Contemporary Climate Change of the African Monsoon Systems

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Purpose of Review

Our current understanding of current climate change in the West African, East African, and Congo Basin monsoon systems is reviewed. The detection of observed trends, the analysis of the physical processes of change, and model projections are discussed.

Recent Findings

An increase in Sahel precipitation has been associated with a summer intensification and northward shift of the West African monsoon system as a response to amplified warming over the Sahara. Declines in the boreal spring rains over East Africa, and in spring and summer rains in the Congo Basin, are also reported in the literature, but with less corroboration through physical analysis and model projections. Confident analysis and accurate simulation are hampered by a relative scarcity of observations in these regions.

Summary

The West African monsoon system is trending to bring more precipitation to the Sahel in summer, and some is delivered through increasingly intense rainfall events. It is not yet clear whether recent trends observed in the East African and Congo Basin monsoon systems can be expected to persist as the global climate continues to warm. We cannot expect these monsoon systems to change linearly through the 21st century because, as the ocean basins warm at different rates and with different distributions, different forcing factors may become dominant.

Keywords climate change, monsoons, tropical rainfall, Sahel rainfall, East African precipitation, Congo Basin, climate prediction

I. Introduction

Three regions of Africa are generally described as having monsoon climates, although the historic definition of a seasonal reversal of the low-level wind direction may not apply everywhere. Here we review recent research on climate change in the West African, East African, and Congo Basin monsoon systems. While these regions are termed "monsoon systems", they are not all characterized by one summer rainy season. Rather, as mapped by Hermann and Mohr (2011), some regions have complicated seasonality with multiple wet seasons that may or may not be distinct from each other.

It is easy to motivate the study of climate change in these regions because the populations are among the most vulnerable to climate change, including both in terms of economic loss and mortality. Throughout the continent there is a high dependence on rain-fed agriculture, and many regions have relatively low adaptive capabilities (Busby et al. 2014). The sheer size of Africa also motivates its study - the surface area of the African continent exceeds the sum of the United States, India, China, Japan, and most of Europe, yet the attention paid to the study of African climates is far from proportional. African climate also influences regional and global climate. African wave disturbances influence Atlantic hurricane development (e.g., Dieng et al. 2017) and the Sahara Desert is estimated to emit 70% of the global total of mineral dust aerosols each year (Huneeus et al. 2011) modifying global climate and weather (e.g., Pan et al. 2018), ocean and terrestrial fertilization (e.g., Korte et al. 2017), and carbon sequestration in the deep ocean (Pabortsava et al. 2017).

Each of the three African monsoon climates is discussed below. We review background on each that reflects our current knowledge, and follow this with a discussion of recent work aimed at understanding the potential for – and reality of - climate change in African monsoon systems.

II. The West African Monsoon System

The West African summer monsoon season begins in late April or early May with the onset of spring rains along the Guinean coast near 4°N. The Hovmöller diagram of the 1998-2017 TRMM Multisatellite Precipitation Analysis (TMPA; Huffman et al. 2007) daily precipitation climatology averaged over West Africa between 12°W and 6°E (Fig. 1) shows that precipitation intensifies through May until early June, and remains over the Guinean coast even as the marine Atlantic ITCZ to the west moves farther north (Cook 2015), until late June or early July. At that time, the rainfall maximum transitions into the southern Sahel, near 12°N, often over the course of a few days.

This abrupt shift in the latitude of the rainfall maximum from the Guinean coast into the Sahel is known as the West African monsoon jump (Sultan and Janicot 2000, 2003; Hagos and Cook 2007; Peyrille et al. 2016). As seen in Fig. 1b, which shows the climatological time series of rainfall over the Guinean coast and over the Sahel, the abruptness of the shift in the precipitation maximum is associated with a rapid end of the springtime Guinean coast rainy season – Sahel rainfall builds smoothly from May to a maximum in mid-August. The abrupt end of the spring rains over the Guinean coast is associated with the seasonal movement of the African easterly jet (AEJ), a mid-tropospheric jet that is controlled primarily by surface temperature gradients (Cook 1999; Thorncroft and Blackburn 1999; Wu et al. 2009). The strong negative meridional zonal wind gradient over the coast that is associated with the AEJ preserves an inertially-stable environment until the end of June. When the AEJ moves farther north, these gradients weaken and reverse, satisfying the threshold condition for inertial instability. In the presence of this instability, low-level convergence cannot be maintained, even over a surface temperature maximum, and a rapid demise in the Guinean coast rainy season follows (Cook 2015). The southward progression of the West African monsoon system is smooth because the inertial instability mechanism does not apply for the equatorward movement of the AEJ. An early end of the Sahel rainy season is associated with a strong North Atlantic subtropical anticyclone (NASH) that extends eastward over the Mediterranean and Sahara (Zhang and Cook 2014). The opposite occurs for a late rainfall demise. This is relevant for the 21st century as future projections from the fifth phase of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012) indicate a later demise in summer monsoon Sahel rainfall (Seth et al. 2013; Monerie et al. 2016).

The spring rainy season along the Guinean coast is supported primarily by the meridional convergence of moisture-laden winds from the Gulf of Guinea (Fig. 1c; Sultan and Janicot 2003; Nguyen et al. 2011; Thorncroft et al. 2011; Cook 2015; Schlueter et al. 2019), similar to the hydrodynamics of the marine ITCZ. The Sahel rainy season is supported by moisture from the Gulf and also from the eastern Atlantic by the low-level West African westerly jet (WAWJ). This jet is distinguished from the meridional monsoon flow across the Guinean coast by its structure, physical processes, variations, and connections with precipitation variability (Pu and Cook 2010, 2012; Liu et al. 2019).

On seasonal and sub-seasonal time scales, the Sahara thermal low, or West African heat low, also has an important role in determining summer rainfall distributions over the Sahel (e.g., Lavaysse et al. 2009). During the boreal summer when heating is strongest over the Sahara, the thermal low intensifies, enhancing the low-level transport of moisture over the West African Sahel and southern Sahara needed to fuel convection (e.g., Parker et al. 2005; Lavaysse et al. 2009, 2010, 2016; Liu et al. 2019).

African easterly wave disturbances organize convection over large portions of West Africa. Earlier investigations of these waves suggested that dynamical (shear) instability of the mid-level

African easterly jet (Burpee 1972; Cook 1999) generates them (Rennick 1976; Simmons 1977; Thorncroft and Hoskins 1994; Thorncroft 1995). More recently, however, the strong concentration of diabatic heating across the Sahel has been identified as the fundamental cause of the waves (Hsieh and Cook 2005, 2007, 2008; Hall et al. 2006; Kiladis et al. 2006; Thorncroft et al. 2008). The meridionally-narrow band of diabatic heating across the Sahel creates an environment (i.e., reversed potential vorticity gradients and Charney-Stern instability) that supports wave growth and maintenance. Initial disturbances originate over the central Sahel and East Africa, including over the Darfur Mountains and Ethiopian Highlands (Berry and Thorncroft 2005; Laing et al. 2008). Vertical and meridional wind shears associated with the African easterly jet contribute to the scale and structure of the waves. African easterly waves contribute to the organization of convection on synoptic time scales (Mekonnen et al. 2006; Maranan et al. 2017; Vizy and Cook 2018a, b; Schlueter et al. 2019), and help determine the diurnal cycle of rainfall (Zhang et al. 2016a).

Several papers document observed increases in rainfall amounts and intensity over the West African monsoon region in recent decades, reversing a multi-decadal drying trend from the late 1950s until the late 1980s. For example, shading in Figure 2a displays trends over the 1981-2018 period from Climate Hazards group InfraRed Precipitation with Stations dataset (CHIRPS; Funk et al. 2015b) for the height of the Sahel rainy season, the July-August-September mean, and indicates positive precipitation trends across the Sahel. Studies by Sanogo et al. (2015), Barry et al. (2018), Bicher and Diedhiou (2018), and Nicholson et al. (2018a) report overall positive trends in large-scale precipitation and heavy rainfall days, but the trends are not necessarily monotonic and exhibit some regionality. Panthou et al. (2018) also detect an intensification of rainfall in daily ground-based observations in the Sahel, and they point out that the ongoing recovery of rainfall amounts does not necessarily mean a return to the steadier pre-1950's rainfall regime.

The observed increases in Sahel rainfall amounts and intensity have been associated with greenhouse-gas forcing in observational (Cook and Vizy 2015; Vizy and Cook 2017; Taylor et al. 2017) and model-based (Vizy et al. 2013; Dong and Sutton 2015; Han et al. 2019) analyses. Warming over the Sahara Desert has been proceeding at an amplified pace – approximately triple the rate of the tropical average, similar to the amplified warming of the Arctic (Cook and Vizy 2015). An examination of the surface heat budget indicates that the cause of the amplified warming is primarily the dryness of the surface over the huge expanse of desert, although increases in atmospheric water vapor content outside of the boundary layer play a role seasonally (Vizy and Cook 2017). The resulting increase in the warm-season's positive meridional temperature gradient across West Africa (Fig. 2a) is accelerating the AEJ and enhancing southwesterly flow and moisture transport into the Sahel. Taylor et al. (2017) suggest that the increased wind shear in the Sahel intensifies convention, and increases in atmospheric water vapor may also play a role.

Vizy et al. (2013) find that the Sahara Desert's amplified warming and related changes in the atmospheric hydrodynamics occur both in regional model and CMIP5 AOGCM projections under greenhouse gas-induced climate change, as shown in Figs. 2b-d. The regional model and some of the AOGCMs translate these changes into significant Sahel precipitation increases by the mid-21st c., and all models examined produce these effects by the end of the 21st c.

In addition to this sensitivity to the recent amplified Saharan warming, the West African monsoon system is known to be sensitive to other local and global forcing. For example, the influence of regional and remote SST forcing has been long identified as a primary forcing factor of drought in the Sahel as highlighted in the review paper by Rodríguez-Fonseca et al. (2015), and Lin et al. (2018) and Giannini and Kaplan (2018) also suggest a potential role for external aerosol forcing. West Africa, and the Sahel in particular, is known to experience both short-term (interannual) and long-term (decadal) droughts. The most notable long-term drought occurred during the 1970s – 1980s (Nicholson 2013; Nicholson et al. 2018a). Since this time rainfall over the Sahel has recovered to some extent. Occasional local and regional droughts still occur, and they have become shorter and more frequent (Bicher and Diedhiou 2018; Han et al. 2019). Variations in Atlantic SSTs characterized by warming in the equatorial Gulf of Guinea and cooling off the West African coast north of 10°N are associated with a rainfall dipole over West Africa with drying over the Sahel and enhanced rainfall along the Guinean Coast (Ward 1998; Vizy and Cook 2002; Polo et al. 2008). Mediterranean SST anomalies also influence West African rainfall variability, with warm SSTs associated with enhanced Sahel rainfall as the West African monsoon shifts further north and intensifies (Rowell 2003; Fontaine et al. 2010).

Pacific and Indian Ocean SST anomalies can also impact rainfall variations over the Sahel and West Africa. In some studies, warm tropical Pacific SSTs are associated with subsidence and reduced rainfall over the Sahel during the peak of the summer monsoon (Mohino et al. 2011). Indian Ocean warming has also been tied to Sahel drought, especially in the 1980s (Biasutti 2008; Bader and Latif 2011; Fontaine et al. 2011), but Sahel rainfall has recovered since the 1990s while Indian Ocean SSTs continue to warm. The response may depend on such factors as the spatial scale of the Indian Ocean SSTs and interactions with SST forcing from other ocean basins (Suárez-Moreno et al. 2018) including the Atlantic (Hagos and Cook 2008; Dyer et al. 2017a; Kamae et al. 2017).

Effects of changes in land-surface conditions are also discussed in the literature as a mechanism for climate change over West Africa. Land cover over West Africa is changing as more land is cultivated to support a growing population (Taylor et al. 2002; Wang et al. 2016). These changes modify surface roughness and albedo, and can impact surface water distributions and aerosol loading, affecting energy, momentum, and moisture exchanges between the surface and atmosphere. In the 1960's and 70's, concern over declining rainfall in the Sahel led to wide-spread acceptance of the so-called "Charney mechanism", in which deforestation in the Sahel was thought to increase surface albedo and atmospheric stability, thereby decreasing rainfall (Charney 1975). This idea has been proven incorrect – the sensitivity arises in large part from a strong boundary constraint in the original simple model used - and replaced by an understanding of the dominate role of SST forcing on decadal time scales. This is not to say that West African rainfall is completely insensitive to land-surface conditions. However, large-scale surface albedo changes associated with human activity in West Africa are small and soil moisture feedbacks provide a stronger forcing mechanism (Koster et al. 2006; Patricola and Cook 2008). In a recent review, Spracklen et al. (2018) note the complexity of the issue, with deforestation leading to precipitation reduction over the devegetated areas.

The Sahara is the largest source at atmospheric mineral dust, and is thought to suppress severe storm activity in the tropical Atlantic and Caribbean, the southern U.S., and Europe as well as within the West African monsoon (e.g., Lau et al. 2009; Chen et al. 2010; Knippertz et al. 2015, 2017; Bretl et al. 2015; Bahino et al. 2018; Pan et al. 2018). In addition to the Saharan dust, rapid socioeconomic development in southern West Africa is increasing atmospheric loadings of atmospheric pollutants, including particulate aerosols volatile organic particles (VOCs), with the potential to modify the monsoon system through direct radiative effects and cloud interactions (Deetz et al. 2018, Keita et al. 2018). However, even with high resolution and the inclusion of explicit convection, climate and forecast models perform poorly in simulating African dust lifting (Chaboureau et al. 2016), for example, in association with cold-pool outflows (haboobs) from the numerous warm-season mesoscale convective systems (MCSs) that deliver West African rainfall (e.g., Roberts et al. 2018) and low-level jets (Allen and Washington 2014).

With so many potentially potent forcing factors in play over such a large area with nonlinear dynamics, it is probable that changes within the West African monsoon system as the global climate warms will be non-linear (e.g., Neupane and Cook 2013) and regional. To support prediction, there is a pressing need to continue progress on understanding the basic hydrodynamics of the West African monsoon and the factors that cause this complex system to vary. Modeling at convective-permitting (CP) resolutions (< 4 km) offers promise. Unlike GCMs, CP models are able to capture the diurnal cycle of rainfall over West Africa (Zhang et al. 2016b), which requires that both locally-generated and propagating mesoscale complexes are simulated accurately (Vizy and Cook 2180a, b).

III. The East African Monsoon System

East Africa is a large area that extends some 40 degrees of latitude, from the Horn of Africa (Ethiopia, Eritrea, Djibouti, and Somalia), through the equatorial region (Kenya, Uganda, Rwanda, Burundi and Tanzania), to Mozambique and Malawi in the south. It is home to many of the most food-insecure nations in the world according to the Food and Agriculture Organization of the United Nations (FAO 2015) due to a dependence on rain-fed agriculture, low adaptive capability, and the potential for climate-driven conflict. At least a portion of this insecurity derives from a vulnerability to climate variability and change (O'Loughlin et al. 2012; Busby et al. 2014). Climate is often viewed as a "threat multiplier" that is superimposed on other stressors in the region.

Seasonality of the East African monsoon system is highly regional and complex. East African rainfall is often described as having two rainy reasons – the "long rains" in boreal spring and the “short rains” in boreal fall - that are related to “two passes” of the intertropical convergence zone (ITCZ), but this is an over-simplification (e.g., Nicolson 2018). As shown by the characterization of observed East African rainy seasons by Hermann and Mohr (2011), while some regions of East Africa, for example, south-central Ethiopia, southern Somalia, and most of Kenya, experience two, unimodal wet seasons, each with single peak, the remainder of Ethiopia, all of Tanzania, and much of Southern Hemisphere East Africa (e.g., Malawi and Mozambique) experience a single wet season in summer, sometimes with a mid-season break in precipitation which is classified as a “bimodal regime”. The lack of zonal uniformity in the seasonal characteristics of the East African monsoon and the presence of abrupt changes in the location of rainfall maxima (Riddle and Cook 2008) indicates that the region’s rainfall cannot simply be understood as the smooth movement of an ITCZ.

Many factors have been shown to influence East African rainfall, depending on location and season. These factors include the region’s complicated topography and proximity to the Indian Ocean, and sensitivity to global-scale modes of variability such as ENSO (e.g., Dunning et al. 2016). However, an understanding of the seasonality and variability of the highly regional precipitation regimes is far from complete. As shown in Figure 3, there are significant seasonal variations in the large-scale circulation (and moisture transports), including a reversal in wind direction in DJF compared with the rest of the year. During the transition seasons (Figs. 3b and d), on-shore flow brings moisture from the Indian Ocean to Southern Hemisphere East Africa, while flow through the orographic gap known as the Turkana Channel (5ºN, 36ºE) and southerly low-level flow channeled along the coast in part by the topography support Northern Hemisphere precipitation. Due to warm Indian Ocean SSTs, the associated onshore transport of high moist static energy is sufficiently large to overcome the divergence of the overlying air and produce a rainy season over equatorial East Africa (Yang et al. 2015). Some of the large-scale moisture transport patterns at 850 hPa that characterize boreal spring persist into the summer (Fig. 3c), with added features due to the development of the West African summer monsoon system and the development of the zonal branch of the Somali jet across the Arabian Sea supporting the single rainy season of northern Ethiopia. The northeasterly flow of the Asian winter monsoon dominates the circulation in DJF (Fig. 3a).

Because of this regional and seasonal complexity, evaluating East African rainfall trends is not straightforward. Figure 4 shows the 1981-2018 CHIRPS linear precipitation trends over East Africa for three seasons. CHIRPS incorporates 0.05° resolution satellite imagery with in-situ station data to create gridded rainfall time series over land for trend analysis. We choose to show it here because high spatial resolution is advantageous over the complex East African topography, and the time series is relatively long (1981-present).

During boreal spring (Fig. 4a), drying trends with significance at the 90% level are evident over central Somalia and the Ethiopian Highlands, with increasing precipitation trends around Lake Victoria, west of the Ethiopian Highlands over Sudan and South Sudan, and eastern Tanzania. Regions with weaker positive and negative trends are scattered.

In boreal summer (Fig. 4b), trends are small and incoherent over most of equatorial East Africa where there is little rainfall, but positive trends associated with the Sahel rainfall recovery occur over northern Ethiopia and the Sudan. Positive rainfall trends near Lake Victoria persist through the summer and into the boreal fall (Fig. 4c), when the wetting trends extend over the southern Ethiopian Highlands and the Turkana channel region, consistent with the gauge-based analysis by Schmocker et al. (2016) for Mount Kenya. Farther south, there is evidence of significant drying trends over southern Tanzania, northern Mozambique, and isolated areas of Madagascar.

Recent impactful droughts (including the regional-scale drought of 2011 and another centered in Ethiopia in 2015) heighten the concern about potential climatological precipitation declines over East Africa. A number of recent papers evaluate East African rainfall trends, usually for approximately the 1980-2010 time period for the spring long rains, and report negative trends averaged across large areas (e.g., Lyon and DeWitt 2012; Funk et al. 2015a; Maidment et al. 2015). There is especially a tendency for drying in the first decade of the 21st c. reported in the above studies, although the exact location varies in different studies. For example, Ongoma and Chen (2017) find a reduction in long rains (MAM) over Tanzania and Lake Victoria area in the 2001-2010 decade, while Funk et al. (2015a) report reductions in the long rains (MAMJ) over northern Tanzania, northern Kenya, and southeastern Ethiopia. These findings contrast with the results shown in Fig. 4a, in which the spring drying is restricted to Ethiopia and Somalia. Analyses of the trends associates the changes in rainfall with Indian Ocean SSTs and Walker circulation dynamics, either due to natural variability or greenhouse gas increases (Williams and Funk 2011; Liebmann et al. 2014; Yang et al. 2014), and western Pacific SSTs and/or the Pacific Decadal Oscillation (Omondi et al. 2013, Yang et al. 2014, Funk et al. 2015c)

A trend in extreme events, including dry periods and intense rainfall, may be starting to emerge over East Africa in observational analyses, but a clear picture has not yet formed. While Grebrechorkos et al. (2019) do not find significant trends in rainfall extremes from 1981-2016 in Ethiopia, Kenya, and Tanzania, Gummandi et al. (2018) report positive trends in precipitation intensity over southern Ethiopia during both the spring and fall rainy seasons during a similar period (1980-2010). These trends are accompanied by an increase in consecutive dry days as precipitating systems intensify but become less frequent.

Projections of climate change over East Africa are hampered by the poor representation of the rainy seasons in coupled GCMs. In Figs. 5a and, the black lines show the satellite-observed TRMM rainfall climatology over a region in equatorial East Africa (primarily Kenya), with the long rains in March, April and May and the short rains in October and November depicted. The blue lines are rainfall averaged over the same region and 1986-2005 in two coupled GCMs representative of the CMIP5 archive. In both models, the long rains are extremely weak, and the short rains are too strong (see also Otieno and Anya 2013; Yang et al. 2015b; Rowell et al. 2015). These inaccuracies in the simulations are, in some large part, associated with the ocean component of the coupled GCMs – atmosphere only versions of the models with SSTs prescribed from observations – produce much better simulations (purple lines in Figs. 5a and b). As seen in Fig. 5c, the 30-km, atmosphere-only regional model ensemble simulations of the late 20th c. described in Vizy et al. (2013), also produce credible simulations of the East African rainy seasons.

Hirons and Turner (2018) relate the simulations of too-wet short rains in coupled GCMs to mean-state biases in coupled GCMs, with easterly wind biases amplifying upwelling in the eastern Indian Ocean and boreal fall precipitation. Yang et al. (2015b) compare coupled and atmosphere-only versions of the MRI GCM and relate an improved simulation by the atmosphere-only model to improvements in the low-level MSE and convective instability, ultimately related to having a better representation of western Indian Ocean SSTs. Rowell and Chadwick (2018) associate uncertainty in projections of East African rainfall with uncertainties in the regional responses to projected SST warming in coupled GCMs. Strong influence from Indian Ocean SSTs on East African rainfall is a major cause of the coupled models' rainfall inaccuracy, including the ability to produce prominent modes of Indian Ocean SST variability (Weller and Cai 2013, Du et al. 2013) and connections to ENSO (Ha et al. 2017).

Ummenhofer et al. (2018) shed some light on whether the poor simulation of the long rains in a coupled GCM indicates that the model will be unable to accurately estimate changes in rainfall due to greenhouse gas forcing (via SST anomalies) by examining observed modes of variability and their connection to Indo-Pacific SSTs in a 1300-year simulation. They find that both the long rains and the short rains respond to ENSO-like SSTAs, in contrast to observations in which the only sensitivity is during the short rains. On decadal time scales, East African rainfall anomalies were found to be responsive to Indian Ocean and Indo-Pacific, and not eastern Pacific, SSTA variations, but these sensitivities were not well represented in the model.

The design of future simulations in a regional model skirts this problem to some extent by generating future SSTs by adding coupled GCM-generated anomalies to observed present day SSTs. This methodology depends on the ability of GCMs to project SST anomalies, but not on their ability to produce a correct current SST distribution. The approach was first developed and tested in paleoclimate applications by Cook and Vizy (2006) and Patricola and Cook (2007). Under the RCP8.5 greenhouse-gas emissions scenario, ensemble simulations of the last 2 decades of the 21st c. at 90-km (Cook and Vizy 2012, 2013) and 30-km resolution (as in Vizy et al 2013) indicate little change in the long rains along the equator, and a strengthening of the short rains (red line in Fig. 5c). Elsewhere in East Africa, the simulations project a weakening of the long rains over Ethiopia and Somalia (similar to Fig. 4a) in connection with a strengthening of the Saharan and Arabian highs, which has been related to amplified surface warming in these desert regions (Cook and Vizy 2015; Vizy and Cook 2017).

As for the West African monsoon, CP-modeling holds promise for improving our ability to simulate the East African monsoon system. Along with the usual benefits of explicitly resolving convective processes in the governing equations rather than through parameterization, an added benefit for East Africa is the accurate portrayal of the region's complex and influential topography. More broadly, Finney et al. (2019) evaluate the benefits of CP modeling for the water budget across East Africa and show that there are widespread improvements in rainfall intensity and diurnal cycle in comparison with a coarser-resolution version of a similar model.

IV. Congo Basin

The monsoon system of equatorial western and central Africa, which we refer to here as the Congo Basin region for efficiency, supports a tropical forest that is second in size to the Amazon rainforest and the Congo River watershed which is the second-largest river basin on the planet. The Congo River Basin, which is roughly equivalent to the monsoon region, covers 3.4 x 10^6 km^2 (nearly half the surface area of the contiguous U.S.) in central and western equatorial Africa, including the Republic of Congo, the Democratic Republic of Congo, and the Central African Republic, as well as portions of Tanzania, Cameroon, Angola, and Zambia. The Congo Basin deforestation rate is estimated at 0.17% per year for 2000-2005, with additional degradation (Ernst et al. 2012). This is much smaller than the deforestation rates in the Amazon, Central America, and Southeast Asia forests, but it is nearly double the rate for the 1990-2000 period in the Congo Basin and may change significantly in the future (Dargie et al. 2018).

Figure 4 displays a satellite-based estimate of the seasonal excursion of precipitation averaged from 10ºE to 30ºE. Within 5º of latitude from the equator, precipitation persists for 10 months or more of the year. A broad swath of strong (> 5 mm/day) precipitation progresses smoothly from 10ºS to 10ºN from January through early August. After August, maximum rainfall rates increase significantly, and the southward migration of the rainfall maximum occurs over 4 months.

Through most of the year, with the exception of boreal fall (DJF), southeasterly flow onto the African continent from the South Indian Ocean transports large amounts of moisture into the Congo Basin (Fig. 3). In JJA (Figs. 3c and d), this southeasterly flow is part of the full Somali jet that consists of a southerly, cross-equatorial component and a zonal branch extending across the Arabian Sea to India. At the 850 hPa level shown in Fig. 3, which clears much of the topography as resolved in ERAI, a seasonal reversal of the meridional flow over the Congo Basin occurs.

An additional source of moisture for the Congo Basin is low-level westerlies from the equatorial Atlantic, which are confined below 900 hPa (Pokam et al. 2012, Dezfuli and Nicholson 2013). Schwendike et al. (2014) and Pokam et al. (2014) discuss the existence of an over-turning circulation, with a rising branch over the Congo Basin associated with descent over the equatorial eastern Atlantic. Cook and Vizy (2016) show that the strength of this Congo Basin Walker circulation is not correlated with land surface temperatures on seasonal time scales, and it only exists in June through October when the Atlantic cold tongue is in place. Even then, the diabatic heating maximum over the Congo Basin is primarily supported by inflow from the Indian Ocean (see Dyer et al. 2017b), and the Congo Basin Walker circulation is not a closed system.

Even more than the other monsoon regions of Africa, the study of the Congo Basin monsoon system is hampered by a lack of ground-based measurements of meteorological variables. The number of reporting stations has declined over recent decades, especially since the 1980s in association with economic declines (Washington et al. 2013; Nicholson et al. 2018b). This decrease hampers the evaluation and calibration of satellite-based observations, and can introduce spurious trends (Maidment et al. 2015).

Mahli and Wright (2004) first raised concerns about a possible decline in Congo Basin rainfall in analyzing CRU rain gauge data. Zhou et al. (2014) also suggested that precipitation declines accompanied forest browning in boreal spring (April-May-June; AMJ) over the 2000-2012 period. Maidment et al. (2015) examine trends in eight precipitation datasets for 1983-2010 and find disagreement among them over Central Africa, suggesting that spurious negative trends associated with declines in gauge density occur in some datasets. Nicholson et al. (2018b) produced gridded monthly-mean datasets at 2.5º and 5º resolution from gauge data for 1921-2014, and examined AMJ trends in 14 sub-regions of the Congo Basin. Positive trends that are significant at the 90th percentile are found in the northern portion (5ºN – 10ºN) of their analysis region for the April, May, and June average, with negative trends elsewhere (10ºS – 5ºN), half of which are significant.

Table 1, adapted from Cook et al. (2019), provides an overview of annual mean and seasonal precipitation trends averaged over the entire Congo Basin for two time periods from six datasets that are either satellite derived or based on blending satellite and gauge data. For the annual mean over 1979-2017, all four available datasets produce negative trends, but only one is statistically significant. For the six datasets analyzed for 1998-2017, all but one are negative, with a highly significant negative trend in CHIRPS. Taken together, these data suggest annual-mean Congo Basin drying, but confidence is not especially high.

Seasonal rainfall trends are also shown in Table 1. In DJF, negative trends predominate for 1998-2017, but only CMORPH indicates high confidence. In boreal spring, represented as MAM, negative linear trends that emerge weakly over the 1979-2017 period are not representative of the 1998-2017 period, when positive precipitation trends (if anything) occur. (The result is similar when boreal spring is represented by AMJ; negative trends for these months are more significant for the 1979-2017 period, but also disappear for 1998-2017.) Negative trends emerge most strongly, especially for the JJA period, during boreal summer. Cook et al. (2019) associate these trends, and the somewhat less certain DJF trends, with poleward shifts in the thermal lows over Africa. Given the findings of Guan et al. (2015) that the annual rainfall in the Congo Basin is barely sufficient for maintaining the tropical forest, these reductions in rainfall even during the less-rainy seasons may be critical. This is especially true if the poleward shifts of the thermal lows will be continuing due to increasing greenhouse gas concentrations (as part of the observed tropical expansion) and/or the recovery of the ozone hole in the Southern Hemisphere, or if possible boreal spring rainfall reductions are related to warming SSTs (Hua et al. 2016).

Additional changes in Congo Basin rainfall are emerging in observations, especially for extreme events. The Congo Basin climate is characterized by exceptionally strong MCS activity. Taylor et al. (2018) find significant increases in intense rainfall events associated with MCS activity since 1999 in February, suggesting an earlier onset of the boreal spring rainy season in association with increasing meridional surface temperature gradients across the continent.

Large discrepancies in precipitation climatologies over the Congo Basin occur in the current generation of coupled GCMs, and even the ensemble mean of CMIP5 historic simulations likely does not produce a best estimate (Creese and Washington 2016). Discrepancies between observed and modeled SSTs, especially in the southern Indian Ocean and the equatorial Atlantic, may be related to the models' poor simulation of Congo Basin rainfall distributions, exacerbated by the scarcity of observations with which to compare. However, even regional simulations at 30 km resolution with observed SSTs significantly over-produce Congo Basin rainfall in spring and fall. In these simulations, the frequency of rainfall events is well-simulated, but precipitation rates on wet days is too high (Han et al. 2019).

Adopting a more physically-based approach that emphasizes the transport and convergence of moisture into the Congo Basin, and downplays the models' ability to translate this moisture into precipitation using parameterizations, may provide leverage for extracting information about Congo Basin climate change from AOGCMs. Creese and Washington (2018) explain and adopt this approach for the boreal fall rainy season. During most of the year, models tend to have either wet or dry biases throughout the basin. In boreal fall, however, biases in rainfall distributions between the eastern and western basins are uncorrelated and this suggests associations with physical processes may dominate. For example, they find that models with wet biases in the western Congo Basin have unrealistically-warm SSTs in the eastern tropical Atlantic; models with this characteristic may be less credible for predicting climate change in the Congo Basin.

The Congo Basin monsoon system is the least well-understood of the three African monsoon systems reviewed here. Discrepancies in precipitation climatologies over the Congo Basin produced by coupled GCMs and observational deficiency combine to raise considerable uncertainty about climate change trends and future projections for the region and call for more extensive study of this critical region.

V. Concluding Remarks

The school-book characterization of a monsoon climate is a region with strong seasonal rainfall variations that are associated with land/sea contrasts and a seasonal reversal of winds. This concept has led to suggestions that monsoon systems will intensify with increasing atmospheric greenhouse gas levels as the land warms more quickly than the ocean. However, this simple construct neglects the complexity and individuality of the world's monsoon systems in general, and the African monsoon systems in particular. This review of the potential for climate change in the African monsoons has emphasized the need for careful and in-depth regional study that is based in the physical analysis of these complicated systems.

There is strong corroboration from observations and modeling, combined with a physical understanding of the processes, that amplified warming over the Sahara due to increasing greenhouse gas levels is contributing to positive precipitation trends over the Sahel. Indian Ocean warming is likely playing a supportive role as well. These increases represent an intensification and northward shift of the West African monsoon system during boreal summer. However, this does not mean that we can be sure that this positive trend will continue through the 21st century. The sensitivity of the West African monsoon system, including its association with rainfall along the Guinean Coast as well as over the Sahel, to SSTs is well known. For example, from analysis on interannual time scales, we know that warming SSTs in the Gulf of Guinea in isolation are associated with increased rainfall along the coast and reduced rainfall over temperature Sahel. Consequently, as the ocean basins warm at different rates and with different distributions over the 21st century, other forcing factors may become dominant over the amplified-Sahara-warming mechanism.

Over East Africa, several papers report a decline in the historical long rains, but results depend on the exact analysis periods and averaging regions. The precipitation decreases are especially strong in the first decade of the 21st c., but the presence of an ongoing trend has not been established. This monsoon system is characterized by extremely strong regionality associated with its complex topography and geographical location. Regional rainfall has strong coupling with the Indian Ocean, which provides much of the moisture advected into East Africa, but the mechanisms that control the convergence of that moisture and atmospheric stability are not well known.

The monsoon system that supplies moisture to the Congo Basin also exhibits strong coupling with the Indian Ocean. Reports of boreal spring drying are not yet certain, but they are of great concern since the region's tropical forests are thought to receive marginal rainfall amounts. Poleward shifts of the African continental thermal lows in both hemispheres can lead to Congo Basin drying since the equatorward flow to the east of the lows supports moisture convergence over central and eastern equatorial Africa. Improved understanding of the region hinges on improvements in both observations and modeling.

Perhaps more than other regions of the world, African monsoon systems are dependent on SSTs in seasonally- and regionally-dependent ways. This dependence suggests that we may not be able to fully understand climate change in these regions without fully-confident projections of SST distributions, especially in the Atlantic and Indian Oceans. However, there are also local processes to consider that are currently not well understood, including regional circulation systems, connections among the African monsoon systems, and the role of complex topography, not to mention the MCSs that deliver up to 80% of the rainfall and control impactful intense events and the diurnal cycle of rainfall.

With so many potentially potent forcing factors in play over such a large area, it is probable that changes within the African monsoon systems as the global climate warms will be non-linear and highly regional. To support prediction, there is a pressing need to continue progress on understanding the basic hydrodynamics of the African monsoon systems and the factors that cause these complex systems to vary on different time scales.

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Table 1. Linear slopes of Congo Basin (10ºE-30ºE, 5ºS-5ºN) precipitation with confidence levels that account for autocorrelation using the method of Zwiers and von Storch (1995). Asterisks indicate significant drying at the 90th (*), 95th (**), and 99th (***) percentiles.

Dataset	Annual		DJF		MAM		JJA		SON	
	1979-2017	1998-2017	1979-2017	1998-2017	1979-2017	1998-2017	1979-2017	1998-2017	1979-2017	1998-2017
TRMM	----	-0.114	----	-0.093	----	0.024	----	**-0.467****	----	-0.001
GPCP	-0.015	-0.035	0.093	0.058	-0.083	0.116	-0.099	**-0.420****	0.055	0.064
CHIRPS	**-0.061****	-0.089	-0.059	-0.246	-0.027	0.293*	-0.04	**-0.280***	-0.113	-0.157
ARC2	-0.147	0.065	-0.139	-0.065	-0.137	0.576***	-0.11	-0.162	**-0.244***	-0.052
CMORPH	----	**-0.516*****	----	**-0.879*****	----	-0.07	----	**-0.723*****	----	**-0.531****
PERSIANN	-0.027	-0.065	0.053	-0.041	-0.118	0.11	-0.113	**-0.379****	0.043	-0.003

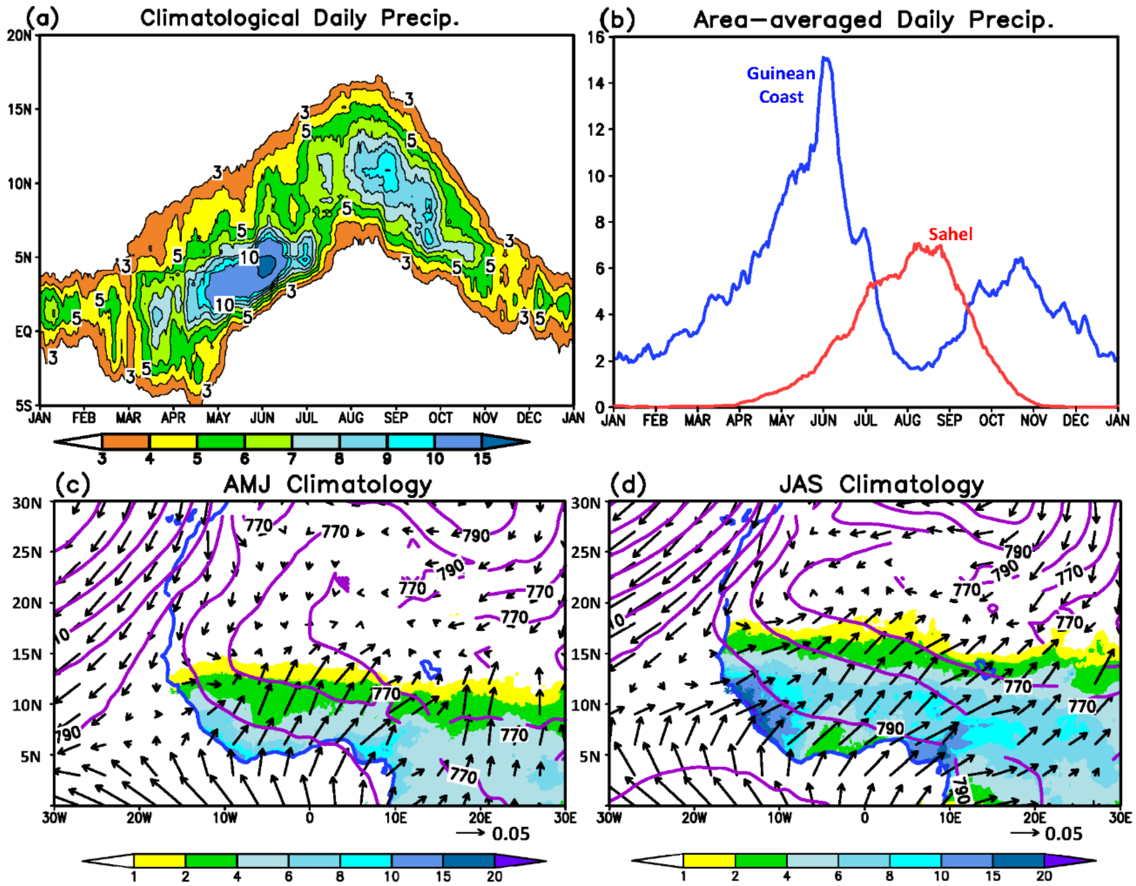

Figure 1. (a) 1998-2017 climatological daily TRMM TMPA precipitation (mm day^{-1}) averaged over West Africa between 12°W - 6°E and smoothed using an 11-day running mean filter. (b) 1998-2017 climatological smoothed daily TRMM TMPA precipitation averaged over the Guinean Coast (3°N - 6°N; 12°W - 6°E) and Sahel (12°N - 15°N; 12°W - 6°E) regions. 1981-2018 climatological (c) April – June (AMJ), and (d) July – September (JAS) CHIRPS precipitation (shaded; mm day^{-1}) and ERAI reanalysis 925 hPa geopotential heights (contours; m) and moisture transport vectors (kg m kg^{-1} s^{-1}).

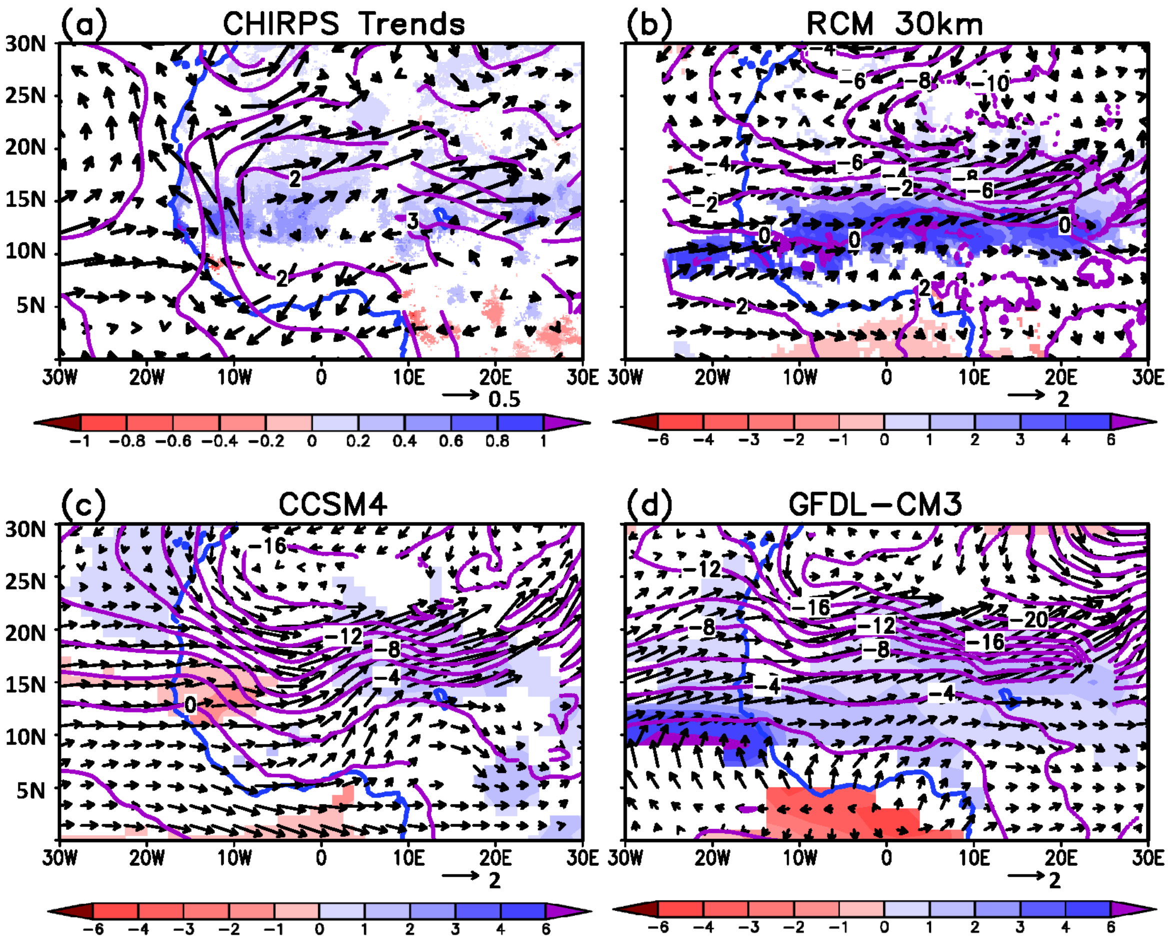

Figure 2. (a) CHIRPS V2.0 1981-2018 July – September precipitation trends (shaded; mm day^{-1} per decade) with 925-hPa geopotential height (m per decade), and wind (m s^{-1} per decade) trends from the ERAI reanalysis. Only precipitation trends statistically significant at the 90% level of confidence according to a Student's *t*-test are shaded. Late 21st century (2081-2100) July – September precipitation (shaded; mm day^{-1}), 925 hPa scaled geopotential height (contours; m) and winds (vectors; m s^{-1}) anomaly projections for the (b) Vizy et al. (2013) 30-km RCM simulation, and the CMIP5 RCP8.5 (c) CCSM4, and (d) GFDL CM3. Only precipitation trends and anomalies statistically significant at the 95% level of confidence according to a Student's *t* test are shaded.

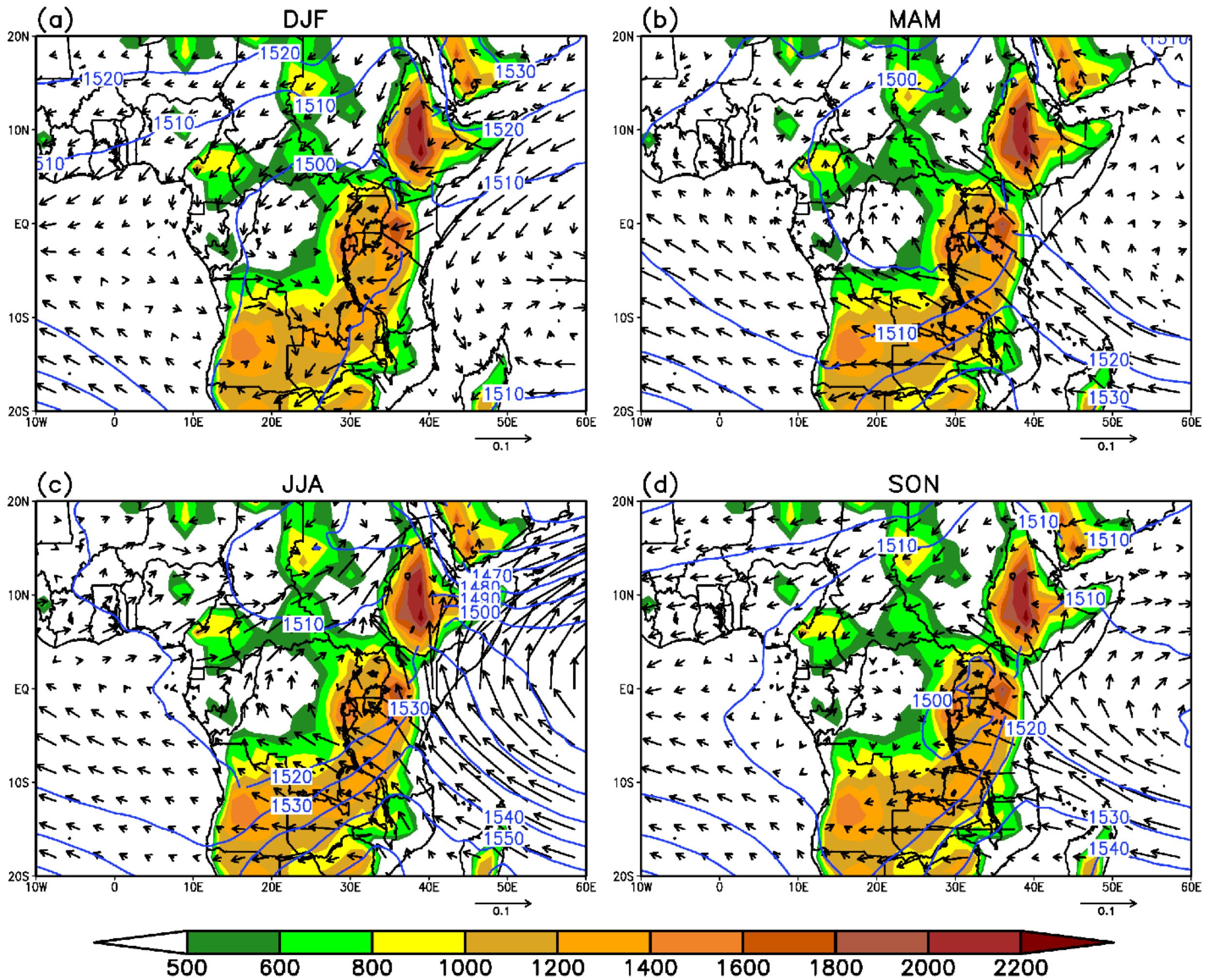

Figure 3. ERAI 1981-2018 climatological 850 hPa moisture flux transport (vectors; kg m kg^{-1} s^{-1}) and geopotential height (contours; m) for (a) December – February (DJF) (b) March – May (MAM), (c) June – August (JJA), and (d) September – November (SON). Shading denotes the topography (m) as resolved in the ERAI at 1.5° resolution.

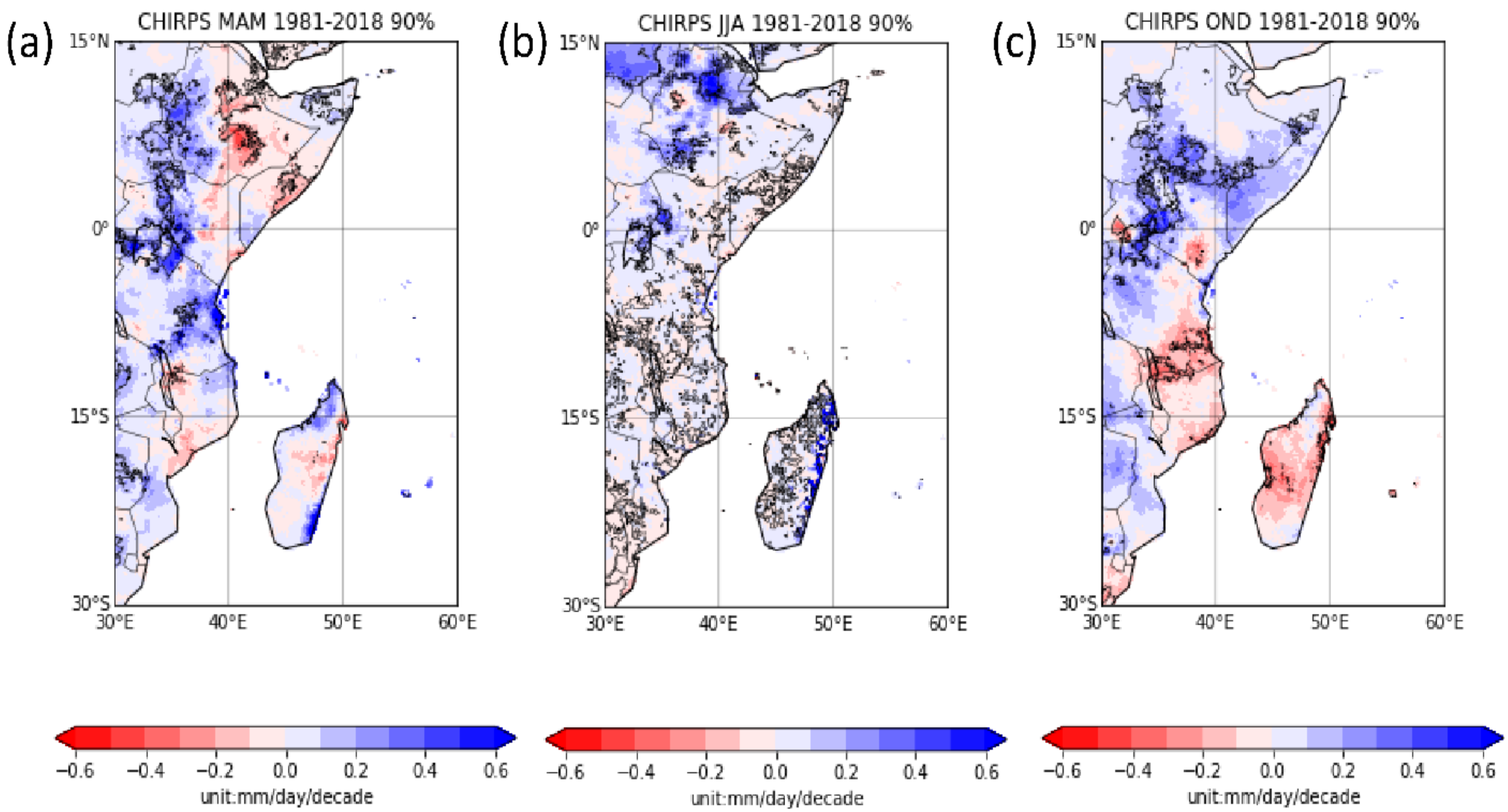

Figure 4. CHIRPS V2.0 1981-2018 precipitation trends (shaded; mm day-1 per decade) for (a) March-May (MAM), (b) June – August (JJA), and (c) October – December (OND). Only precipitation trends and anomalies statistically significant at the 95% level of confidence according to a Student's t test are shaded. (Figure courtesy Siyu Zhao.)

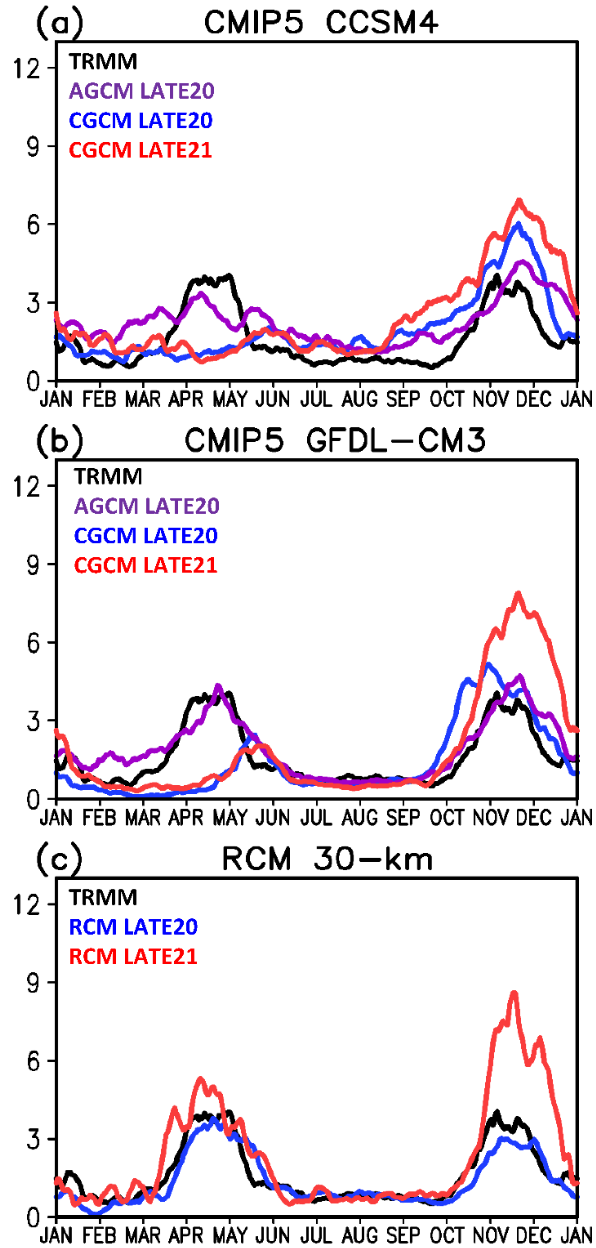

Figure 5. (a) Daily climatological precipitation (mm day^{-1}) averaged over 35°E - 42°E and 3°S - 3°N and 1986-2005 from atmosphere-only (AGCM LATE20; purple) and coupled (CGCM LATE20; blue) versions of the CCSM4 GCM. The black line denotes climatological TRMM TMPA precipitation observations, and the red line is precipitation from CGCM simulations for 2081-2100 (CGCM LATE21) with greenhouse gas increases from the RCP8.5 emissions scenario. (b) Same as (a) but using output from GFDL-CM3 GCM simulations. (c) Same as (a) but using output from atmosphere-only regional model simulations with 30-km horizontal resolution (Vizy et al. 2013).

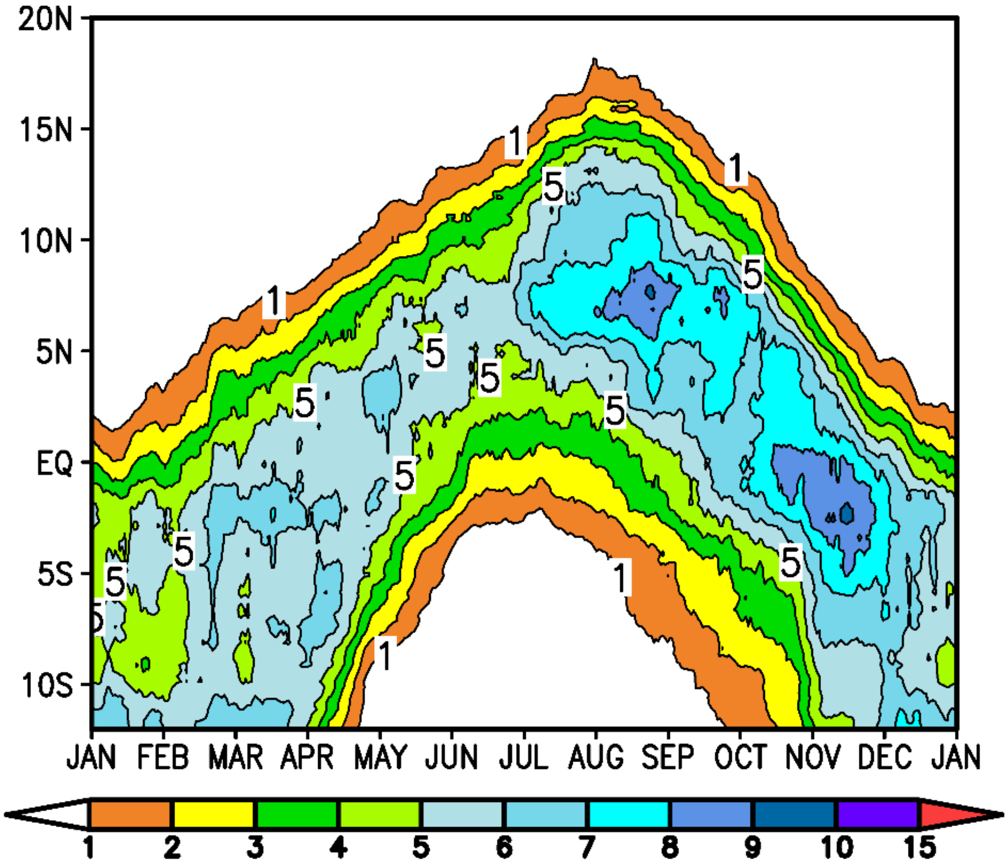

Figure 6. 1998-2018 climatological daily TRMM TMPA precipitation (mm day^{-1}) averaged over the Congo Basin/central Africa between 10°E - 30°E and smoothed using an 11-day running mean.