the single peak. Slight disparities also exist in both observations, with GPCP showing tendency to be greater throughout the season than TRMM.

The Guinean coast (5°–8°N; 10°W–15°E, Fig. 2.9b) features a double maximum rainfall pattern, the first in June (May) and the second in September (October) in observations (CEU), with a dip in rainfall in August in both observations and CEU. Several authors have suggested that the minimum in August, often termed the “Little Dry Season” of coastal West Africa, is related to the static stability that exists during this period over the coastal areas, which prevents the development of convection (Adedokun, 1978). In essence, this “dry season” does not occur for the same reason as the winter dry
season. In fact, substantial evidence in the literature suggests that the atmosphere during July–August in the coastal areas is moist. Recall that during this period the monsoon layer has become very deep (Fig. 2.4) reaching up to about 800 hPa (Omotosho et al., 2000) with the coastal region rich in atmospheric moisture. The dry season rainfall in the region occurs in November–February with rainfall on the order of about 2 mm month\(^{-1}\). The model simulates the seasonal rainfall pattern, but it underestimates the June, July, August and September precipitation, while it overestimates both the onset and cessation of rainfall.

Rainfall pattern in Soudano-Sahel, 12°–18°N 10°W–15°E (Fig. 2.9c) features a more pointed inverted V-shaped single mode distribution with the maximum rainfall in August. While the dry season rainfall of November–February, and the onset period are well simulated by CEU in comparison to observations, the main rainy season period is conspicuously overestimated especially in July–September. The difference in the Sahelian rainfall as seen in observations and simulation may be due to the strength of the simulated westerly jet discussed earlier. In general, the West African rainfall characteristics suggests that the summer rainfall is dominated by the Soudano-Sahel rainfall dynamics, while the onset and cessation periods are mainly controlled by the Guinean coast rainfall dynamics.

### 3.3.2. Temperature

The seasonal evolution of the mean surface air temperature averaged over West Africa, Guinean coast, and central Soudano-Sahel appears in Fig. 2.10. In the ERAIM, the seasonal evolution is characterized by double peaks and a single minimum (Fig. 10a). The primary temperature maximum occurs in May, while the secondary maximum occurs in October, consistent with both the monsoon onset and retreat. The peak of the monsoon system (Fig. 2.4) is associated with the temperature minimum in August. The observed seasonal cycle is fairly well simulated, although CEU underestimates both the temperature maxima and minimum. This cool bias is consistent with
excessive rainfall (Fig. 2.9). Nicholson and Grist (2003) suggest that the effect of the rainfall is to cool the surface, which results in lower surface temperatures.

Figure 2.10b features the Guinean coast with the observed double maximum, but with primary peak in March and secondary peak in November-December. The CEU captures the primary peak, after which its temperature drops off gradually to become approximately constant at about 24.5°C. Generally, the model tends to exhibit a cool bias except during summer season when a slight warm bias occurs. The central Soudano-Sahel (Fig. 2.10c) portrays a temperature pattern similar to the entire West Africa, both in the reanalysis and CEU simulation. The seasonal pattern over this region is well simulated. There is however, a shift towards higher values in comparison to entire West Africa.

3.3.3. Wind

We examine here the relationships between the seasonal cycle of the rainbelt and the Northern Hemispheric easterly jets, AEJ and TEJ. Figures 11a and 11b show the mean latitudinal locations of the rainbelt from GPCP and CEU and the AEJ and TEJ from ERAIM and CEU, respectively. Figs. 2.11a,b show that the structures of the rainbelt, the AEJ and the TEJ are similar between the ERAIM and the model, though the model places them too far north.

During the course of the year, Fig. 2.11 shows a pronounced meridional migration of the rainbelt. Beginning in January at its position between latitudes 2° and 4°N, the rainbelt migrates northward, reaching its northernmost extent in August, between latitudes 12° and 14°N. Immediately after August, the rainbelt retreat southwards in September reaching its southernmost position in November/December. Studies have shown that the seasonal cycle of the rainbelt is consistent with the latitudinal variation of the surface position of the ITD, the meeting point between the hot, dry north-easterly winds and the moist south-westerly winds (cf. Figs. 2.5 and 2.6). The ITD attains its most
southerly position in January between latitudes 2° and 4°N, while it attains its most northerly position between latitudes 20° and 25°N in August (Adedokun, 1978). The difference in the latitudinal position of the rainbelt and the ITD confirms that the rainbelt is embedded within the convection zone (see Fig 2.4) south of the ITD surface position, consistent with the ITCZ.

The seasonal shift in the rainbelt is consistent with the seasonal migration of the AEJ. The AEJ lies approximately north of the rainbelt during the course of the annual cycle except in February (when the rainbelt can be seen slightly north of the jet core) and November (Figs. 2.11a,b). The jet core is about 4°–7°N north of the rainbelt in May – October, while it is about 1°–3°N in all other months. The similarity in the seasonal cycle suggests that a strong relationship exists between the rainbelt and the AEJ. Since the jet equally exhibits similar annual pattern as the ITD, as discussed earlier, there is an indication that a close association exists between the two. It can therefore be inferred from the figures that maximum convective instability occurs south of the AEJ.

In contrast to the rainbelt and the AEJ, both ERAIM and CEU (Figs. 2.11a,b) show the existence of the TEJ in the SH during the first and last quarters of the year. The TEJ begins its northward excursion to the Northern Hemisphere during late March and crosses the equator in early April. It continues its northward advancement until July, when it attains its most northerly position. Between late July and early August, the TEJ begins to move southward and crosses the equator again in October back to the SH. It is interesting to note that the TEJ exists in the NH over West Africa for only six months. During this period, the rainbelt lies approximately between the AEJ and TEJ cores. As simulated by the model, during the annual cycle, the TEJ is stronger and more northward but otherwise well simulated. Since the TEJ appears to be absent in West Africa during the onset of rainfall over coastal zones, but persistent during the peak rainy season over Sahel, then a question comes to the mind, what is the role of TEJ in West African rainfall?
Until recently, the TEJ has been considered a passive system. However, the similarity between its annual cycle and ITD in summer suggests, there is a relationship between the two, which implies that TEJ may contribute to rainfall over West Africa. TEJ attains its northernmost position consistent with the period of maximum rainfall in the Soudano-Sahel and little dry season in the Guinea coast. The implication of this as can be inferred from Figs. 11a,b is that, the only influence of the TEJ is to affect the intensity of rainfall, especially in the Sahel when in phase with the AEJ. However, Nicholson and Grist (2003) suggests that the variability of TEJ over the Guinean coast is a response to rainfall variability. Fig. 2.11 further suggests that a close link exists between AEJ and TEJ confirming the work of Nicholson et al. (2007). The authors, in a study of wave activity on the TEJ suggest that the interaction between the AEJ and TEJ promotes the development of wave disturbances over West Africa. The variability of these disturbances has however been known to modulate rainfall. Hence, the TEJ may be a factor in the development of the rainy season over West Africa.

4. Conclusions

The intraseasonal variability of West African rainfall and associated large-scale dynamics has been described in detail from the viewpoint of observations and simulation. We use a coupled atmospheric global climate model, CAM-EULAG (CEU), constructed using EULAG dynamics (Prusa et al., 2008) and the CAM physics (Collins et al., 2004, 2006) to examine the intraseasonal variability of West African climate. We compare the CEU rainfall with both TRMM and GPCP datasets, and the atmospheric circulations are compared with ERAIM. The focus of this study is to access the performance evaluation of CEU to simulate the spatiotemporal variability of rainfall and associated atmospheric features. We run CEU with grid stretching centered over West Africa. The analysis of the 10-year simulation (1998-2007) show that the West African summer mean climate variability compared well with observations.
Features of the CEU rainfall for West Africa are realistically simulated. The model simulates the mean rainfall variability at the peak of the West African summer rainy season (July–September), and it captures the rainbelt associated with the ITCZ, though it locates it farther north compared to TRMM and GPCP. In addition, CEU reproduces rainfall over highland regions, but tends to underestimate the rainfall maxima over the Cameroon Mountains and the Jos Plateau. The intraseasonal variability of rainfall averaged over 10°W to 10°E is equally well represented (Fig. 4). In agreement with earlier studies (Sultan and Janicot, 2003b; Zhang et al., 2006), and consistent with TRMM and GPCP, the model captures the three distinct phases (onset, intensification and cessation phases) of the West African monsoon rainfall. The onset phase is simulated earlier, starting from early February/March to early May, compared to observations. The CEU captures the monsoon jump during the intensification phase, but moves it more northward, between 10° and 14°N, as compared to 9° and 12°N in both TRMM and GPCP. The isoline of 4 mm day\(^{-1}\) during this period is located at about 16°N in TRMM and GPCP, but at about 20°N in CEU. The monthly evolution of rainfall exhibits meridional migration of the rainbelt reaching the northernmost position in August over Sahel, consistent with the northernmost limit of the ITCZ (Fig. 5).

The spatiotemporal variability of the accompanying physical mechanisms responsible for the rainfall variability are well simulated. The model captures the mean summer temperature. The temperature increases from the Guinean coast to the Saharan region. The coastal region is associated with low temperatures linked to the cool, moist southwesterly winds from the Gulf of Guinea, while the northern zones are associated with high temperatures consistent with the hot, dry harmattan winds from the Saharan desert. Since there is a close relation between rainfall and temperature, the model shows a tendency to underestimate temperature over high-grounds as expected where the rainfall is equally underestimated.
The simulated vertical structure of zonal wind component compares very well with ERAIM. The weaker and more northward-displaced AEJ lies north of the core of the rainbelt in the region of the maximum rising motions, with the cores of the rainbelt and the ascent underlain by the strengthening TEJ at 200 hPa level. Consistent with previous studies (e.g., Sylla et al., 2010), the monthly migration of the rainbelt previously discussed is associated with the northward migration and weakening of the AEJ and the appearance and strengthening of the TEJ at 10°N. Similarly, in agreement with earlier studies (e.g., Grist and Nicholson, 2001; Nicholson and Grist, 2001; Patricola and Cook, 2010b), the variability in the AEJ and TEJ has been found to play an important role in the variability of rainfall over West Africa. In this study, the core of the rainbelt as stated earlier appears between the axes of the two jets, suggesting that the instability of the AEJ may lead to rainfall variability over West Africa. The model simulates a broader and stronger rising motion with core extending from the surface and covering the entire troposphere. The core of the West African rainbelt corresponds to the column of the ascending motion, and both are associated with the surface ITCZ. The rainbelt lies between the axes of AEJ and TEJ. The coincidence of the core of the rainbelt and the core of the deep ascent suggests that the vertical motion plays a significant role in rainfall.

We further evaluate the performance of the model by examining the sub-regional simulations of rainfall, temperature, and wind circulations. The analysis of the Guinean coast, Soudano-Sahel, and the entire West Africa climate shows that the model tends to underestimate (overestimate) rainfall in Guinean coast (Soudano-Sahel) during the peak monsoon season. Consistent with rainfall, the model underestimates temperature along the Guinean coast during summer, while there was a warm bias in every other month. Contrary to the Guinean coast, the model shows a warm bias in every month in Soudano-Sahel. The mean latitudinal position of the rainbelt, the AEJ, and the TEJ is reasonably well represented.
Overall, the GCM simulations capture several of the observed relationships between rainfall and associated atmospheric circulations, though there are some notable discrepancies in the simulations. Most prominent are the lower rainfall over orographic regions consistent with a cold bias in temperature, overestimation of the depth and strength of the monsoon associated with overly strong tropical, low-level westerly flow and a stronger African westerly jet, accompanied by a weaker AEJ on the northern side of the monsoon and a stronger TEJ in the upper troposphere, and stronger ascent between the cyclonic side of the AEJ and the anticyclonic side of the TEJ.

The findings presented in this study are consistent with previous modeling and observational studies performed for West Africa. The limitations encountered are not limited to our model; they are consistent with other regional and global models. There are several causes that have been cited in literature that might be the potential reasons for the model disparities. While we have carefully assessed CEU and found it to produce reasonable simulation of West African rainfall and associated large-scale dynamics, we must also note that there are notable discrepancies in the simulations, and we therefore encourage further modifications to the model physics with particular emphasis on the convection schemes. We suggest that the model as it is can be used for further studies of climate variability over West Africa, and any other regions in Africa because of its stretched-grid capability.

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Figure Captions

Figure 2.1 Horizontal grid resolution for CEU-SG simulation. Panel (a) shows the zonal distribution for the grid points along 18°N (line). Panel (b) shows the double nested distribution used for the meridional grid points. Panel (c) shows the combined effect of meridional and zonal mappings for grid. Only 20% of the grid points are shown for clarity.

Figure 2.2 Mean JAS rainfall (mm day$^{-1}$, shaded) from (a) TRMM, (b) GPCP and (c) CEU for 1998-2007. The mean JAS ERAIM surface temperature (in °C) is superimposed on both TRMM and GPCP, with CEU surface temperature (in °C) shown in panel (c) in contours.

Figure 2.3 The spatial distribution of mean JAS zonal wind at 850 mb for (a) ERAIM and (b) CEU for 1998-2007. The black contour line is the isoline of zero zonal wind component, and zonal winds ranging from -2<u<2, are shaded white to distinguish the westerlies from the easterlies. The mean JAS wind field at 850 mb represented in vectors is superimposed. The vector scale in m s$^{-1}$ is shown at the bottom of the panel. The blue vectors show regions with wind speed greater than 10 m s$^{-1}$, while the black vectors show regions with wind speed less than 10 m s$^{-1}$. White patch areas at the top right hand corner on the right panels are missing data.

Figure 2.4 The vertical cross section of mean JAS zonal wind component (m s$^{-1}$, contour), the vertical wind (mb s$^{-1}$, shaded), and vector wind (v, w) averaged between 10°W and 10°E for (a) ERA-Interim and (b) CEU for the period 1998-2007. The scale of the vector is shown below the panel.

Figure 2.5 The time-latitude cross-section of mean rainfall averaged between 10°W and 10°E for (a) TRMM, (b) GPCP, and (c) CEU for 1998–2007. Rainfall values greater than 4 mm day$^{-1}$ are shaded, and white areas are values less than 4 m s$^{-1}$. The ERAIM 925 hPa wind field
is superimposed on the TRMM and GPCP, and is expressed in vectors. The model’s 925 hPa wind field represented in vectors is superimposed on the mean rainfall. The vector scale in $\text{m s}^{-1}$ is shown at the bottom of the panels. The black line represents the zero isoline of the 925 hPa zonal wind component.

**Figure 2.6** Mean monthly rainfall fields (mm day$^{-1}$) for TRMM (left panel) and CEU (right panel) for (a,d) July, (b,e) August, and (c,f) September for 1987–2007. Rainfall values greater than 4 mm day$^{-1}$ are shaded. The ERAIM 925 hPa wind field is superimposed on the left panels, and is expressed in vector in $\text{m s}^{-1}$. The model’s 925 hPa wind field represented in vectors is superimposed on the mean rainfall on the right panels. The vector scale in $\text{m s}^{-1}$ is shown at the bottom of the panel. The blue vectors show regions with wind speed greater than 20 m s$^{-1}$, while the black vectors show regions with wind speed less than 20 m s$^{-1}$. White patch areas at the top right hand corner on the right panels are missing data. The red line represents the zero isoline of the 925 hPa zonal wind component.

**Figure 2.7** Monthly variability of mean zonal wind at 600 mb for ERA-Interim (a – d) and CEU-SG (e – h). Only easterly speeds are plotted and shaded are areas where zonal wind greater than 6 m s$^{-1}$.

**Figure 2.8** Monthly variability of mean zonal wind at 200 mb for ERA-Interim (a – d) and CEU-SG (e – h). Only easterly speeds are plotted and shaded are areas where zonal wind greater than 9 m s$^{-1}$.

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Table 2.1. The percentage contribution of individual months to the mean annual rainfall for every 2°N latitudinal interval

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CHAPTER 3. West African extreme daily precipitation in observations and stretched-grid simulations by CAM-EULAG

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Abayomi A. Abatan and William J. Gutowski Jr.

Abstract

Extreme precipitation and its possible alteration by climate change can have substantial societal impact. We evaluate the ability of a global, stretched-grid model CAM-EULAG (CEU) to simulate extreme precipitation and its physical causes. CEU combines the EULAG dynamics core, developed by Smolarkiewicz and colleagues, with the physics package of the NCAR Community Atmospheric Model (CAM). CEU can use grid stretching to focus high resolution in selected regions. Here we analyze observed and simulated extreme daily precipitation and its underlying processes in observations and in a ten-year (1998-2007) CEU stretched-grid simulation, where the stretching gives 0.5° x 0.5° resolution over West Africa.

We focus on a core monsoonal region in West Africa: (6°-16°N, 5°W-5°E). In both the onset (April-May-June; AMJ) and mature-monsoon (July-August-September; JAS) seasons, the model reproduces well the observed climatological annual and diurnal cycles of precipitation in this region, though with somewhat greater than observed time-average precipitation. Daily precipitation extremes at the 99% level and higher are stronger in the observations, but the spatial scale of extreme events in the model is comparable to the observed scale. The model also simulates fairly well the interannual and intra-seasonal variability of the extreme events. The model’s anomaly vertical wind and humidity
fields on simulated extreme-event days correspond to ERA-Interim anomaly fields on observed extreme-event days. However, the three-hourly maximum precipitation on extreme-event days in the observations for both AMJ and JAS varies with extreme event, typically appearing during a period ranging from the late afternoon to early morning, in contrast to the late afternoon maximum in the overall, observed ten-year climatology. The model has difficulty replicating diurnal behavior of the extreme precipitation, tending to show a three-hourly maximum that appears only in the late afternoon. A chief reason for the difference appears to be the model’s inability to simulate squall lines, which produce much of the observed extreme precipitation.

1. Introduction

In the last 100 years, the global surface temperature has increased (IPCC, 2007). The anthropogenic greenhouse gas-induced warming can cause changes in other atmospheric variables and thus impact changes on temporal and spatial distribution of precipitation characteristics including extremes behaviour (Diffenbaugh et al., 2005). Changes in the occurrence of precipitation extremes leading to extreme events such as erosion and flooding can have tremendous impact on infrastructures, biological, agricultural and socio-economic systems (Nicholls and Alexander, 2007; Gutowski et al. 2010; Paeth et al. 2010).

In order to better understand the changes in extreme events in a warming climate and their physical causes, substantial effort has been devoted to research on extreme temperature and precipitation events in many regions of the U.S (Aguilar et al. 2005; Gutowski et al. 2007, 2008, 2010; Peterson, 2008), the Caribbean (Peterson, 2002), Mexico, Canada (Wang and Zhang, 2008), Europe (Christensen and Christensen, 2003), Asia (Griffiths et al. 2005), Australia, regions of Africa (Aguilar, 2009; New et al. 2006), and China (Chen et al. 2010; Wang and Zhou, 2005; Zhai and Pan, 2003) to mention but a few. These studies have shown that there are remarkable increases in intensity
of precipitation extremes. However, observational and modeling studies of extreme precipitation events in West Africa are limited.

There already exists an increase in the frequency of extreme precipitation events over most land areas in many parts of the world under the current global warming (Groisman et al. 2005; Christensen et al. 2007). The increase in extreme precipitation is consistent with increasing temperatures (Easterling et al. 2000; Trenberth et al. 2007) and atmospheric water vapor (Christensen et al. 2007). Studies have shown that there are large regional differences in precipitation variability. The increase in extreme precipitation has been most pronounced in middle and high latitudes, where total precipitation has increased. However, tropical regions are not excluded from the changes in extreme precipitation.

Previous studies on climate change and extremes using precipitation data have focused on changes on mean behavior. However, studies such as Alexander et al. (2006) have shown that analyzing changes in extremes requires daily data. Even at that, there have been only a few studies (Easterling et al. 2003; New et al. 2006) that have examined changes in African daily precipitation extremes. Extensive studies on extremes using daily data has been difficult in West Africa partly because of networks of observing stations with sufficient quality and quantity of data have not been available to the scientific community. This problem stems from a number of factors ranging from political instabilities to government bureaucracy on data policy issues, which, starting from the mid-1980s, plagued most of the countries in the region. More recently, the Working Group on Climate Change Detection and Indices, part of the joint World Meteorological Organization Commission for Climatology (CCI)/World Climate research Programme (WCRP) project on Climate Variability and Predictability (CLIVAR) organized regional climate change workshops to address some of the issues of data availability in many parts of the world including West Africa. One outcome of these workshops was, the result that extreme precipitation events had increased at some stations (for
example, parts of western Africa) but decreased at others (Easterling et al. 2003), though large areas remained unanalyzed. From all the international daily data sets available, Alexander et al. (2006) found that precipitation indices show a tendency toward wetter conditions throughout the 20th century. New et al. (2006) found that there has been statistically significant increase in regionally averaged daily precipitation intensity. However, none of these studies have attempted to study changes in daily precipitation extremes using numerical simulation. There is little that can be said about changes in climate variability or extreme events in West Africa (Christensen et al. 2007).

This study aims at using a climate model, CAM-EULAG (CEU), to examine extreme precipitation events over West Africa subregion where extensive modeling study is lacking (Jenkins et al. 2002). We also examine large scale atmospheric circulations responsible for the extremes. The goal here is to establish the capabilities of the stretch grid CEU to simulate extremes and their processes so that we can have a foundation for further research, such as climate change.

The paper is organized as follows. The observational datasets and climate model experiment, including methods of analysis, are discussed in Section 2. In Section 3, we examine the capability of the model in simulating West African precipitation with emphasis on precipitation characteristics, the knowledge of which we use to determine the widespread extreme events. Also, the statistical relationships between extreme precipitation events and large-scale atmospheric circulations are examined. Conclusions appear in Section 4.

2. Model description and observational dataset

This study uses an atmospheric global climate model, CAM-EULAG (hereafter, CEU) to study West African climate. CEU is constructed using EULAG dynamics (Prusa et al., 2008) and the physics package of the Community Atmospheric Model (CAM; Collins et al. 2004, 2006), developed at the National Center for Atmospheric Research (NCAR).
2.1 EULAG

EULAG is a non-hydrostatic, anelastic, parallel computational model for simulating all-scales geophysical flows (Smolarkiewicz and Prusa, 2005). It features a generalized coordinate formulation enabling grid adaptation (Smolarkiewicz and Prusa, 2005), non-oscillatory forward-in-time integration algorithms (Smolarkiewicz, 2006; Prusa et al. 2008), and robust elliptic solver (Smolarkiewicz et al. 1997). EULAG’s all-scale capability has permitted its successful application at scales ranging from centimeters (Andrejczuk et al. 2004) to $10^6$ km (Elliott and Smolarkiewicz, 2002). For example, EULAG has a proven record of being successfully applied in the areas of turbulence, gravity wave dynamics, urban flows, flows past complex/moving boundaries and micrometeorology (Smolarkiewicz and Prusa, 2002, Prusa et al. 1996; Prusa and Gutowski, 2006; Ortiz and Smolarkiewicz, 2006). Further details on EULAG can be found in Smolarkiewicz and Pudykiewicz (1992), Prusa et al. 1996, Prusa and Smolarkiewicz, 2003, Prusa and Gutowski, 2006, Abiodun et al. (2008), Prusa et al. (2008), and Abiodun et al. (2011).

2.2 CAM

The physics used in CEU come from CAM 3.1, which has been the atmospheric component of the NCAR Community Climate System Model (CCSM). The CAM physics include precipitation processes (e.g., moist deep convection, shallow convection and large-scale stable condensation), turbulence mixing, clouds and radiation, and ocean-surface and land models. Of relevance to this study, deep convection is based on a plume ensemble approach where updrafts and downdrafts are related to the interaction between cloud base mass flux and the convective available potential energy (CAPE) (Zhang and McFarlane, 1995; Collins et al., 2004). The CAM employs Hack (1994) mass flux scheme to represent shallow convection, and Sundqvist (1988) bulk microphysical parameterization scheme for evaporation of the convective precipitation, while the parameterization of non-convective clouds processes uses prognostic condensate scheme (Rasch and Kristjánsson,
1998) combined with a bulk microphysical parameterization (Zhang et al., 2003) closer to that used in smaller scale cloud resolving models. The cloud fraction in CAM is evaluated through a diagnostic method (Slingo, 1987; Hack et al., 1993; Kiehl et al., 1998). Collins et al. (2004, 2006) give details of CAM physics parameterizations.

2.3 CEU

Detailed description of the coupling of CAM and EULAG appears in Abiodun et al. (2008). We run CEU with a static stretched grid that concentrates grid points in the vicinity of West Africa (Fig. 3.1). The highest grid stretching centered over West Africa has latitude-longitude resolution of 0.5° X 0.5°, and extend zonally from about 0° – 40°N. The resolution elsewhere becomes as coarse as 4° (lat) and 2.8° (lon). The model uses 26 vertical levels, with model top at 30 km. Like other grid-point GCMs, the grid boxes of CEU tend to become elongated at the poles, a feature than can produce noisy behavior on small scales and contribute to numerical instability (Abiodun et al., 2008). To minimize such behavior, latitudes poleward of 75° have a sponge layer to damp the state of the atmosphere toward the zonal average. Abiodun et al. (2008) show that this damping has little influence on simulations at lower latitudes.

We ran model using observed sea surface temperatures and sea ice distributions obtained from the standard CAM package. The simulation ran from 1 January 1996 to 30 January 2008. We discarded the first two years for spin up and analyzed the 10-year period 1998-2007.

2.4 TRMM, GPCP, and ERA-Interim data sets

We used three sources of observation-based data to evaluate the simulation. For precipitation we used the gridded precipitation provided by Tropical Rainfall Measurement Mission (TRMM). Specifically, we used the multi-satellite 3B-42, version 6, product, which is on a 0.25° by 0.25° grid in a global belt extending from 50°S to 50°N. The TRMM 3B-42 product consists of rainfall
estimates derived from several satellite measurements including the TRMM Microwave Imager (TMI), Visible-Infrared Scanner (VIRS) and TRMM precipitation radar. Huffman et al. (2007) give the details of the data assimilation and their quality control. We used daily rainfall aggregated from 3-h values for 1998 to 2007, consistent with our simulation. In order to facilitate comparison of the simulation with observations, we regridded the rainfall data set to a 0.5° X 0.5° grid. Analysis of a subset of the data showed that regridding did not alter substantially the daily spatial pattern of precipitation seen on the 0.25° X 0.25° grid.

In addition, we compared simulated precipitation with a gridded precipitation data set from the Global Precipitation Climatology Project (GPCP), specifically the Global Daily Merged Precipitation analyses on a 1° X 1° grid (Huffman et al. 2001). GPCP precipitation estimates are derived from satellite sensors and rain gauge measurements.

We compared the simulated atmosphere with output from the ERA-Interim reanalysis (Berrisford et al. 2009; Dee et al. 2011). ERA-Interim reanalysis (ERAIM) datasets are available at a 1.5° X 1.5° resolution, with 37 pressure levels from 1000 to 1 hPa, for the period 1979 - 2010. The gridded data products available online include, among others, 6-hourly (0000, 0600, 1200, and 1800 UTC) upper-air fields covering the troposphere and stratosphere. For this study, we use data from 1998 to 2007.

2.4 Analysis methods

The analysis of extreme precipitation events in this study is similar to those of Bell et al. (2004) and Gutowski et al. (2007). We define a precipitation event as any day at a grid point that has precipitation exceeding 0.1 mm day$^{-1}$. We then pool all events from all the grid points in our target region in the core of the West African monsoon: (6°-16°N, 5°W-5°E).

In AMJ, we obtained 166,798, 52,589, and 278,860 daily precipitation events for TRMM, GPCP and CEU respectively, while in JAS we have 216,207, 66,405, and 323,559 daily precipitation events.
for TRMM, GPCP and CEU respectively. With these events, we produced frequency versus intensity histogram (with bin widths of 2.5 mm day\(^{-1}\)) that we normalized using the total number of precipitation events.

We define precipitation at a grid point as extreme when the amount exceeds the 99% level (Table 1). From these events, we then determine what we term widespread extremes by searching for multiple extreme events occurring on the same day. We designate days with extreme events occurring simultaneously on 12 or more grid points as widespread extreme events. We assume that these events cover sufficiently large area that their occurrence is at least partly governed by resolved behavior in the reanalysis or the model (as opposed to being only the product of a subgrid parameterization). We form composites of precipitation and atmospheric features on the days of widespread extreme events, April-May-June (AMJ) and for July-August-September (JAS).

3. Results

3.1. Model validation: Climatology

3.1.1 Mean precipitation

We give here a brief diagnosis of the model’s climatology as background for analysis of extreme precipitation. Further validation of CEU’s climatology for West Africa appears in Abiodun et al. (2011). Figure 3.2 shows the mean AMJ and JAS precipitation for TRMM, GPCP and CEU datasets. The model captures many of the large-scale features of seasonal precipitation. In AMJ, prominent precipitation maxima appear along the Guinean coast and in the Intertropical Convergence Zone (ITCZ) between 2°N and 6°N with maximum precipitation of about 12 mm day\(^{-1}\) in TRMM, and 10 mm day\(^{-1}\) in both GPCP and CEU. Immediately above the rainbelt, precipitation decreases northward as far as the southern fringe of the Sahelian region at about 16°N. TRMM observation (Fig. 3.2a) shows precipitation maxima (12 mm day\(^{-1}\)) over Cameroon highlands oriented northeast/southwest,
and extending westward to the Jos Plateau in Nigeria. The GPCP observations (Fig. 3.2b) show a similar localized precipitation maximum around the Cameroon Mountains but with lower maximum (10 mm day$^{-1}$) and a broader spatial extent. The CEU (Fig. 3.2c) simulates fine scale features in West African precipitation consistent with TRMM. The model reproduces the zonally extended structure of precipitation between 2°N and 6°N, and with a northeast-southwest tilt. In comparison to observations, CEU underestimates precipitation over Nigeria/Cameroon border and along the ITCZ.

JAS precipitation over West Africa (Figs. 3.2d–f) shows a northward shift in the rainbelt to within 7°N and 12°N, in contrast to the latitudinal location of the rainbelt in AMJ (Figs. 3.2a–c). This season is characterized by northward migration and intensification of the rainbelt, leaving the coastal areas with reduced precipitation. The northward shift in the zone of maximum precipitation is consistent with the northward migration of the location where meridional wind changes sign, the Intertropical Discontinuity (ITD). TRMM (Fig. 3.2d) places the rainbelt in a broader zonal band between 7°N and 13°N, with the core located at about 8°N. Precipitation increases to a maximum of about 17 mm day$^{-1}$, and extends westward into the Atlantic Ocean along the ITCZ. Another interesting feature of the mean precipitation over land in TRMM dataset is the local regions of maximum precipitation in the Guinea highlands, Cameroon Mountains and Jos Plateau. Precipitation over Cameroon Mountains is comparable to the maximum embedded within the ITCZ over the Atlantic Ocean. The JAS precipitation pattern over Cameroon Mountains is similar to AMJ precipitation pattern, but with greater extent. In the GPCP dataset, the core of the ITCZ, with precipitation maximum of about 12 mm day$^{-1}$ at 8°N is embedded within the zonal rain band located at 6°–12°N (Fig. 3.2e). A prominent precipitation maximum of about 12 mm day$^{-1}$ centered between 8°N and 12°N along 13°–17°W exist over the coastal region of Guinea and Guinea-Bissau. The second orographically related precipitation maximum of about 11 mm day$^{-1}$ found over Cameroon Mountains, extends northward encompassing Nigeria on the left and Lake Chad to the right.
The CEU JAS simulation (Fig. 3.2f) shows a slight difference in mean precipitation compared to TRMM (Fig. 3.2d) and GPCP (Fig. 3.2e). It captures the precipitation maximum over Guinea highland region to a reasonable extent in agreement with GPCP, but underestimates precipitation maxima over Cameroon Mountains and Jos Plateau. Similarly, region of reduced precipitation over land, 8°–13°W, 6°–16°N, can be seen. However, the fine scale features associated with the eastward extension of the rainbelt over 5°W–5°E, 6°–16°N are well captured by the model. Over the Atlantic Ocean in both seasons, the rain band associated with the ITCZ are well simulated, though CEU underestimates the westward extend of the ITCZ and the magnitude of the precipitation. Generally, precipitation decreases south and north of the maximum rainbelt. It is indicated from the figures that significant differences exist in TRMM and GPCP mean precipitation, which could be related to the higher resolution of the TRMM data. The discrepancies between observed data sets thus emphasize the uncertainties of precipitation validation over West Africa (Druyan et al., 2008). On the other hand, since the model captures the precipitation pattern, but not the intensity, then the discrepancies in the simulated precipitation may be due to the convection parameterization used in the model. Overall, the precipitation pattern is well captured by CEU.

3.1.2. Precipitation frequency and intensity

Previous studies have shown that climate models produce reasonable patterns of precipitation amount, but often overestimate the frequency of light precipitation and underestimate the intensity of heavy precipitation (Dai and Trenberth, 2004; Sun et al., 2006, 2007). Fig. 3.2 shows that CEU produces a reasonable spatial pattern of the mean precipitation. However, this does not guarantee that the model produces high intensity precipitation with the same frequency as observed (Dai et al., 1999; Sun et al., 2006).

Figure 3.3 shows the mean histograms of daily precipitation frequency as a function of daily precipitation intensity for 1998–2007 for AMJ and JAS in observations and in the model. The arrows
show the precipitation intensity corresponding to the 99th percentile. As shown in AMJ (JAS), the model overestimates the frequency of precipitation for precipitation less than 12.5 mm day\(^{-1}\) (17.5 mm day\(^{-1}\)), but underestimates the frequency for precipitation greater than 12.5 mm day\(^{-1}\) (17.5 mm day\(^{-1}\)), consistent with previous studies using climate models (Sun et al., 2007). This means that the model produces less intense precipitation than observations.

Table 3.1 shows the average daily precipitation rate (mm day\(^{-1}\)) for the 99th percentile for all daily precipitation events for the two seasons and for CEU and the observations. The 99th percentile of daily precipitation rate is about 3 – 11 mm day\(^{-1}\) smaller in AMJ than JAS. This could be because the region receives most of its annual precipitation during the peak summer season. The differences in 99th percentile precipitation between TRMM and GPCP ranges from 18 – 23 mm day\(^{-1}\), and this could be due to the relative coarse resolution of the GPCP data. The model underestimates the 99th percentile rate in both seasons, difficulty seen in other simulations (Gutowski et al., 2007). These authors suggest that model errors are partly a consequence of the dynamics.

### 3.2. Widespread extreme precipitation

We use the 99th percentile threshold (Table 3.1) to extract days for which extreme precipitation occurs simultaneously on 12 or more grid points, defining these as widespread extreme events. For comparison of model results with both GPCP and TRMM, we normalized GPCP results to the equivalent number of half-degree grid boxes by assuming that 1-grid box in GPCP is equivalent to 4-grid boxes in both TRMM and CEU. From the 910 days analyzed in this study for AMJ, there are 32 days in TRMM, 51 days in GPCP, and 58 days in CEU with such widespread events. For the 920 days analyzed for JAS, there are 46 (TRMM), 85 (GPCP), and 55 (CEU) days with such widespread events. We find that the model tends to produce extreme events covering a wider spatial scale than TRMM at all grid points lower and greater than 12 grid points (Fig. 3.4).
3.3. Analysis of extreme events

Having obtained the widespread events, we form composite of precipitation and associated atmospheric circulations for those days to analyze atmospheric features associated with the extreme.

3.3.1. Precipitation and wind anomalies

Figure 3.6 shows the positive precipitation anomaly for the days with widespread extreme events. The observations show similar structure in the anomaly patterns. In AMJ, the anomaly is concentrated at the southern end of the analysis box, which is where the climatological mean precipitation is also largest (Fig. 3.2). In JAS, the observed anomaly is concentrated farther north, similar to the seasonal shift in the climatological mean precipitation. The maximum intensity of the composite anomaly is weaker in JAS compared to AMJ, mimicking seasonal changes in TRMM, but not GPCP, climatology. The model simulates a precipitation anomaly concentrated in the domain of interest, but its anomaly has weaker magnitude versus observations, consistent with Fig. 3.3 and Table 3.1.

3.3.2. Interannual variability of extreme precipitation

Figure 3.5a shows the intraseasonal variability of days with widespread extreme events. The root-mean-square difference (RMSD) between the TRMM and GPCP frequencies (0.05) is roughly the same as that between CEU and TRMM (0.08) and CEU and GPCP (0.07). These RMSD values can provide an estimated scale for assessing the magnitude of seasonal variations. By this scale, the observed extreme precipitation events show strong seasonal changes with maxima in June and August, straddling a relative minimum in July. The model simulates the seasonal variation of extreme events well in that it has the June and August maxima and a relative minimum in July, and its month-to-month changes are larger than the RMSD values above. The model does produce too many events in May and too few in July and August compared to TRMM.
The interannual variability histogram (Fig. 3.5b) shows somewhat larger RMSD between the different precipitation sources: 0.08 (TRMM-GPCP), 0.09 (TRMM-CEU) and 0.08 (GPCP-CEU). By this measure, only the years with highest or lowest frequencies should be compared for agreement. There is somewhat greater agreement between TRMM and CEU than between GPCP and either TRMM or CEU. Both TRMM and CEU have the same three years with lowest occurrence of extreme events, 2001, 2004 and 2006 (Table 3.2). Also, the model tends to produce years with many widespread extreme events during the same year as TRMM, for example, 1999, 2005, and 2007 (Table. 3.2). Similar agreement does not occur between GPCP and either TRMM or CEU. The model thus replicates some of the interannual variability in extreme events seen in the TRMM output.

The model also produces quite frequently consecutive days with extreme widespread events (Table 3.3). In contrast, consecutive days of observed widespread extreme events are substantially fewer in both observational data sets. The model thus demonstrates persistence in extreme precipitation behavior not occurring in the observations.

3.3.3. The synoptic conditions

3.3.3.1. Statistical significance of anomalies

Understanding the physical processes responsible for extreme events is important for understanding why they occur and how they might change with climate change (e.g., Gutowski et al. 2008). In diagnosing possible relationships between large-scale atmospheric features and precipitation during extreme events, we also examine the statistical significance of the departure of examined fields from climatology over domain of study, partly because some characteristics of the extremes are similar to climatology (e.g., intraseasonal variability) and partly because there is substantial daily and longer variability in the atmospheric fields. We hypothesize that composite fields on extreme precipitation days are statistically different from climatology. We accomplish this through a sequence of steps:
First we detrend the time series of each field examined at each grid point to remove seasonal trends that might give a false indication of significant differences. For each field, we detrend our two seasons, AMJ and JAS, separately. We also detrend each year separately to remove potential effects of interannual variability. Departures from a linear fit to the time series at each point provided the detrended fields.

We then determined the degrees of freedom in our data set. For this, we used the detrended time series to compute the autocorrelation with 1-day lag at each grid point for each field. A lag-1 autocorrelation of 0.8 emerged as a conservative estimate applicable to all fields at all grid points. For our 10 years of approximately 90 days sampled each year, this implied about 100 degrees of freedom. For this relatively large number, the final results are not sensitive to the precise value of the degrees of freedom.

We then assess the statistical significance of differences between composite anomalies and climatology using Welch’s t-test for two samples with unequal variance. The test examines the null hypothesis that the two samples have equal means. We focus on all fields and locations where the null hypothesis is rejected at the 95% confidence level. Further details of the test statistic appear in Agresti and Finlay (1986), and Wilks (1995).

We focus on fields that display field significance. At the 95% confidence level, random noise could yield statistically significant differences between two fields at about 5% of the points examined. We did not perform a formal evaluation of field significance (e.g., Livezy and Chen, 1983; Wilks, 2006) for the fields studied because they clearly had significance at much more than 5% of their points.
3.3.3.2. Significant anomalies

Figure 3.7 shows the vertical structure, averaged over 5°W-5°E, of the detrended vertical velocity field in AMJ on (a) the day of widespread extreme events (left panels), and (b) 1-day before widespread extreme events (right panels) from ERAIM for days with TRMM widespread precipitation extremes and from CEU for days with simulated widespread extremes. The shaded regions show where the vertical velocity is statistically significant at the 95% confidence level. A region of enhanced ascent between the surface and upper atmospheric levels is located between 2°N and 10°N. The core of the ascending motion at about 300 hPa in ERAIM (Fig. 3.7a) coincides with the region of widespread extreme precipitation (Fig. 3.6a). A shallow ascent at 14°N located between 760 hPa and 850 hPa north of the ITCZ, associated with northeasterly trade winds (Fig. 3.6a), is characterized by dry convection (Fig. 3.9a). The model reproduces the latitude-height pattern of the vertical velocity anomaly (Fig. 3.7b), but it underestimates the magnitude. The core of the ascending motion between 5°N and 12°N at about 650 – 400 hPa is associated with the core of widespread extreme precipitation events (Fig. 3.6c). In contrast, the shallow ascent at 14°N is not well defined in the model. In JAS, Figure 3.8 suggests intensification and northward shift of the core of the ascending air, consistent with the northward migration of the zone of the positive extreme precipitation events (Fig. 3.6, right panels). In ERAIM, three zones of maximum ascent appear (Fig. 3.8a). The first region is located at 3°N to 7°N between lower levels and the mid-troposphere and is linked with another one in the upper troposphere. The other two shallow regions of rising motion are located at 16°N and 20°N respectively, with the core of the ascending motion at 16°N, the upper limit of the study domain, located at 800 hPa. Figure 3.8b indicates a fairly good agreement between model and observations on the day of widespread extreme events. The location of the vertical velocity anomaly is well simulated by the model, though the magnitude is slightly higher than observed. Also, the model simulates a single, deep ascent in contrast to ERAIM. The core of the ascending motion is consistent with the model’s ITCZ (Fig. 3.2f), located between 12°N and 16°N, and extends from
about 750 hPa to 400 hPa. Similarly, comparisons of Figs. 3.7 and 38 with Fig. 36 suggest a possible link between convection and extreme precipitation events, consistent with earlier studies. Nicholson (2009) suggests that wet years over West Africa are associated with the enhancement of the vertical motion within the rainbelt embedded between the axes of African easterly jet (AEJ) and Tropical easterly jet (TEJ). During an investigation of the 2007 flood occurrence in West Africa, Paeth et al. (2010) shows that anomalous upward vertical motion dominated the upper-levels, with peak anomalies located north of latitude 10°N, the location of the strongest August 2007 rainfall.

Examining the latitude-height cross sections of simulated and observed vertical velocity 1-day before extreme events, and comparing simulation with observations reveals several important features (Figs. 3.7 and 3.8; right panels). One can see that there are differences in ERAIM between vertical velocity on the day of extreme events and one day before. For AMJ, the ascent one day before is weaker. Also, in JAS, there is anomaly subsidence the day before at 20°N which is replaced by ascent there on the day of the extreme events. The model roughly captures the spatial pattern in AMJ (Fig. 3.7c,d), but has a larger difference from observations in JAS (Fig. 3.8c,d). Although the anomaly vertical motion patterns in the model are different in AMJ and JAS, in both seasons, the motion the day before is quite similar to the anomaly vertical motion on the day of the event. The model produces consecutive days with extreme events fairly frequently, and this persistence appears in the anomaly vertical motion fields on the day of the event and one day before.

Figure 3.9 shows the meridional cross section of the specific humidity anomalies for AMJ, averaged over 5°W-5°E, on (a) the days of widespread extreme events (left panels), and (b) 1-day before widespread extreme events (right panels). Only statistically significant anomalies appear. On the day of the extreme events, the observations and the model both have significant humidity in the region 0°-15°N. However, the positive moisture anomaly extends through much of the troposphere in the model, whereas the observations have no significant near-surface humidity anomaly. The
model also shows two layers of positive anomaly, in contrast to the single layer in the observations that appears midway between the two simulated layers of high anomaly humidity.

In JAS, the detrended specific humidity anomalies (Fig. 3.10), are comparable in magnitude to the specific humidity fields in AMJ (Fig. 3.9a,b) for each source, though in both ERAIM and the model, the positive moisture anomalies shift northward. The negative anomaly, located earlier at 15°-20°N in AMJ (Fig. 3.9a), in ERAIM shift northwards and weakens, while in the model, negative anomalies north and south of the positive moisture anomaly intensify. From Figs. 3.9 and 3.10, one sees that the model simulates approximately the same positive specific humidity anomalies as observed, but with stronger magnitude. The latitudinal location of the positive moisture anomaly coincides with the domain (6°-16°N) of the extreme precipitation events implying that high humidity is an important factor for the extreme precipitation events examined here. The humidity anomalies one day before the extreme events show the same behavior produced by the vertical velocity fields: the model shows more day-to-day persistence than the observations for the anomaly specific humidity when widespread extreme precipitation events occur.

The African easterly wave (AEW) is an important westward propagating system that modulates daily precipitation distribution during boreal summer in West Africa is (e.g., Carlson, 1969; Burpee, 1972; Chen 2006a,b). The wave originates between 20° and 30°E, propagate westward in the lower atmosphere between 5° and 15°N with maximum amplitude in the meridional wind located between 850–650 hPa. Chen (2006b) observed maximum root-mean-square (RMS) of the 2–7 day filtered meridional wind, at 5°W, near the surface and 600 hPa levels. Figure 3.11 shows the spatial distribution of composite 600 hPa streamline superimposed with isotach of detrended meridional wind during AMJ and JAS in ERAIM and model.

Over West Africa, the figures depict different wind circulations. However, as indicated by the 600 hPa streamline analysis in Fig. 3.11a from ERAIM in AMJ, four cyclonic (C) and three divergence
(D) centers, and one anticyclonic (H) center are found. First, over land, the cyclonic centers are located at 0°E 20°N and 5°W 27°N, the divergent centers at 10°W 15°N and 7°E 15°N, while the anticyclonic center is at 7°E 30°N. Second, the cyclonic centers over the oceanic area are located at 7°E 3°N in the Gulf of Guinea and 25°W 7°N inside the ITCZ, the divergent center at 22°E 17°N in the northern part of the eastern tropical Atlantic off the coast of Mauritania/Senegal. A concave shape trough-axis aligned roughly north-south orientation merges with the convergence over the Gulf of Guinea. As inferred from the dipole patterns of the easterly wave activity, the negative phase of the wave within the square box associated with the trough axis is accompanied by northeasterly trade winds. In JAS (Fig. 3.11c), two cyclonic and an anticyclonic centers are observed. The southern track cyclonic vortex associated with the northeast-southwest oriented trough axis is located at 8°W 12°N. The cyclonic and anticyclonic vortices along the Saharan thermal low separated by maximum southerly flow are located at 4°E 24°N and 12°E 22°N respectively. The positive phase of the wave during widespread extreme precipitation events accompanied by southwesterly trade winds is associated with the cyclonic vortex. The circulation structure over West Africa during these periods suggests that the waves are being transported by the cyclonic circulations, and are generally associated with moist convection as indicated by Figs. 3.9 and 3.10, consistent with earlier studies (Diedhiou et al., 1999). The locations of the wave activity and extreme precipitation events coincide, suggesting a link between them.

The model simulates realistic circulation patterns and wave activity, but the dipole patterns of the easterly waves are somewhat different from ERAIM. In AMJ (Fig. 3.11c), the cyclonic vortices linked together by a trough axis oriented northeast-southwest are located at 11°E 28°N and 18°W 8°N. The eastward extension of the cyclonic circulation over the Atlantic Ocean, between west coast and 20°W, merge with the southwesterly flow from the Gulf of Guinea to become southeasterly flow within the box. This suggests the availability of abundant moisture over the region during widespread extreme events, consistent with Fig. 3.9. The core of the positive phase of the easterly wave
embedded within the southwesterly-southeasterly trade winds is associated with the eastward extension of the cyclonic vortex and the trough axis. However, in JAS (Fig. 3.11d), the negative phase of the wave within the square box accompanied by westerly winds is associated with the trough axis and cyclonic vortex at 3°E 15°N (C). This vortex linked two other vortices at 12°W 18°N (C) and 18°E 21°N (C) respectively. Figure 3.10 suggests that the wave behavior is associated with moist convection. Given the nature of the wave's dipole patterns, it can be inferred from the figures that the negative phase of the easterly wave formed ahead of the trough axis while the positive phase formed behind the cyclonic vortex within the trough axis during extreme events. Figure 3.11 confirms that easterly waves coupled with their associated cyclonic vortices/wave troughs are associated with extreme precipitation events.

3.3.3.3. Correlation coefficients

An additional way to examine the relationship between atmospheric circulations and moisture with respect to extreme precipitation events is to find the correlation coefficients between them. We calculated the correlation coefficients between the simulated specific humidity and vertical velocity, and between simulated specific humidity and meridional wind. Correlation coefficients are a measure of the relationship between the variables. Table 3.4 gives the spatial correlation coefficients between simulated specific humidity and each of the indicated variables during the two seasons, and during the widespread extreme precipitation events for some specific levels. Some of the correlation coefficients are quite high, and indicate significant values. The correlations between specific humidity and vertical velocity are highly negative and significant in both seasons for all the levels. The correlation between specific humidity and meridional wind is quite different. It gives both positive and negative correlations, though highly correlated.

The spatial pattern of the correlation coefficients between moisture and the indicated variable averaged over 5°W-5°E, for AMJ and JAS 1998–2007 during the widespread extreme events is
shown in Fig. 3.12. The figure shows that correlations are significant, highly positive and negative. The negative correlations are more pronounced in Fig. 3.12a,b over the region, consistent with the coincidence of enhanced upward motion with moist atmosphere. The enhanced upward motion acts as an agent transporting the moisture upward to saturation level where it precipitates. Such interaction explain their role in extreme precipitation events, which supports our earlier suggestion that widespread extreme precipitation events are associated with enhanced rising motions in the vicinity of moist atmosphere. Similarly, there is a good correlation between moisture field and meridional wind. The correlations are highly positive at the surface up to 800 hPa in JAS and up to 700 hPa in AMJ. Thus, more moisture at the lower atmosphere supplies the latent heat necessary to drive the easterly wave. Taken together with the enhanced upward motion, we can infer from these figures that there is coupling between atmospheric circulations and moisture to promote the occurrence of extreme events.

3.3.4. Diurnal variation of extreme precipitation events: Case study

As indicated in Figures 3.7 and 3.8, and Figures 3.9 and 3.10, the model shows persistent behavior in the pattern of the vertical velocity and moisture fields respectively. The persistence observed in the model behavior compared to the variability observed in observation could be explained by Fig. 3.13. We examine the characteristics of precipitation in observation and model on one of the extreme days. We chose August 25, 1998 for observation and August 14, 1998 for model. We found that, in the real world as given by TRMM (Fig. 3.13a-c), heavy precipitation is produced by a propagating squall line, whereas in the model (Fig. 3.13d-f), the squall line did not appear. In the model (Fig. 3.13d-f), heavy precipitation is due to convection scheme simply turning off and on with no horizontal propagation. Thus, in the TRMM data, when extreme precipitation occurs depends on when a rain-bearing convective system propagates through the region (see Fig. 3.14a,b). In contrast, in the model, extreme precipitation always occurs at the time of day that the convection scheme is triggered, and that is nearly the same time for every event.
4. Conclusion

The relationship between extreme daily precipitation events and associated large-scale processes has been examined over a portion of West Africa for the periods April–June (AMJ) and July–September (JAS) 1998 to 2007 in observation-based data sets and in an atmospheric global climate model, CAM-EULAG. The comparison between CEU and the observational sources helps to establish the degree to which the model can replicate the extreme events and the underlying physical behavior supporting them. We compared simulated precipitation with gridded precipitation observational data sets provided by Tropical Rainfall Measurement Mission (TRMM) and Global Precipitation Climatology Project (GPCP). We compare simulated atmospheric fields with output from the ERA-Interim reanalysis (ERAIM).

The model simulates the spatial pattern of precipitation, but underestimates precipitation maxima over highland regions of Cameroon and Nigeria, and along the ITCZ. Comparison of precipitation from the two observational data sets shows that TRMM is higher than GPCP. The diurnal cycles of precipitation averaged over 5°W–5°E and 6°–16°N for TRMM and model are compared for the two seasons. Although the model captures the pattern of the diurnal cycle of precipitation, it produces less (excess) precipitation in the morning (evening) over the domain compared with TRMM observation, suggesting that convection is stronger in the evening hours, consistent with previous study (Lee et al., 2007).

The model’s simulated precipitation frequency versus intensity is different from observations. The model simulates a higher frequency of drizzle, and much lower frequency of events exceeding 25 mm d$^{-1}$. We extracted days that precipitation at each grid point is greater than the 99% level and occurring simultaneously on 12 or more grid points over the core monsoonal region in West Africa, calling these widespread extreme events. An assumption underlying our examination of these events is that they are linked to resolved behavior of the reanalysis or model fields. The model simulates the fairly
well the intraseasonal and interannual variability of days with widespread extreme precipitation events. For seasonal variability, the model replicates the observed double maxima in frequency occurring June and August and the relative minimum in July. On an interannual basis, the model tends to simulate higher frequency of days with widespread extreme events during the same years when TRMM produces higher events, and fewer events during the same years with fewer events in TRMM. The composite anomaly of days with widespread extreme precipitation events in the model also reproduces fairly well the observed behavior, at least within our analysis region.

We employed a statistical approach to determine the significance of synoptic conditions associated with widespread extreme daily precipitation. We show that detrended vertical wind, specific humidity, and meridional wind anomalies associated with widespread extreme daily precipitation are significantly different from their corresponding time average climatology at the 95% confidence level over much of the region, suggesting that there is a link between extreme precipitation events and these synoptic conditions. Compared to ERAIM, the model simulates well the latitude-height pattern of the vertical velocity anomaly, though it underestimates (overestimates) the magnitude in AMJ (JAS). The core of the ascending motion lying along the ITCZ is associated with the core of widespread extreme precipitation events. Similarly, the model simulates fairly well the detrended composite moisture field anomalies associated with widespread extreme daily precipitation. The significant, positive moisture anomaly coincides with the extreme precipitation events over the region. The model produces abundant moisture in the lower atmosphere up to about 450 hPa levels.

We also analyzed the characteristics of African Easterly Waves during extreme events. The spatial pattern of the wave on extreme event days shifts westward between AMJ and JAS. The model also captures the spatial patterns of the wave behavior, but its phases differ from the phases of observed behavior on extreme event days. The phase of the wave associated with cyclonic vortices is consistent
with the location of the positive anomaly of the extreme precipitation event. Therefore, the wave's behavior also appears to play a significant role in promoting precipitation extremes.

In fact, the spatial relationship between extreme precipitation and each of the associated atmospheric features could not be a coincidence. Their separate interactions suggest that there is a link between them. Correlation coefficients between humidity and either vertical or meridional motion can be quite high, as large as 0.86 in magnitude, indicating significant values. Negative correlations are more pronounced over the region of enhanced upward motion, consistent with a coupling of ascent (negative pressure velocity) with positive humidity anomaly. Also, the correlations between moisture field and meridional wind can be highly negative, especially at the level of the maximum amplitude of AEW.

Examining the latitude-height cross sections of simulated and observed synoptic conditions 1-day before extreme events reveals that there is persistence in the model behavior but not the observed behavior. Finally, we examined the spatial distribution of the precipitation on one of the extreme days in observation and simulation. We found that, in TRMM, heavy precipitation is produced by propagating squall lines, whereas, in the model, there is no indication of a squall line. Heavy precipitation is a result of convection scheme in the model firing-off continuously at any one day at the same time.

The relationship observed between large scale atmospheric features and extreme precipitation events in observation and model in this study will help understand the physical mechanisms conducive to extreme events in this region under the present and future climate change scenarios. These results should be considered with caution because of the limited numbers of large-scale features (e.g., vertical velocity, humidity and meridional wind) considered. Therefore, further study is needed to examine other large-scale circulation processes (e.g., geopotential height, etc.) conducive to extreme events since the region’s precipitation is affected by different factors.
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References


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Figure Captions

Figure 3.1 Horizontal grid resolution for CEU-SG simulation. Panel (a) shows the zonal distribution for the grid points along 18°N (line). Panel (b) shows the double nested distribution used for the meridional grid points. Panel (c) shows the combined effect of meridional zonal mappings for grid. Only 20% of the grid points are shown for clarity.

Figure 3.2 Composite mean precipitation (mm day$^{-1}$) from (a,d) TRMM, (b,e) GPCP, and (c,f) CEU for AMJ (left panels) and JAS (right panels) during the period 1998-2007. The scale is shown at the lower right corner for all figures, and only precipitation greater than 5 mm day$^{-1}$ is shaded grey.

Figure 3.3 Normalized frequency of precipitation as a function of daily intensity for (a) AMJ and (b) JAS during 1998-2007 in TRMM, GPCP, and CEU. Arrows mark the 99th percentile of daily precipitation.

Figure 3.4 Days with simultaneous extremes on 12 or more grid points for CEU and observations for (a) AMJ and (b) JAS.

Figure 3.5 (a) Intraseasonal and (b) interannual normalized distribution of the frequency of days with widespread extreme precipitation events.

Figure 3.6 Composite precipitation anomalies (mm day$^{-1}$) for positive widespread precipitation events overlaid with 925 hPa level wind vector anomalies in AMJ (left panel) and JAS (right panel) for (a,d) TRMM, (b,e) GPCP, and (c,f) CEU. Only precipitation anomaly values greater than 2 mm day$^{-1}$ are shaded. The vector scale for (a and d) is 15 m s$^{-1}$, (b,e) is 15 m s$^{-1}$, and for (c,f) is 5 m s$^{-1}$. Note the difference in the scale between model and observations.

Figure 3.7 Vertical structure of composite detrended vertical velocity statistically significant at the 95% confidence level during widespread extreme events (left panels) and 1-day before
widespread extreme events (right panels) averaged over 5°W-5°E during AMJ 1998 – 2007 for (a,c) ERAIM and (b,d) CEU. White areas indicate detrended vertical velocity not statistically significant at the 95% confidence level (p > 0.05). Contour scale for all plots is shown at the lower right, in mb s⁻¹.

**Figure 3.8** Same as in Fig. 3.7, except for JAS.

**Figure 3.9** Vertical structure of composite detrended specific humidity statistically significant at the 95% confidence level during widespread extreme events (left panels) and 1-day before widespread extreme events (right panels) averaged over 5°W-5°E during AMJ 1998 – 2007 for (a,c) ERAIM and (b,d) CEU. White areas indicate detrended specific humidity not statistically significant at the 95% confidence level (p > 0.05). Contour scale for all plots is shown at the lower right, in kg/kg (x 10⁴).

**Figure 3.10** Same as in Fig. 3.9, except for JAS.

**Figure 3.11** The 600 hPa streamline and isotach of detrended meridional wind (shaded) during widespread extreme events (left panels) in (a,c) ERAIM and (b,d) CEU during AMJ(left panels) and JAS (right panels) 1998 – 2007. Shading contour for all plots is shown by the scale in the lower right side of (d) in m s⁻¹.

**Figure 3.12** Vertical structure of correlation coefficients between (a and b) simulated specific humidity and vertical velocity fields, and (c and d) simulated specific humidity and meridional wind fields during widespread extreme events averaged over 5°W-5°E.

**Figure 3.13** Spatial distribution of diurnal precipitation (mm day⁻¹) during one of the extreme days in TRMM (a-c) and CEU (d-f). The model values have been multiplied by a factor of 16 for comparison with TRMM.

**Figure 3.14** Diurnal cycle of precipitation during some of the extreme days in TRMM (a-c) and CEU (d-f), averaged over 5°W-5°E, 6°N-16°N for AMJ and JAS during the period 1998 – 2000.
Fig. 3.1: Horizontal grid resolution for CEU simulation. Panel (a) shows the zonal distribution for the grid points along 18°N (line). Panel (b) shows the double nested distribution used for the meridional grid points. Panel (c) shows the combined effect of meridional and zonal map pings for grid. Only 20% of the grid points are shown for clarity.
Fig. 3.2: Composite mean precipitation (mm day$^{-1}$) from (a,d) TRMM, (b,e) GPCP (1x1), and (c,f) CEU for AMJ (left panels) and JAS (right panels) during the period 1998-2007. The scale is shown at the lower right corner for all figures, and only precipitation greater than 5 mm day$^{-1}$ is shaded grey.
Fig. 3.3: Normalized frequency of precipitation as a function of daily intensity for (a) AMJ and (b) JAS during 1998-2007 in TRMM, GPCP, and CEU. Arrows mark the 99th percentile of daily precipitation.
Fig. 3.4: Days with simultaneous extremes on 12 or more grid points for CEU and observations for (a) AMJ and (b) JAS.
Fig. 3.5: (a) Intraseasonal and (b) interannual normalized distribution of the frequency of days with widespread extreme precipitation events.
Fig. 3.6: Composite precipitation anomalies (mm day\(^{-1}\)) for positive widespread precipitation event overlaid with 925 hPa level wind vector anomalies in AMJ (left panel) and JAS (right panel) for (a,d) TRMM, (b,e) GPCP, and (c,f) CEU. Only precipitation anomaly values greater than 2 mm day\(^{-1}\) are shaded. The vector scale for (a,d) is 15 m s\(^{-1}\), (b,e) is 15 m s\(^{-1}\), and for (c,f) is 5 m s\(^{-1}\). Note the difference in the scale between model and observations.
Fig. 3.7: Vertical structure of composite detrended vertical velocity statistically significant at the 95% confidence level during widespread extreme events (left panels) and 1-day before widespread extreme events (right panels) averaged over 5°W-5°E during AMJ 1998 – 2007 for (a,c) ERAIM and (b,d) CEU. White areas indicate detrended vertical velocity not statistically significant at the 95% confidence level (p > 0.05). Contour scale for all plots is shown at the lower right, in mb s⁻¹.
Fig. 3.8: Same as in Fig. 13, except for JAS.
Fig. 3.9: Vertical structure of composite detrended specific humidity statistically significant at the 95% confidence level during widespread extreme events (left panels) and 1-day before widespread extreme events (right panels) averaged over 5°W-5°E during AMJ 1998 – 2007 for (a,c) ERAIM and (b,d) CEU. White areas indicate detrended vertical velocity not statistically significant at the 95% confidence level (p > 0.05). Contour scale for all plots is shown at the lower right, in kg/kg (x 10^{-4}).
Fig. 3.10: Same as in Fig. 3.9, except for JAS
Fig. 3.11: The 600 hPa streamline and isotach of detrended meridional wind (shaded) during widespread extreme events (left panels) in (a,c) ERAIM and (b,d) CEU during AMJ (left panels) and JAS (right panels) 1998 – 2007. Shading contour for all plots is shown by the scale in the lower right side of (d) in m s$^{-1}$. 
Fig. 3.12: Vertical structure of correlation coefficients between (a and b) simulated specific humidity and vertical velocity fields, and (c and d) simulated specific humidity and meridional wind fields during widespread extreme events averaged over 5°W-5°E.
Fig. 3.13: Spatial distribution of diurnal precipitation (mm day$^{-1}$) during one of the extreme days in TRMM (a-c) and CEU (d-f). The model values have been multiplied by a factor of 16 for comparison with TRMM.
Fig. 3.14: Diurnal cycle of precipitation during some of the extreme days in TRMM (a-c) and CEU (d-f), averaged over 5°W-5°E, 6°N-16°N for AMJ and JAS during the period 1998 – 2000.
Table 3.1 Precipitation rate for the 99th percentile of daily precipitation (mm day$^{-1}$).

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<th>JAS</th>
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<td>GPCP</td>
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<td>45.4</td>
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<td>CEU</td>
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<td>19.0</td>
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Table 3.2  Ranked percentage of widespread extreme precipitation events occurring in each year.

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<thead>
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<th>GPCP</th>
<th>CEU</th>
</tr>
</thead>
<tbody>
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<td>Normalized Days</td>
<td>Year</td>
<td>Normalized Days</td>
</tr>
<tr>
<td>---------</td>
<td>-----------------</td>
<td>------</td>
<td>-----------------</td>
</tr>
<tr>
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<td>2003</td>
<td>0.13</td>
</tr>
<tr>
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<td>1998</td>
<td>0.12</td>
</tr>
<tr>
<td>1999</td>
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</tr>
<tr>
<td>2005</td>
<td>0.13</td>
<td>2005</td>
<td>0.11</td>
</tr>
<tr>
<td>2003</td>
<td>0.10</td>
<td>2006</td>
<td>0.11</td>
</tr>
<tr>
<td>2000</td>
<td>0.08</td>
<td>2002</td>
<td>0.10</td>
</tr>
<tr>
<td>2002</td>
<td>0.05</td>
<td>2007</td>
<td>0.10</td>
</tr>
<tr>
<td>2006</td>
<td>0.05</td>
<td>1999</td>
<td>0.08</td>
</tr>
<tr>
<td>2004</td>
<td>0.03</td>
<td>2000</td>
<td>0.07</td>
</tr>
<tr>
<td>2001</td>
<td>0.03</td>
<td>2004</td>
<td>0.07</td>
</tr>
</tbody>
</table>
Table 3.3 Percentage of widespread extremes occurring as part of consecutive-day events.

<table>
<thead>
<tr>
<th>Data</th>
<th>AMJ</th>
<th>JAS</th>
</tr>
</thead>
<tbody>
<tr>
<td>TRMM</td>
<td>13%</td>
<td>26%</td>
</tr>
<tr>
<td>GPCP</td>
<td>20%</td>
<td>39%</td>
</tr>
<tr>
<td>CEU</td>
<td>81%</td>
<td>69%</td>
</tr>
</tbody>
</table>
Table 3.4  Spatial correlation coefficients between the simulated specific humidity and the indicated variable during AMJ and JAS over the region of study during extreme precipitation events.

<table>
<thead>
<tr>
<th>Variable</th>
<th>AMJ</th>
<th>JAS</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>925 hPa</td>
<td>850 hPa</td>
</tr>
<tr>
<td>Vertical velocity</td>
<td>-0.38*</td>
<td>-0.54*</td>
</tr>
<tr>
<td>Meridional wind</td>
<td>+0.76*</td>
<td>-0.03</td>
</tr>
</tbody>
</table>

* Statistically significant at the 95% confidence interval.
CHAPTER 4. GENERAL CONCLUSIONS

1. Discussion

Hurrell, et al. (2006) states that climate system models are needed to understand the wide range of complex interacting processes (physical, chemical, and biological) that govern the atmosphere, ocean, and land. However, before we can gain confidence in the simulations of any climate model, it is very important to access the strengths and weaknesses of the model. Therefore, this study summarizes the performance of one such model, CAM-EULAG (CEU) when simulating the major synoptic features of the West African monsoon system and extreme precipitation events and their accompanying large-scale processes.

The climatological mean fields of precipitation, winds, and temperature in July-August-September (JAS) compare well with observations. The model captures the rainbelt associated with the ITCZ and the three phases of the monsoon circulation over West Africa: onset, intensification, and cessation. Also, the northward shift and weakening of the AEJ, the appearance and intensification of the TEJ, and the strong ascent between the levels of the AEJ and TEJ associated with the rainbelt are realistically simulated. A northward shift in the core of the AEJ and appearance of the TEJ during the peak summer monsoon season is observed in conjunction with the zone of maximum precipitation. The main deficiencies depicted by the model over West Africa are the underestimation of rainfall over orographic regions, the weakening and northward displaced AEJ, and too strong and northward displaced TEJ. The observed model deficiencies result in underestimation (overestimation) of the Guinean coast (Soudano-Sahel) precipitation. Previous studies have shown that these are common to many models.

In Chapter 3, we examined the potential of CEU to simulate extreme precipitation events and associated atmospheric circulations over West Africa. We compared the simulated precipitation with
gridded precipitation data sets provided by the Tropical Rainfall Measurement Mission (TRMM) and the Global Precipitation Climatology Project (GPCP). We also compared the simulated atmospheric fields with output from the ERA-Interim reanalysis (ERAIM). We found that the model did well in simulating the patterns of the mean diurnal cycle of precipitation over the target region. Also, the spatial patterns of the precipitation anomalies during widespread events in AMJ and JAS were realistic. The model tended to mimic observations in that it reproduces the high frequency of days with widespread extreme precipitation events for the same periods as in TRMM. The model simulated fairly well the latitude-height structure of the vertical velocity and moisture anomaly fields that had statistically significant departures from climatology. The core of the enhanced upward motion inside the ITCZ, consistent with the core of maximum moisture, coincided with the location of the maximum precipitation anomaly. Similarly, the spatial patterns of the African easterly wave’s activity depicted by the 600 hPa meridional wind in AMJ and JAS during widespread extreme precipitation events were fairly well simulated by the model. CEU simulated the observed amplitudes of the westwards propagating system.

Despite the performances of the model, there were major deficiencies. First, the magnitude of the composite precipitation anomaly for widespread extreme precipitation events was underestimated. We found that model underestimated precipitation intensity greater than 12.5 mm day$^{-1}$ (17.5 mm day$^{-1}$) in AMJ (JAS) and produced too little high-intensity precipitation compared to observation. The magnitude of the simulated anomaly vertical velocity is too weak in AMJ, though of the similar strength to the observations in JAS, but it is located farther north than in ERAIM in both seasons. Similarly, the model produces too much anomaly moisture especially in the lower atmosphere. Examination of the model’s performance in simulating vertical velocity and moisture fields 1-day before extreme events shows that there is daily persistence in the model behavior; the spatial patterns bear resemblance to those on the extreme days. An examination of the precipitation on one of the extreme days in TRMM and the model indicates that heavy precipitation in TRMM is a result of a
propagating squall line, whereas heavy precipitation in model is due to convection scheme simply turning on and off with no horizontal propagation.

2. **Some Recommendations for Future Research**

This study has shown many important positive features of the climatology of the global climate model, CAM-EULAG. Additional positive features appear in Abiodun et al. (2011). Despite these good results, there are notable differences in the model behavior compared to observations. Therefore, there are several tasks that could improve the deficiencies noted in this study. Among them are:

1) Resolution sufficient for squall lines may help with simulating extreme precipitation events.

2) The convection parameterization may need adjustment to avoid day-to-day persistence in extreme events.

3) The representation of land cover may need modification to properly simulate the dynamics of West African circulation (Abiodun et al., 2007).

**References**

