A Nonlinear Response of Sahel Rainfall to Atlantic Warming

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ABSTRACT

The response over West Africa to uniform warming of the Atlantic Ocean is analyzed using idealized simulations with a regional climate model. With warming of 1 and 1.5 K, rainfall rates increase by 30%–50% over most of West Africa. With Atlantic warming of 2 K and higher, coastal precipitation increases but Sahel rainfall decreases substantially. This nonlinear response in Sahel rainfall is the focus of this analysis. Atlantic warming is accompanied by decreases in low-level geopotential heights in the Gulf of Guinea and in the large-scale meridional geopotential height gradient. This leads to easterly wind anomalies in the central Sahel. With Atlantic warming below 2 K, these easterly anomalies support moisture transport from the Gulf of Guinea and precipitation increases. With Atlantic warming over 2 K, the easterly anomalies reverse the westerly flow over the Sahel. The resulting dry air advection into the Sahel reduces precipitation. Increased low-level moisture provides moist static energy to initiate convection with Atlantic warming at 1.5 K and below, while decreased moisture and stable thermal profiles suppress convection with additional warming. In all simulations, the southerly monsoon flow onto the Guinean coast is maintained and precipitation in that region increases. The relevance of these results to the global warming problem is limited by the focus on Atlantic warming alone. However, confident prediction of climate change requires an understanding of the physical processes of change, and this paper contributes to that goal.

1. Introduction

An improved understanding of the physical processes associated with climate change in West Africa and the Sahel is needed for producing reliable predictions to aid resource management and planning as the planet warms. Nonlinear responses in the climate system can lead to surprises and even abrupt climate change, presenting greater challenges for adaptation. Paleoclimate proxy data show that the West African climate is prone to rapid changes between wet and dry conditions. Even the seasonal evolution of the monsoon system is characterized by abrupt transitions because the northward migration of rainfall from the Guinean Coast into the Sahel does not occur smoothly.

The purpose of this paper is to contribute to an improvement in our physical understanding of how the West African monsoon will respond under global warming. We isolate one aspect of the problem, namely the hydrodynamics of the response to uniform warming of the Atlantic Ocean within the model domain (19°S–34°N, 58°W–48°E; see section 3). Forcing from the Atlantic Ocean is chosen because of the known sensitivity of the monsoon system to Atlantic sea surface temperature anomalies (SSTAs). Uniform warming is chosen for simplicity and clarity, and because the modeling results of Patricola and Cook (2011) indicate that much of the global warming signal forced from the Atlantic in West Africa and the Sahel is caused by uniform warming, with a dependence on structure in the SSTAs present only along the west coast.

Climate change and variability in West Africa and the Sahel is complicated, with demonstrated dependence on SSTAs in the Indian, Atlantic, and Pacific Oceans, land surface feedbacks, aerosol forcing, midlatitude–tropical interactions, and other factors—and we do not mean to imply otherwise by focusing on Atlantic sea surface temperature (SST) forcing alone. However, confident prediction of twenty-first-century climate change in the region requires that we build a solid understanding of the physical processes of that change, and here we contribute to that goal in a process study that isolates the response to uniform Atlantic SSTAs.
Background is provided in section 2, including an overview of the region’s sensitivity to Atlantic SSTs and our current understanding of the potential for change under greenhouse gas forcing. The regional climate model and experimental design are described in section 3, and the model is validated against observations, reanalysis, and satellite measurements in section 4. Results are reported in section 5, and conclusions and implications are discussed in section 6.

2. Background

Paleoclimate proxy data show that the West African climate can change substantially on millennial time scales. For example, 14.8–5.5 thousand years (ka) before present (BP) is known as the African Humid Period (AHP), when the western Sahel and Sahara received rainfall sufficient to support large lakes, small trees, and large fauna such as hippopotamus (e.g., Gasse and Van Campo 1994; Petit-Maire and Guo 1996). The onset of the AHP was associated with some major atmospheric changes. For example, the atmospheric CO$_2$ concentration increased by about 50 ppmv, the Atlantic meridional overturning circulation (AMOC) weakened and the northern ice sheets retreated during the onset, and summer insolation in the Northern Hemisphere was about 5% greater than today at 8 ka BP (White et al. 1994; Monnin et al. 2001; McManus et al. 2004).

Of potential relevance to the development of the global warming response over West Africa and the Sahel is the ability of the region’s climate to change abruptly. For example, the transition into and out of the AHP was abrupt, at least in some regions, occurring within decades to centuries despite gradual and smooth changes in the precessional forcing of insolation (e.g., deMenocal et al. 2000; Russell et al. 2003; Peck et al. 2004). Also, in a more recent model simulation, Timm et al. (2010) demonstrate that the abrupt onset of the AHP was caused by the combined effects of increases in local insolation and drastic melting of the Northern Hemisphere ice sheets.

Paleoclimate studies also provide an indication of the sensitivity of the West African climate to Atlantic SSTs. The African Humid Period was interrupted by an arid interval corresponding to the Younger Dryas event (~12.9–11.5 ka BP), when a weakening of the AMOC occurred in response to a sudden influx of freshwater from deglaciation in North America (Alley 2000; Lowell et al. 2005). Heinrich events are also reflected in African paleoclimate records; there is evidence that North Atlantic cooling at 17 ka BP associated with Heinrich event 1 caused a widespread drought in the African monsoon region (Stager et al. 2011).

Present-day climate records also indicate that the West African climate is sensitive to Atlantic SSTs (e.g., Folland et al. 1986; Fontaine and Bigot 1993; Camberlin et al. 2001; Liu and Chiang 2012). One particular SSTA forcing pattern known to influence West Africa rainfall is centered in the Gulf of Guinea, where the onshore monsoon flow originates. With positive SSTAs of 1–2 K in the Gulf of Guinea, rainfall is enhanced along the Guinean coast and suppressed in the western and central Sahel (Ward 1998; Nicholson 2000; Mo et al. 2001). An opposite pattern occurs with cool SSTAs in the Gulf of Guinea. Vizy and Cook (2001, 2002) explain the physics of this mechanism. Evaporation is enhanced over the warm SSTAs, but a precipitation anomaly does not develop over the ocean because of regional subsidence associated with the Walker circulation. As a consequence, the moisture content of the northward monsoon inflow is increased along with rainfall along the coast. Anomalous subsidence and drying in the Sahel is a secondary response to enhanced rainfall along the coast generated by conservation of potential vorticity requirements.

Although the focus of this paper is on the physics of the response to uniform Atlantic Ocean warming, many factors are known to influence precipitation over West Africa. These include changes in SST patterns in the Atlantic (e.g., Mohino et al. 2011a), SSTAs in the Pacific and Indian Oceans (e.g., Paeth and Friederichs 2004; Hagos and Cook 2008; Mohino et al. 2011b), changes in land surface/vegetation (e.g., Li et al. 2007; Patricola and Cook 2008), aerosol forcing (e.g., Gu et al. 2012; Joseph and Zeng 2011; Zhao et al. 2011), and relative positions of the African easterly jet and the tropical easterly jet (Nicholson 2009). Because the model domain does not include the Indian and Pacific Oceans, the impacts of large-scale circulations such as the Madden–Julian oscillation, the El Niño–Southern Oscillation, and the Indian summer monsoon on the West African monsoon are not included.

The need to improve our basic understanding of the physics of the African monsoon’s response to climate forcing is evident in the uncertainty about how the region’s rainfall will change in the future (Biasutti and Giannini 2006; Biasutti et al. 2008; Cook 2008). Cook and Vizy (2006) examined output from 18 coupled GCMs from phase 3 of the Coupled Model Intercomparison Project 3 (CMIP3; Meehl et al. 2007) over West Africa. Half of the CMIP3 simulations failed to produce the West African monsoon at all in the sense that they did not bring the summer precipitation maximum onto the continent. Three GCMs were selected based on their ability to represent major features of the twentieth-century climate, including its variability. But in the future simulations, these three models diverged widely,
demonstrating that a reasonable simulation of the present climate is not a sufficient condition for accurate projection. Inclusive analyses of the CMIP3 coupled GCMs (Cook 2008) show that most of the CMIP3 models predict only small increases or decreases of annual precipitation over West Africa by the end of the century.

There is no overall improvement in simulating the West African monsoon system in the coupled GCMs from the CMIP5 (Vizy et al. 2013). Some GCMs, such as the Community Climate System Model, version 4.0 (CCSM4.0; Cook et al. 2012), produce an improved representation over West Africa compared with the CMIP3, but in other models the simulation is worse. Maynard et al. (2002) used a coupled GCM to simulate African climate at the end of the twenty-first century under the Intergovernmental Panel on Climate Change (IPCC) B2 scenario and found that the summer monsoon precipitation over the Sahel increases.

Patricola and Cook (2010, 2011) use a regional model at 90-km resolution centered over West Africa to produce two 9-member ensembles for the end of the twentieth and twenty-first centuries, with boundary conditions for future simulations applied as anomalies based on output from nine coupled GCM. Despite the differences in SSTAs and lateral boundary conditions in the nine GCMs, there is good agreement in the predictions over most of West Africa and the Sahel among the nine ensemble members. Annual precipitation changes are relatively small, similar to the CMIP3 multimodel ensemble mean (Cook 2008), but regional and monthly analysis reveals large changes in temperature, heat stress, precipitation, and extreme rainfall events that would produce pronounced impacts. In particular, rainfall in the Sahel increases in July and August and decreases along the Guinean coast in June and July.

The reason for the agreement among the ensemble members in the regional model simulations of Patricola and Cook (2010) is that the first-order response over West Africa and the Sahel at the end of the twenty-first century is to overall ocean warming, which is demonstrated by an additional simulation with uniform SSTAs of 2 K. The changes in precipitation in the uniform warming case are similar to those from the ensemble mean across West Africa and the Sahel, with the only exception in the far western Sahel where the response is sensitive to the structure of the imposed SSTAs.

Building on these results, we explore the physics of the response to uniform Atlantic warming alone in a series of idealized simulations. We are particularly interested in identifying and understanding nonlinearity in the monsoon system’s response to gradual warming, because these can be a source of abrupt change.

### 3. Model description and experimental design

The National Center for Atmospheric Research (NCAR) Advanced Research Weather Research and Forecasting model, version 3.1.1 (WRF 3.1.1), is used for this study (Skamarock et al. 2008). The model is fully compressible and nonhydrostatic, with 28 vertical levels in this case. The top of the model is set to 50 hPa. The model is used at a spatial resolution of 90 km with a time step of 3 min.

The outcome of past regional climate modeling efforts over Africa (e.g., Vizy and Cook 2002, 2003; Patricola and Cook 2010, 2011) and extensive testing were used to select parameterizations, a model domain, and lateral and surface boundary conditions that produce a realistic simulation of the present-day West African climate. Parameterizations chosen include the Mellor–Yamada–Janjic planetary boundary layer scheme (Mellor and Yamada 1982; Janjic 1990, 1996, 2002), the Monin–Obukhov–Janjic surface layer scheme (Monin and Obukhov 1954; Janjic 1994, 1996, 2002), the new Kain–Fritsch cumulus scheme (Kain and Fritsch 1990, 1993), the Purdue–Lin microphysics scheme (Lin et al. 1983), the Rapid Radiative Transfer Model longwave radiation parameterization (Mlawer et al. 1997), the fifth-generation Pennsylvania State University–NCAR Mesoscale Model (MM5) shortwave parameterization (Dudhia 1989), and the unified Noah land surface model (Chen and Dudhia 2001).

Figure 1 shows the model domain selected for this study. The large domain minimizes the effects of the lateral boundaries on the region of interest. It includes most of northern Africa as well as tropical and eastern North Atlantic.

Initial and lateral/surface boundary conditions for all simulations are initialized at 0000 UTC 15 March using climatological March conditions interpolated onto the regional model grid. Lateral boundary conditions for horizontal winds, temperature, relative humidity, and geopotential height are derived from the 1958–2001 National Centers for Environmental Prediction (NCEP)–NCAR reanalysis (Kalnay et al. 1996) monthly climatology. Initial surface boundary conditions for skin temperature, soil temperature, soil moisture, and sea level pressure are derived from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; 1958–2001) monthly climatology (Uppala et al. 2005). The ERA-40 is used for the surface initialization instead of the NCEP reanalysis climatology because the NCEP reanalysis has a cold and wet bias over the Sahara. Patricola and Cook (2010) show that WRF produces a more realistic simulation of the monsoon when ERA-40 surface conditions are used to initialize the simulation.
Lateral boundary conditions are updated every 6 h using the NCEP reanalysis climatology interpolated onto the model boundaries. Monthly-mean values from the reanalysis are taken to represent the middle of the month, and linear interpolation in time is used to generate 6-hourly boundary conditions. Boundary conditions produced by this method include seasonality, but diurnal and synoptic time scales are suppressed. This “climate mode” approach works for large domains in the tropics (Cook and Vizy 2008; Hagos and Cook 2007; Patricola and Cook 2010) but fails for applications in middle latitudes where transients propagating into the domain are important for determining the time-mean state (Patricola and Cook 2013a,b; Pu et al. 2012). Seven 231-day simulations (15 March–31 October) are produced, and 3-hourly output is averaged to generate daily and monthly means. This period covers the entire evolution of the boreal summer monsoon. The first 2.5 months of each simulation, from 15 March to 31 May, are devoted to model spinup and not used in the analysis. Such a long period for spinup is necessary to avoid errors associated with the persisting initial soil moisture properties (Simmonds and Hope 1998).

A control simulation (CTL) uses climatological SSTs that are prescribed and updated every 6 h. These 6-hourly SSTs are derived from the 1958–2001 ERA-40 monthly climatology using the same linear interpolation method discussed above for the lateral boundary conditions. Six additional simulations are produced with uniform Atlantic SSTAs added to the climatological SSTs, but otherwise identical to CTL. These SSTs preserve the structure of current Atlantic SST gradients, but create an overall warmer ocean basic state. Uniform positive SSTAs are applied in six additional simulations referred to as 1K, 1.5K, 2K, 2.5K, 3K, and 4K.

4. Model evaluation

Figure 2 shows boreal summer monthly rainfall from the 0.25°-resolution National Aeronautics and Space Administration (NASA) Tropical Rainfall Measuring Mission (TRMM 3B42V6) satellite-derived 1998–2009 climatology (Kummerow et al. 1998), the 0.5°-resolution Climatic Research Unit (CRU; CRUTS3.0) 1979–2006 climatology (Mitchell and Jones 2005) over land and the 2.5°-resolution Global Precipitation Climatology Project (GPCP; Adler et al. 2003) 1979–2006 over the ocean, and the control simulation. Each climatology is presented at its original resolution to preserve as much spatial information as possible. The regional model simulation is able to realistically capture the seasonal migration of rainfall over West Africa, including the movement of the precipitation maximum into the continental interior that is often missed in coupled GCMs.

Overall, the simulated rainfall intensity is larger than observed. This wet bias varies based on location and can range from 2 to more than 18 mm day$^{-1}$. For example, over the far western Sahel (10°–12°N, 6°–12°W) the difference between the simulated and observed precipitation is approximately 8 mm day$^{-1}$ in July and 4 mm day$^{-1}$ in August. Also, the summer precipitation minimum over the Guinean Coast in July and August is more pronounced.
in the control simulation than the TRMM and CRU climatologies.

The accuracy of the simulated rainfall distribution is similar to that of the most accurate GCMs in this region, and better than most GCM simulations. This includes the sudden shift of the rainfall maximum from the Guinean coast into the Sahel in early July known as the monsoon jump (not shown).

The 925-hPa June geopotential heights and winds from the 1979–2010 ERA-Interim climatology at 1.5°
resolution (Dee et al. 2011) and the control simulation are shown in Figs. 3a and 3b, respectively. The summertime thermal low has developed over the central Sahel and Sahara in the reanalysis (Fig. 3a). The model (Fig. 3b) produces geopotential heights of realistic magnitude, but the low is centered farther to the north and west than in the reanalysis. The position and strength of the North Atlantic subtropical high are realistically simulated, but the onshore flow along the Guinean and west coasts is stronger in the model simulation than in the reanalysis. The West African westerly jet (Grodsky et al. 2003; Pu and Cook 2010, 2011) near 8°N is in place on the west coast in the model, earlier than its formation in the reanalysis.

Figures 3c and 3d show reanalysis and model 925-hPa geopotential heights and winds from July through September. The magnitude of the thermal low is captured accurately in the simulation, but it is located about 5° to the south of its center in the reanalysis. The West African westerly jet has moved a few degrees to the north in

![Fig. 3](image-url)

**Fig. 3.** The 925-hPa June geopotential heights (gpm) and winds (reference wind vectors provided; m s$^{-1}$) from the (a) ERA-Interim climatology and (b) CTL. Also, the 925-hPa geopotential heights (gpm) and winds (reference wind vectors provided; m s$^{-1}$) averaged over July–September from the (c) ERA-Interim climatology and (d) CTL. Contour and shading interval is every 10 m.

In the July–September mean, the Saharan high expands northward over the Sahara in both the reanalysis and the model (Figs. 4c,d) compared with the June average, positioning the African easterly jet a little farther to the north, consistent with the repositioning of geopotential height, surface temperature, and moisture gradients (Cook 1999).
In summary, the regional model produces a state-of-the-art representation of the major circulation and precipitation features of the West African climate and their seasonal evolution.

5. Results

a. The dependence of West African precipitation anomalies on Atlantic Ocean warming

Figure 5a shows July–September precipitation differences between the control simulation and the SSTA 1K simulation in the Atlantic, denoted 1K-CTL. Rainfall rates are enhanced by 2–6 mm day\(^{-1}\) along the Guinean and west coasts to the south of 15\(^{\circ}\)N, and over the western Sahel. There are negative precipitation anomalies up to 2 mm day\(^{-1}\) to the north of the Guinean coast (near 12\(^{\circ}\)N, 5\(^{\circ}\)W) and in the central Sahel (14\(^{\circ}\)N, 10\(^{\circ}\)E). Differences are small to the north of 20\(^{\circ}\)N.

Figure 5b shows summer rainfall anomalies for 1.5K-CTL. Positive rainfall differences in the western and Guinean coasts are further enhanced in 2K-CTL (Fig. 5c). The two small regions of drying in Figs. 5a and 5b are also larger, and the drying in the central Sahel extends to the west. The positive anomalies between 13\(^{\circ}\) and 15\(^{\circ}\)N near the Greenwich meridian in Figs. 5a and 5b are now replaced with negative anomalies.

Negative precipitation anomalies over the central Sahel continue to strengthen and extend westward with additional Atlantic warming. Figures 5d–f display precipitation differences for 2.5K-CTL, 3K-CTL, and 4K-CTL, respectively. The 3K-CTL and 4K-CTL simulations produce a response similar in structure to 2.5K-CTL, but with larger anomalies and more extensive Sahel drying. Rainfall rates over the western and Guinean coasts increase by more than 18 mm day\(^{-1}\) in 2.5K-CTL, for example, and rainfall rates over the Sahel decrease by up to 10 mm day\(^{-1}\) in 3K-CTL and 4K-CTL.
Fig. 5. Precipitation differences from CTL averaged over July–September for simulations with (a) 1-, (b) 1.5-, (c) 2-, (d) 2.5-, (e) 3-, and (f) 4-K warming of the Atlantic Ocean. Contours are drawn for precipitation differences of $-8$, $-2$, $+2$, and $+8$ mm day$^{-1}$. 
Figure 6a shows percentage differences in summer (July–September) precipitation averaged from 12°W to 6°E over West Africa for 1K-CTL. Each bar represents a 1° latitude band, and Sahel latitudes (13°–16°N) are shaded. Rainfall rates increase up to 35% over West Africa between 8° and 10°N. They do not change, or decrease slightly, between 10° and 13°N. To the north of 13°N, precipitation rates increase everywhere. The largest percentage change occurs between 16° and 19°N where rainfall rates increase by 90%–167%.

Figure 6b displays the results for 1.5K-CTL. Between 8° and 16°N, the response is similar to the 1K-CTL differences (Fig. 6a). Rainfall increases are not as large as in 1K-CTL to the north of 17°N.

With 2-K warming in the Atlantic (Fig. 6c), precipitation increases between 8° and 12°N are a little larger than with 1.5-K warming but similar in structure, but there is a sign change in the precipitation anomaly over the Sahel. The 1.5K-CTL has 4%–66% precipitation increases between 12° and 17°N, but Sahel rainfall rates in 2K-CTL decrease by 3%–18%. Percentage precipitation enhancements north of 17°N are greater than in 1K-CTL and 1.5K-CTL.

Figures 6d–f display precipitation differences for 2.5K-CTL, 3K-CTL, and 4K-CTL, respectively. Precipitation differences between 8° and 11°N are positive, from 38% to 140% in 2.5K-CTL and from 45% to 225% in 3K-CTL. The Sahel drying that was relatively modest in 2K-CTL (Fig. 6c) is larger and more extensive with 2.5-K Atlantic warming, with precipitation increases to the north of 17°N replaced by drying. With greater Atlantic warming, the same pattern of drying in the Sahel persists, becoming more severe and spreading southward.

Figure 6 demonstrates that the precipitation response to Atlantic SSTAs over the Sahel is nonlinear. For Atlantic SST increases up to 1.5 K, Sahel rainfall rates increase. When the warming reaches and exceeds 2 K, Sahel rainfall decreases. The existence of a threshold Atlantic SSTA between 1.5 and 2 K in the model complicates climate prediction and suggests a possibility for abrupt climate change over the Sahel in association with SST forcing from the Atlantic basin.

Reference to Fig. 5 indicates that there is some regional dependence for the threshold warming value. In general, rainfall rates increase (decrease) to the west (east) of 6°W from 12° to 19°N in all of the warming simulations. In the Sahel, the rainfall rates increase from 6° to 12°W in 1K and 1.5K SSTA simulations compared with CTL, but decrease with warming of 2 K and higher. In 1K-CTL and 1.5K-CTL, the overall increases are offset by a simulated decrease to the east of 6°W. However, from 17° to 19°N, the decrease to the east of 6°W does not
offset the increase over the western portion of the averaging region until SSTAs exceed 2.5 K. This regional increase in rainfall rates to the west of 6°W is largely confined to the month of August.

Two other recent model simulations have been conducted with uniform SST increases, both with 2-K warming. Held et al. (2005) simulated annual-mean drying in the Sahel, especially from 10° to 15°N, in an atmospheric GCM. Patricola and Cook (2010, 2011) found that July–September (JAS) Sahel rainfall decreased, especially during July, up to 6 mm day\(^{-1}\) in response to uniform Atlantic warming in a regional climate model. These results are consistent with the 2K-CTL presented here.

To focus on the mechanisms of this nonlinear dependence on Atlantic SSTAs, we examine the time period from 26 June to 26 July. This time period is largely responsible for the fact that Sahel JAS rainfall is less in 2K-CTL than in 1K-CTL and 1.5K-CTL (not shown). Average daily precipitation over the Sahel in this period in the CTL simulation is about 5 mm day\(^{-1}\). It increases to about 8 mm day\(^{-1}\) in 1.5K and decreases to about 3 mm day\(^{-1}\) in 2K. At the end of July, rainfall rates in 2K are greater than in CTL. During the rest of the summer (e.g., August and September) rainfall rates in CTL, 1.5K, and 2K are similar.

Figure 7 displays rainfall differences for four of the warming simulations averaged from 26 June to 26 July. As was the case for the full summer average (Figs. 5a,b), the basic spatial structure of rainfall anomalies over the Sahel in 1K-CTL (Fig. 7a) and 1.5K-CTL (Fig. 7b) is similar. Precipitation differences in the southern Sahel (near 13°N) and the eastern Guinean coast region are greater in 1.5K-CTL than in 1K-CTL, as is the rainfall reduction in the central Sahel (along 14°N east of 6°W).

Precipitation decreases in the central Sahel become more pronounced with Atlantic warming of 2 and 2.5 K.
As Atlantic SSTAs are increased above 2.5 K, the drying becomes stronger but the spatial structure is similar to that in 2K-CTL and 2.5K-CTL (not shown). Precipitation over the Gulf of Guinea increases in all the Atlantic warming simulations.

b. The hydrodynamic response to small Atlantic SSTAs: 1K-CTL and 1.5K-CTL

The hydrodynamics of the West African and Sahel precipitation increases with Atlantic warming less than 2 K is examined first, and then compared with the response to SSTAs of 2 K and greater in the following section to understand the sign change in the response.

Figure 8a shows winds and geopotential heights at 925 hPa averaged from 26 June to 26 July from the CTL simulation. The southerly monsoon flow from the Gulf of Guinea flows into the Guinean coast region and turns southwesterly over the continent under the influence of Coriolis accelerations. The West African westerly jet is in place on the west coast centered near 10°N (Grodsky et al. 2003; Pu and Cook 2010). The pronounced meridional gradients that characterize the region (Fig. 3) are strongest between 9° and 17°N.

Difference fields for 1.5K-CTL are shown in Fig. 8b. Here, the differences in 925-hPa geopotential heights are normalized by the domain-averaged difference of 13.4 gpm to highlight changes in gradients. The anomalous geopotential height gradient is positive, indicating a weakening of the negative climatological gradient (Fig. 8a). This weakening is greatest in the Sahel, centered around 15°N, and it is accompanied by a westward wind anomaly of up to 2 ms\(^{-1}\), indicating that the eastward curvature of the flow over the Sahel (Fig. 8a) is reduced. Warming in the Atlantic Ocean—and, in particular, in the Gulf of Guinea—decreases the large-scale meridional geopotential height gradients across West Africa. As a result, the southerly monsoon flow onto the Guinean coast is not diverted to the east as strongly, and it penetrates farther north. This leads to the precipitation increases in the western Sahel in 1K-CTL and 1.5K-CTL, and also to the small region of decreased precipitation in the central Sahel (Figs. 5a,b).

Another consequence of the decreased magnitude of the large-scale meridional geopotential height gradients across the Sahel is a weakening of the African easterly jet. Figure 9 displays the cross section of the difference in the zonal wind averaged from 12°W to 6°E for 1.5K-CTL. Easterly wind anomalies occur below 700 hPa throughout the domain, consistent with Fig. 8b. The positive (westerly) anomalies centered near 13°N and 550 hPa indicate a weakening of the African easterly jet by more than 2 ms\(^{-1}\) in 1.5K-CTL. The African easterly jet is a result of strong coupling between the atmosphere and land surface over the Sahel, and it weakens as a result of geostrophic dynamics (via the thermal wind relation) when meridional temperature gradients below the level of the jet also weaken (Cook 1999). In these simulations there is no shift in the position of the African easterly jet, which is known to be associated with Sahel rainfall variations on interannual time scales (Nicholson and Grist 2001).

The increases in Guinean coast precipitation in response to Atlantic Ocean warming (Figs. 5a,b) are not fully explained by examining differences in the low-level flow in Figs. 8a and 8b. The southerly monsoon flow onto...
the Guinean coast is increased in magnitude somewhat and turned to the northwest by the warming of the Atlantic Ocean, with anomalous convergence along the coast; the West African westerly jet is weaker in 1.5K than in CTL. To fully understand the response on the coast, atmospheric moisture fields must be considered.

Figure 10a displays 925-hPa mixing ratios and moisture transport from the CTL simulation averaged from 26 June to 26 July. Similar to Simmonds et al. (1999), moisture flux is expressed as the sum of the mean and the transient eddy components. The transient eddy component, however, is negligible in this case. While some moisture is transported onto the continent in the southerly monsoon flow, reduced evaporation associated with the relatively cool temperatures of the Atlantic cold tongue lower the moisture content of the southerly flow. A primary source of moisture for the Sahel at this time of year is the West African westerly jet (Pu and Cook 2011). As seen in the difference fields (Fig. 10b), mixing ratio values over the Gulf of Guinea double when 1.5-K warming is imposed in the Atlantic basin. Warming of the sea surface combined with the low-level winds of the monsoon circulation (Fig. 8a) drive enhanced evaporation. The increased moisture content of the inflowing monsoon circulation fuels the Guinean coast precipitation increases. Note also the anomalous westward moisture transport from the central to the western Sahel that is associated with the wind anomalies shown in Fig. 8b.

It may seem counterintuitive that rainfall rates over the Guinean coast increase when Atlantic SSTs are warmed if one thinks about this region’s rainfall as being supported by a monsoon circulation that is ultimately related to the juxtaposition of a cooler ocean surface and a warmer land surface, because warming the ocean would decrease the land–sea temperature contrast in the summer (all other factors remaining the same). This mechanism operates on large space scales and leads to

Figure 9. Latitude–height cross section of the difference in zonal wind speed (1.5K-CTL) for the 26 Jun–26 Jul period averaged from 12°W to 6°E. The contour interval is 1 m s\(^{-1}\) and negative values are shaded.

Figure 10. (a) The 925-hPa horizontal moisture transport (reference vector provided; \(10^{-3}\) g kg\(^{-1}\) m s\(^{-1}\)) and mixing ratios from CTL. (b) Differences in the 925-hPa horizontal moisture transport (reference vector provided; \(10^{-3}\) g kg\(^{-1}\) m s\(^{-1}\)) and mixing ratios for 1.5K-CTL. All values are averaged over 26 Jun–26 Jul. Contour interval is 2 g kg\(^{-1}\) in (a) and 1 g kg\(^{-1}\) in (b).
increases in rainfall in the western Sahel in response to
these mild SSTAs, as discussed above. But along the
Guinean coast, the precipitation increase is explained
through the moisture budget analysis. When Gulf of
Guinea SSTs increase, evaporation rates and low-level
moisture also increase over the SSTAs, but local pre-
cipitation rates are not enhanced because of the pres-
cence of subsidence in this region. As seen in Fig. 11 from
the CTL simulation, and similar to the reanalyses (not
shown), subsiding air over the Gulf of Guinea sup-
presses convection, so the result is an increase in the
moisture flux onto the coast and precipitation increases
to the north of the SSTAs.

A similar mechanism is prominent in the hydrody-
namics of the region’s interannual variability, in which
warm years in the Gulf of Guinea are associated with
enhanced rainfall on the Guinean coast (Vizy and Cook
2002). It is also consistent with the requirements of the
region’s vorticity balance. Cook (1997) showed that the
maintenance of rainfall along the Guinean coast requires
that low planetary vorticity from the south be transported
into the region to balance the positive relative vorticity
tendency associated with midtropospheric condensa-
tional heating and low-level vorticity stretching. Here, we
find that the meridional advection of low relative vorticity
air also contributes to this balance.

c. The hydrodynamic response to stronger Atlantic
SSTAs: 2K-CTL, 2.5K-CTL, 3K-CTL, and 4K-CTL

As seen in Fig. 6, Guinean coast rainfall rates increase
with increased warming in the Atlantic Ocean, but Sahel
rainfall rates reverse the trend and decrease when At-
lanic warming reaches 2 K and higher.

Figure 12 explains why the response over the Sahel is
nonlinear. Increased warming in the Atlantic leads to
further reductions in the large-scale meridional geo-
potential height gradient over West Africa as geo-
potential heights in the Gulf of Guinea decrease. In 2K,
shown in Fig. 12a, the monsoon flow and the West Af-
rican westerly jet continue to impinge on the Guinean
coast supporting rainfall in that region. Farther north,
the meridional geopotential height gradient is weaker,
as seen in the difference fields shown in Fig. 12b, and the
associated easterly wind anomalies over the Sahel be-
come large enough to reverse the direction of the flow.
In 2K-CTL, easterly flow develops in the Sahel east of
5°E. This easterly flow is associated with dry air advec-
tion and precipitation reductions.

With continued warming in the Atlantic—particularly
in the Gulf of Guinea—the easterly full-field flow expands
westward and strengthens as the meridional geopotential
height gradients flatten. In 3K, shown in Figs. 12c, easterly
flow stretches across the western Sahel as a result of very
large easterly anomalies (Fig. 12d).

To understand the mechanisms of the nonlinear re-
sponse, we distinguish between the nonconvective and
convective components of precipitation. Differences
in the nonconvective precipitation among all of the
warming simulations and the control are small (not
shown). The differences in total precipitation, displayed
in Figs. 5a–d, are similar to the differences in the con-
vective precipitation. To connect differences in convec-
tive precipitation with the dynamical fields, we examine
moist static energy (MSE) profiles. The MSE is a linear
combination of the thermal, latent, and geopotential en-
dergy defined as

\[ \text{MSE} = c_p T + Lq + gz, \]

where \( c_p \) is the specific heat of air at constant pressure, \( T \)
is air temperature, \( L \) is the latent heat of vaporization, \( q \)
is specific humidity, \( g \) is the acceleration caused by gravity,
and \( z \) is the geopotential height. A stable atmosphere
will have increasing MSE with elevation. A high value of
MSE at low levels destabilizes the atmosphere.

Figures 13a and 13b display differences in the moist
static energy profiles from the 1.5K-CTL and 2K-CTL
simulations, respectively. The profiles are averaged
from 26 June to 26 July over the region 13°–16°N,
10°W–0°E to emphasize the nonlinear signal in convective
precipitation. Negative slopes of the anomalous MSE
profiles (solid line), especially below 850 hPa, indicate
that the low-level atmosphere is more unstable than the
control in both simulations. The increased instability is
associated with low-level moisture anomalies $Lq$ (dotted line). Atmospheric moisture increases from the surface to 550 hPa in 1.5K-CTL, but it decreases at all levels in 2K-CTL. Consistent with these differences, convective precipitation increases in 1.5K-CTL and decreases in 2K-CTL.

The thermal profiles $c_pT$ (black dashed line) show that there is warming of the entire atmospheric column in 1.5K-CTL and 2K-CTL (Figs. 13a,b). The profile is almost neutral in 1.5K-CTL. On the other hand, there is greater warming at 850 hPa than at the surface in 2K-CTL, and this stabilizes the vertical column. Differences in the geopotential energy $g\zeta$ (gray-dashed line) are negligible in both. Factors associated with increases in precipitation over the Gulf of Guinea and the Guinean coasts in 2K-CTL, 3K-CTL, and 4K-CTL are similar to those in 1.5K-CTL, and there is no nonlinear behavior in this region. To confirm the robustness of the nonlinear response of Sahel rainfall to warming Atlantic SSTAs, a second set of seven simulations for each case was run with a different initialization date in March. The same nonlinear response of the Sahelian rainfall to the warming of the Atlantic was reproduced in the second set of simulations.

6. Summary and conclusions

To contribute to an improvement in our physical understanding of how the West African monsoon will respond under global warming, we isolate one aspect of the problem: the hydrodynamics of the response to a uniform warming of the Atlantic Ocean. Forcing from the Atlantic Ocean is chosen because of the known sensitivity of the monsoon system to Atlantic SSTAs. Idealized simulations with a regional climate model and uniform
Atlantic warming ranging from 1 to 4 K are conducted. The model has 90-km resolution and produces a state-of-the-art simulation of the present-day climate of West Africa, including the precipitation and flow fields.

With 1- and 1.5-K SSTAs prescribed in the Atlantic, rainfall rates increase by 30%–50% over most of West Africa, including both the Guinean coast region (8°–10°N) and the Sahel (13°–16°N). With Atlantic warming of 2 K and higher, precipitation still increases between 8° and 12°N, but Sahel rainfall decreases. These decreases are quite substantial, with approximately 60% reductions with warming of 2.5 K and 85% reductions with warming from the SSTA 4K simulation. This nonlinear response in Sahel rainfall is the focus of the analysis of the model simulations.

Warming less than 2 K in the Atlantic Ocean is accompanied by decreases in low-level geopotential heights in the Gulf of Guinea and, as a consequence, decreases the large-scale meridional geopotential height gradients across West Africa. By a geostrophic argument, this leads to easterly wind anomalies in the central Sahel and the southerly monsoon flow onto the Guinean coast is not diverted as strongly to the east. The moisture from the Gulf of Guinea is able to penetrate farther to the north over the western Sahel, and this leads to precipitation increases in that region. With greater Atlantic warming, the weakening of the continental-scale meridional geopotential height gradient continues, and the easterly anomalies strengthen. With 2-K warming—the threshold value in these simulations—the easterly anomalies reverse the westerly flow over the Sahel. The result is dry air advection into the Sahel and precipitation reductions. A moist static energy analysis shows that the entire atmospheric column warms in all warm SSTAs simulations. The destabilization at low levels that supports convection in 1K-CTL and 1.5K-CTL is because of increases in the low-level moisture content of the atmosphere. Compared with CTL the moisture content of the atmosphere decreases, and the low-level atmosphere is stabilized, with Atlantic warming of 2 K and higher.

In all of the simulations, the southerly monsoon flow onto the Guinean coast is maintained, and precipitation in that region increases. These increases become more closely confined to the coast as the Atlantic warms. The hydrodynamics in this region is dominated by moisture considerations. Warmer Gulf of Guinea SSTs are associated with enhanced evaporation, and the moisture content of the southerly monsoon flow increases. Precipitation anomalies do not develop over the SST anomalies because of the presence of large-scale subsidence over the Gulf of Guinea. Warming even up to 4 K does not break through this subsidence.

Climate change and variability in West Africa and the Sahel is complicated and regional. In this analysis, we focus only on Atlantic SST forcing to develop a deeper understanding of the region’s hydrodynamic response. The application of these results to the global warming problem is limited by this focus. For example, we do not take into account Indian Ocean warming, which is known to be a prominent forcing factor for decadal variability over the Sahel, and we do not increase CO2 levels in these conditions.

![Fig. 13. Average 26 Jun–26 Jul MSE ($\times 10^4$ J kg$^{-1}$) profiles over 13°–16°N, 10°W–0° in (a) 1.5K-CTL and (b) 2K-CTL. The solid, black-dashed, gray-dashed, and dotted lines denote MSE, $c_p T$, gz, and Lq, respectively.](image)
idealized simulations. One could hypothesize that the physical processes that are at the heart of the response analyzed here would be interrupted if increasing greenhouse gas levels cause the Sahara to warm and the continental thermal low to strengthen as the Atlantic warms. However, confident prediction of twenty-first-century climate change in the region requires that we build a solid understanding of the physical processes of that change, and here we contribute to that goal in a process study that isolates the response to uniform Atlantic SSTAs.

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