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Abstract: In this study, the combined effect of the Atlantic Multidecadal Oscillation (AMO) and El Niño Southern Oscillation (ENSO) on the Lake Chad (LC) level variability is explored. Our results show that the lake level at the Bol monitoring station has a statistically significant correlation with precipitation ($R^2 = 0.6$, at the 99.5% confidence level). The period between the late 1960s and early 1970s marked a turning point in the response of the regional rainfall to climatic drivers, thereby severely affecting the LC level. Our results also suggest that the negative impact of the cold phase of AMO on Sahel precipitation masks and supersedes the positive effect of La Niña in the early the 1970s. The drop in the size of LC level from 282.5 m in the early 1960s to about 278.1 m in 1983/1984 was the largest to occur within the period of study (1900–2010) and coincides with the combined cold phase of AMO and strong El Niño phase of ENSO. Further analyses show that the current warm phase of AMO and increasing La Niña episodes appear to be playing a major role in the increased precipitation in the Sahel region. The LC level is responding to this increase in precipitation by a gradual recovery, though it is still below the levels of the 1960s. This

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Churchill Okonkwo

Churchill Okonkwo earned his MS degrees in Geology and Environmental sciences from the Federal University of Technology, Owerri, Nigeria (2005) and the American University, Washington DC (2010), respectively, and his PhD degree in Atmospheric Sciences from Howard University, Washington DC. His doctoral dissertation was on the understanding of teleconnections between climate indices and precipitation in the Sahel region and their impacts on the Lake Chad level variability. He is currently working with the Washington DC Water and Sewer Authority and is also the director of the African Center for Climate Science and Policy Research, Washington DC—a think tank that strives to resolve environmental problems relating to changing climate in Africa based on sound science and policy formulation. His current interest is on climate system/hazard monitoring and diagnostics of climate anomalies aimed at preparing and strengthening institutions for disaster prevention, mitigation, and adaptation.

PUBLIC INTEREST STATEMENT

The desiccation of Lake Chad (LC) in the Sahel region of Africa during the 1960s–1980s has had some devastating socioeconomic impacts on the local economies, and this has been generating increased interest among researchers. In this study, the combined effect of two oceanic indices—the Atlantic Multidecadal Oscillation (AMO) and El Niño Southern Oscillation (ENSO)—on the LC level variability is explored. The late 1960s to the early 1970s marked a turning point in the response of the regional rainfall to climatic drivers, thereby severely affecting the LC level. Thus, the drop in the size of LC level coincided with the combined cold phase of AMO and strong El Niño phase of ENSO. The LC level is responding to recent increases in precipitation by a slow recovery. This slow recovery of the LC level suggests an early warning signal for climate regime change. A proactive intervention in the form of wetland restoration should form an integral part of adaptation and mitigation measures.

understanding of the AMO-ENSO-rainfall-LC level association will help in forecasting the impacts of similar combined episodes in the future. These findings also have implications for long-term water resources management in the LC region.

Subjects: Atmospheric Sciences; Earth Sciences; Earth Systems Science; Environmental Issues; Environmental Studies; Hydrology; Natural Hazards & Risk

Keywords: Lake Chad; Sahel; Africa; AMO; ENSO; drought; lake level; precipitation

1. Introduction

One of the issues of critical importance in ocean-atmosphere interactions is the role of sea surface temperature (SST) anomalies on climate and the implications on regional natural resources. Many studies have shown that precipitation in the Sahel region of Africa is influenced by the impact of oceanic conditions on atmospheric circulations. Such oceanic conditions include the Atlantic Multidecadal Oscillation (AMO) (Zhang & Delworth, 2006), the Mediterranean (e.g. Polo, Rodríguez-Fonseca, & Losada, 2008), the Indian Ocean (e.g. Chung & Ramanathan, 2006), El Niño Southern Oscillation (ENSO), and the Pacific Index (e.g. Joly, Volodio, Douville, Terray, & Royer, 2007). While it has been accepted that SST patterns play a significant role in rainfall variability in the West African Sahel (e.g. Giannini, Biasutti, & Verstraete, 2008), there is still a debate regarding the major oceanic driver (Nicholson, 2013). However, AMO (e.g. Martin, Thorncroft, & Booth, 2014; Polo, Ullmann, Roucou, & Fontaine, 2011) and ENSO (e.g. Caminade & Terray, 2010) have been reported as exerting the most influence on Sahel precipitation (e.g. Martin et al., 2014).

According to Schlesinger and Ramankutty (1994), the North Atlantic SSTs show a 65–75-year variation (0.4°C range). Kerr (2000) referred to this feature as the Atlantic Multidecadal Oscillation. The cold phase of AMO occurred during the 1900s–1920s and 1960s–1980s, while a warm phase occurred during the 1930s–1950s (Schlesinger & Ramankutty, 1994). The warm (cold) phase of AMO is associated with the enhancement (weakening) of Sahel precipitation (e.g. Lu & Delworth, 2005). The increased precipitation during the warm phase has been attributed to increased African easterly wave activity and northward displacement of the Intertropical Convergence Zone (ITCZ) (e.g. Martin & Thorncroft, 2013). Also, the El Niño/La Niña Southern Oscillation (ENSO/LNSO) (Goddard, DeWitt, & Reynolds, 2009; Goddard et al., 2001; Trenberth & Stepaniak, 2001) is one of the greatest sources of global climate variability from seasonal to inter-annual scales. ENSO, a coupled cycle of atmosphere and ocean dynamics (Bjerknes, 1969), has been linked to the devastating droughts of the 1970s and 1980s in the Sahel of Africa (e.g. Giannini, Saravanan, & Chang, 2003). El Niño results in the weakening of the West African Monsoon (WAM) flow, creating a dry condition across the Sahel region (e.g. Janicot, Mounier, Gervois, Sultan, & Kiladis, 2010). La Niña on the other hand creates a wet condition through the enhancement of the Walker circulation (Nicholson, Some, & Kone, 2000).

At the intersection of the dry Sahel and wet tropical region of West Africa is the “shrinking” Lake Chad (LC) affected by climate variability (e.g. Polo et al., 2011) and subjected to immense anthropogenic pressure (e.g. Okonkwo & Demoz, 2014). The desiccation of LC has had some devastating socioeconomic impacts due to the dependence of local economies on agriculture, fishing, and livestock production (Okonkwo, 2013). This has been generating some interest among researchers and policy-makers (e.g. Lemoalle, Bader, Leblanc, & Sedick, 2012). The decrease in the size of LC has been attributed primarily to increased human water use and declining precipitation (IPCC, 2007). The declining precipitation is linked to the complexity in atmospheric dynamics and remote influence of climatic indices like AMO and ENSO. While the effect of AMO and ENSO on precipitation in the Sahel region has been studied repeatedly, though separately, the combined effect of these two oceanic conditions (with both negative and positive phases) has received little attention. Also, most of the studies on LC level variability had focused on the 1970s and 1980s drought years, during which LC split into the northern and southern pools (e.g. Lemoalle et al., 2012). These periods coincide with the cold phase of AMO and increased frequency of El Niño episodes. However, the response of LC

level to precipitation during combined phases of AMO and ENSO dating back to the early 1900s has not been documented.

The objective of this study is to investigate the combined impact of past and current warm and cold phases of AMO and ENSO on the LC level and size. We will do this by documenting the timing of LC level increase and decrease in response to AMO-ENSO forcing on Sahel precipitation. Understanding the remote effect of SST forcing on the decrease in LC level can help in the hydrological modeling of LC level variability, rainfall forecast, and long-term planning in water resources management. Also, knowledge of the AMO-ENSO-rainfall relationship in the Sahel region can help the region to adequately prepare to mitigate the effect of drought on its agricultural sector in the future. The possible physical processes and implications of the findings of this paper in relation to the management of LC are also discussed. The paper is organized as follows. Section 2 describes the methodology, while Section 3 presents and discusses the results. Summary and conclusions are given in Section 4.

2. Data-set and methodology

2.1. Study area

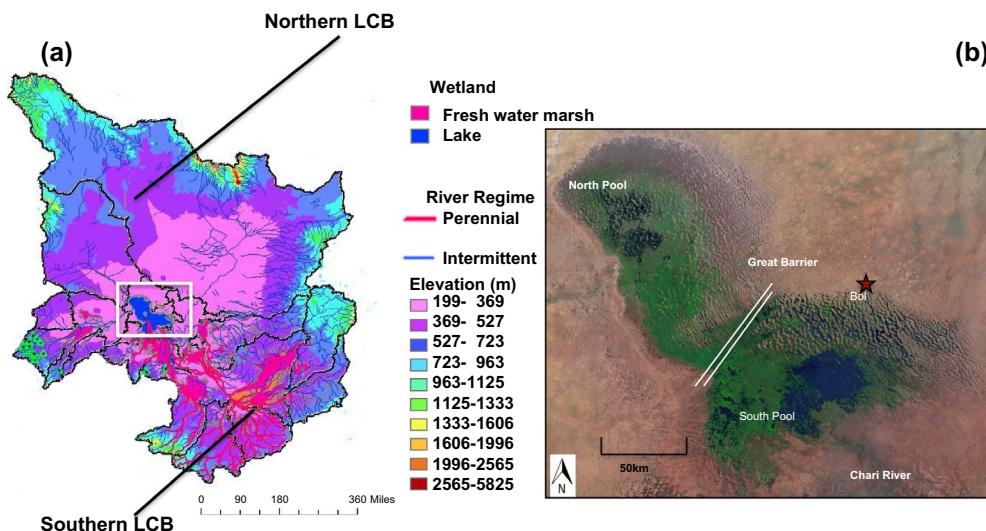
The Lake Chad Basin (LCB) (latitude 6°N–24°N; longitude 7°E–24°E) is part of the African Sahel, a semi-arid region that is prone to drought (e.g. Tarhule & Lamb, 2003). LC, a closed lake at the center of the LCB (Figure 1(a)), is highly sensitive to hydro-climatic events (Leblanc, Lemoalle, Bader, Tweed, & Mofor, 2011). The spatial distribution of the elevation, the perennial and intermittent surface waters, and the freshwater marsh in the LCB are also shown in Figure 1(a). The LCB hydro-climatic system is influenced by many factors including precipitation, river discharge, anthropogenic activities, and some large-scale climate indices and teleconnections (e.g. Lauwaet, van Lipzig, Van Weverberg, De Ridder, & Goyens, 2011). Studies of LC have documented its reduction in size from 22,000 km² in the 1960s to 300 km² by the 1980s (Singh, Diop, & M'mayi, 2006); the presence of the ridge referred to as the “great barrier” (Figure 1(b)) by Olivry, Chouret, Vuillaume, Lemoalle, and Bricquet (1996) that runs between the southern and northern parts of the lake; the splitting of the lake into two smaller lakes when the inflow is below the barrier (Gao, Bohn, Podest, McDonald, & Lettenmaier, 2010); and the inflow from the southern parts of the basin through the Chari River (Figure 1(b)) that significantly determines the lake's behavior (e.g. Gao et al., 2010).

2.2. Data sources

The LC water levels used in this study were based on 1900–2010 measurements at Bol monitoring station (Figure 1(b)) and satellite measurement from Topex/Poseidon satellite, a joint NASA/CNES (National Aeronautic and Space Administration/Centre National d'Etudes Spatiales, France) altimetry mission launched in 1982. Reconstructed levels from hydrological modeling study by Lemoalle et al. (2012) were used to complement for years with no data from the Topex/Poseidon satellite. The figures from Lemoalle et al. (2012) and LC level measurements at Bol, as reported in “Africa's lakes: atlas of our changing environment” by United Nations Environment Programme, were scanned as raster images that were then digitized using geographic information system tools. The digitization was carried out using point mode operation at maximum annual lake levels. Post-processing was applied to the digitized map by checking lake levels against the source figure for accuracy. The primary precipitation data-set used in this study is the gridded station monthly rainfall anomalies (cm) taken from the National Oceanic and Atmospheric Administration (NOAA) Global Historical Climatology Network. The gridded data points were produced at 5° by 5° resolution, and, to address inhomogeneity, averaging for the Sahel region was performed based on a rotated principal component analysis of African precipitation by Janowiak (1988). Also, stations with at least 20 years of data within the 1961–1990 period were used. Further details on the data are described at <http://www.ncdc.noaa.gov/temp-and-precip/ghcn-gridded-products.php>. The precipitation data-set used in this study is for July–September (JAS) coinciding with the WAM season, which is the peak period of the short rainy season in the Sahel (Thorncroft, Nguyen, Zhang, & Peyrillé, 2011).

Figure 1. (a) Geographical, topological, and some hydrological references of the LCB showing the freshwater marsh, elevation (m), and the perennial rivers (Okonkwo & Demoz, 2014). (b) True-color composite image acquired in May 2003 from Landsat 7, showing the north and south pools of LC.

Note: three monitoring stations; Bol, Kalom, and Kindjeria are also shown (red star).



ENSO is represented in this study by SST data from the Niño 3.4 region (5°S – 5°N and 120° – 170°W). The Niño 3.4 SST period used in this study is based on a new strategy of updating by the Climate Prediction Center (CPC) of the US NOAA. In this new approach, multiple 30-year base periods are used to calculate anomalies for successive 5-year periods in the historical record (CPC, 2013). The warm (cold) phase of ENSO is a positive (negative) departure from normal greater (less) than or equal to $+0.5^{\circ}\text{C}$ (-0.5°C). However, it should be noted here that before 1950, there were less quality controlled and fewer SST measurements to reconstruct the Niño 3.4 index (Trenberth & Hoar, 1997). To mitigate this concern, we used a new Niño 3.4 index generated by Bunge and Clarke (2009) with high correlation between physically related and independent time series of $\delta^{18}\text{O}$ from a coral at Palmyra, an atoll inside the region Niño 3.4. For detailed information on these indices (see Bunge & Clarke, 2009).

The AMO part of a coherent temperature variation across much of the Northern Hemisphere (Knight, Allan, Folland, Vellinga, & Mann, 2005) is the average of North Atlantic Ocean SST anomalies and was obtained from the NOAA Earth System Research Laboratory website (<http://www.esrl.noaa.gov/psd/data/correlation/amon.sm.long.data>). Kaplan SST data-set [5×5] is used in calculating the AMO index by area-weighted averaging followed by de-trending. When filtered with a 10-year running mean, the AMO exhibits a positive (negative) phase during the 1930–1959 (1965–1994) period (Enfield, Mestas-Nuñez, & Trimble, 2001). The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis 2 data-set (Kalnay et al., 1996) has been used for the large-scale circulation and physical mechanism discussion.

2.3. Data processing

Attention is focused on the multi-decadal character of the data-sets by applying a 10-year running mean to linearly de-trended time series of the AMO index, LC level, and Sahel precipitation. We applied an eight-year running mean to Niño 3.4 anomalies to obtain decadal variability following Yeh and Kirtman (2005). According to Rose and Darrell (2009), the running means help in minimizing the “noise” from random variability in the observed data and enabled us to focus our analysis on the relationship between ENSO and rainfall variability in the region. Conventional linear correlation was used for quantitative association between the LC level, precipitation, and the atmospheric indices (AMO and ENSO).

3. Results

3.1. Large-scale circulation associated with AMO and ENSO SST anomalies

Precipitation in the West African region is influenced by large-scale atmospheric circulation and climate variability. These large-scale circulation transport moisture and heat inland from the ocean at rates that vary at decadal and inter-annual timescales. In this section, the relations of AMO-ENSO phase and their influence on the West African rainfall by way of affecting the characteristics of atmospheric transport of moisture and heat into the Sahel region are analyzed.

Figures 2–4 show the composite maps of combined AMO and ENSO conditions of precipitation, outgoing longwave radiation (OLR), and surface relative humidity, respectively. The composite maps are for all combinations of AMO and ENSO (i.e. +AMO and +ENSO, +AMO and -ENSO, -AMO and +ENSO, -AMO and -ENSO). All anomalies are with respect to the 1981–2010 mean.

According to Vizy and Cook (2001) and Paeth and Stuck (2004) a warmer than normal Atlantic SST during the warm phase of AMO creates a dipole response leading to wet conditions across West Africa including the Sahel (Nicholson, 2008) as seen in Figure 2(a) and (c). This positive rainfall anomaly is as a result of initiation and increased intensity of meridional propagation of the ITCZ over West Africa (Ward, 1998). The composites (Figures 2 and 3) show relatively more precipitation for +AMO with +ENSO than +AMO with -ENSO. This apparent contradiction of the expected increased precipitation during La Niña suggests that the AMO may have the strongest and more consistent association with precipitation. This will be explored further with correlation analysis in Section 3.2.

In general, precipitation over Western Africa decreases during the positive phase of ENSO. According to Chiang and Sobel (2002) and Lu (2009), warming of the Pacific basin triggers local convection and propagation of Rossby and Kelvin waves and subsequently tropospheric warming of West Africa. The SST anomalies move westward over the equatorial Pacific during the WAM (JJA).

Figure 2. Composite maps of combined AMO and ENSO conditions showing precipitation anomalies for (a) +AMO and +ENSO [10], (b) -AMO and +ENSO [8], (c) +AMO and -ENSO [8], and (d) -AMO and -ENSO [10].

Notes: The period for the anomalies is 1948–2010, while the number of events in each composite is given in square brackets. The anomalies are with respect to the 1981–2010. Source: Data from NCEP-NCAR reanalysis 2.

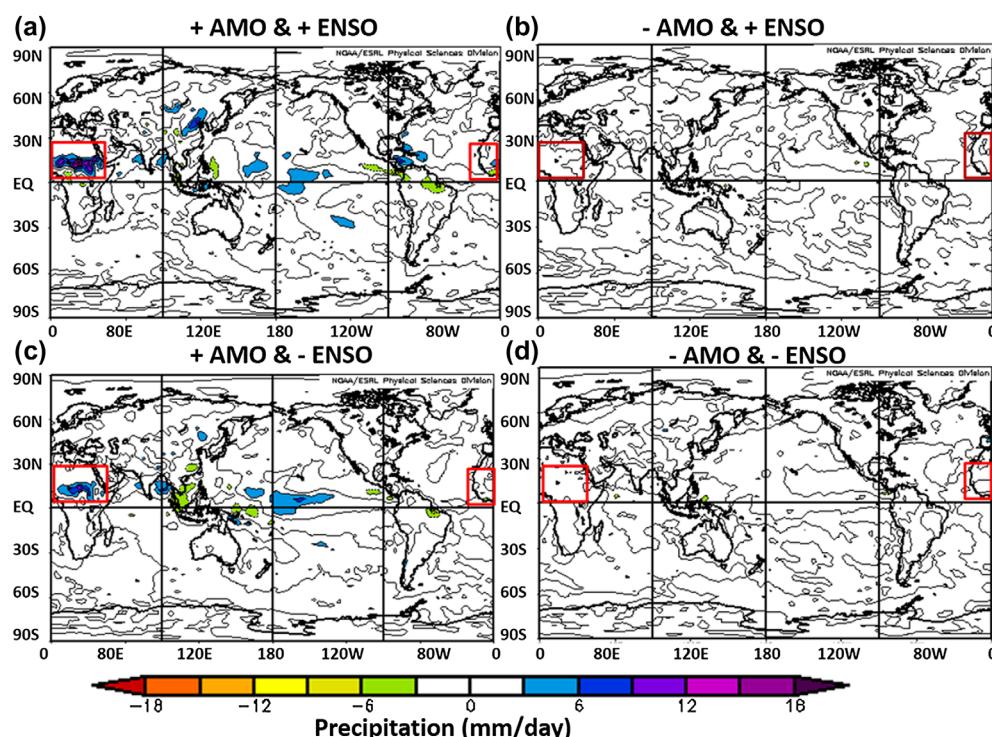


Figure 3. Composite maps of combined AMO and ENSO conditions showing OLR anomalies for (a) +AMO and +ENSO [10], (b) -AMO and +ENSO [8], (c) +AMO and -ENSO [8], and (d) -AMO and -ENSO [10].

Notes: The period for the anomalies is 1948–2010, while the number of events in each composite is given in square brackets. The anomalies are with respect to the 1981–2010. Source: Data from NCEP–NCAR reanalysis 2.

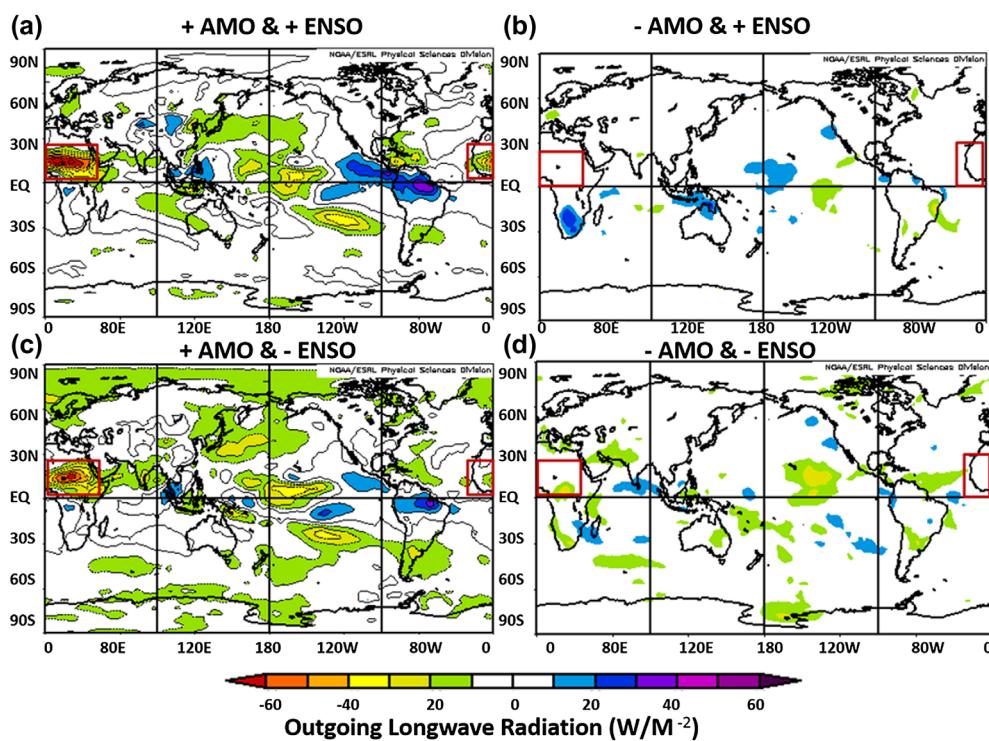
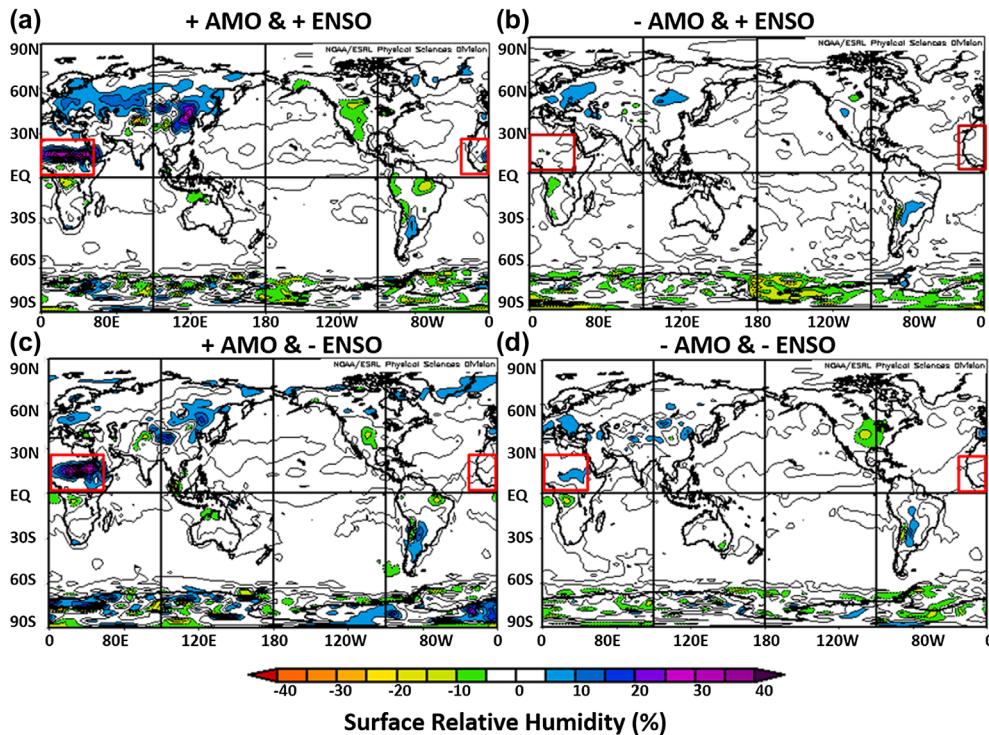


Figure 4. Composite maps of combined AMO and ENSO conditions showing surface relative humidity (RH) anomalies for (a) +AMO and +ENSO [10], (b) -AMO and +ENSO [8], (c) +AMO and -ENSO [8], and (d) -AMO and -ENSO [10].

Notes: The period for the anomalies is 1948–2010, while the number of events in each composite is given in square brackets. The anomalies are with respect to the 1981–2010. Source: Data from NCEP–NCAR reanalysis 2.



This weakens the monsoon and subsequently Sahel precipitation due to anomalous subsidence. This anomalous subsidence is captured by the OLR shown in Figure 3.

A closer look at Figure 3(a) and (c) further reveals that the negative anomaly in OLR over the West African region prevails for the +AMO and +ENSO and +AMO and -ENSO composites. The increased ascent in these regions is associated with enhanced moisture convergence and positive precipitation anomalies as shown in Figure 2(a) and (c). A comparison between the OLR and surface relative humidity (Figure 4) depicts a mirror image of the magnitude of the positive anomaly.

On the other hand, Figure 3(b) and (d) shows the near neutral conditions for -AMO and +ENSO and -AMO and -ENSO composites, similar to the rainfall anomalies. The differential weakening of Walker circulation and Tropical Easterly Jet (TEJ) via the east-west Walker circulation during the warm phase of ENSO was further amplified by the cooler Atlantic.

Zhang and Delworth (2006) and Hwang, Frierson, and Kang (2013) described the meridional shift in the ITCZ and increased rainfall in the Sahel region associated with the differential heating of the Northern and Southern Hemispheres. While much of Southern Africa remained dry, there is a swath of above normal moisture indicated by surface relative humidity in Figure 4(a) and (c) across the stretch of West Africa and the Sahel region during the +AMO and +ENSO and +AMO and -ENSO. This increased the moisture content of the monsoon flow and subsequently precipitation.

3.2. Impact of combined effect of AMO and ENSO on LC level variability

Understanding the past and present multi-decadal variability in the LC level is clearer when the anomalies in AMO, precipitation, and ENSO are considered jointly. Figure 5 shows the comparison of co-variability of the warm (El Niño) and cold (La Niña) phases of ENSO, western Sahel precipitation, and the AMO index on the LC level. The mean LC levels for the periods 1900–2009, 1930–1959, 1960–1989, and 1990–2010 are 281.36, 281.24, 280.23, and 279.36 m, respectively.

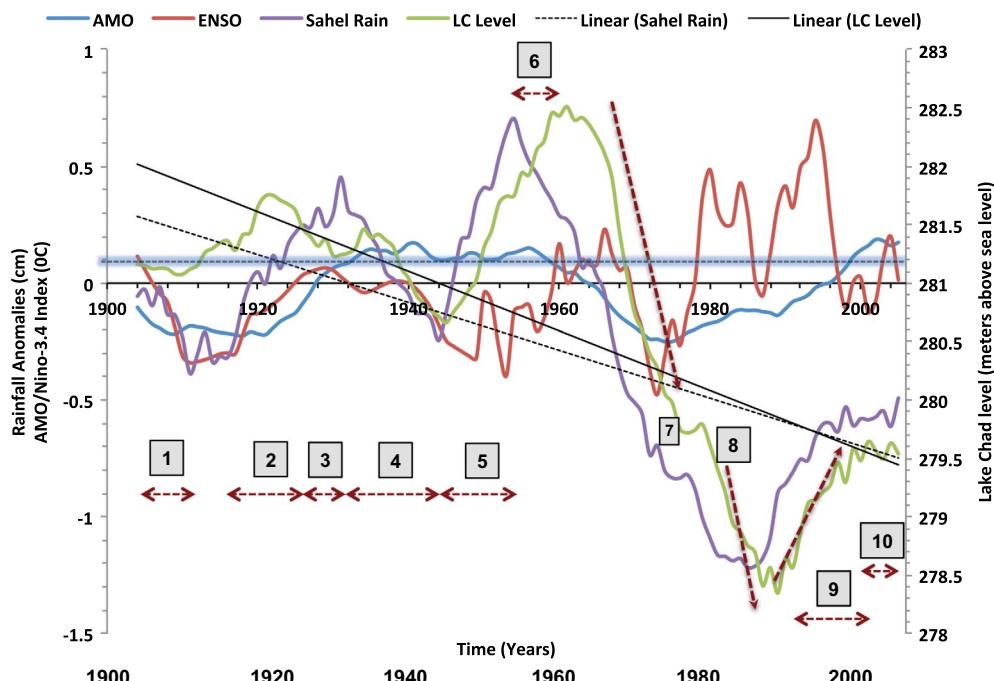
The significant periods in the response of the LC level variability to precipitation and the combined phase of AMO and ENSO are indicated by the grey numbered boxes in Figure 5. The selected periods represent one of the following conditions: (1) years of simultaneous increase or decrease in LC level and precipitation, (2) years of increasing (decreasing) LC level and decreasing (increasing) precipitation, and (3) sharp decrease in LC level. Table 1 shows a summary of the periods, LC level, and signs of AMO, ENSO, and precipitation. The period 1904–1910 (grey box 1 Figure 5) shows decreasing precipitation with a slight decrease in LC level around 1909. The continued decrease in precipitation did not however reflect on the LC level, which initially showed steady-state conditions in the early 1900s and then increasing afterwards. Groundwater seepage during these years of relatively decreasing precipitation could have been recharging LC, thereby explaining the physical disconnect. This period is in the combined cold phase of AMO and El Niño, both of which are known to have a negative impact on precipitation in the region. The period 1916–1924 shows increasing LC level in response to increasing precipitation and transitioning of AMO from cold to warm phase (grey box 2 Figure 5).

However, the 1926–1932 period (grey box 3) shows decreasing lake level with increasing precipitation. This is counterintuitive, as we expect the lake level to increase with increasing precipitation for a closed lake system. A possible explanation of this physical disconnect can be attributed to excessive evaporation from the LC surface. Evaporation is an important component of the hydrologic cycle in a lake system (e.g. IPCC, 2007; Vallet-Coulomb, Legesse, Gasse, Travi, & Chernet, 2001). The interaction between evaporation, rainfall, inflow, and run-off is reflected in the water balance in the LC hydrology (Birkett, 2000). The clearest evidence in support of our evaporation argument could be seen in a North Atlantic SST as a proxy for tropics-wide warming (Giannini et al., 2003).

We believe that the increased water temperatures of LC relative to overlying air resulted in increased evaporation and thus decrease in the LC level. This is consistent with Contoux et al. (2013) that Megalake Chad was an important source for evaporation. The increased precipitation during this period is also supported by an experiment using a mesoscale regional atmospheric model coupled to a soil–vegetation–atmosphere model, that found precipitation increase for filled pre-industrial LC scenario (Lauwaet et al., 2011). One other possible explanation of this physical disconnect is

Figure 5. Evolution of the Atlantic Multi-decadal Oscillation (AMO) index (blue) shown together with ENSO (red), LC level variability (green) measured at Bol, and Sahel rainfall (purple). Eight- and 10-year running means were computed for ENSO and LC level, respectively.

Notes: The shaded blue dotted line represents the 1900–2010 mean lake level. The grey numbered boxes represent some significant periods in the response of the LC level variability to precipitation and the combined phase of AMO and ENSO.



decreased inflow to the lake due to agricultural and irrigational activities within the LCB upstream of LC.

As expected, the 1932–1944 period (grey box 4) saw a decrease in the lake level with decreasing precipitation, while the 1944–1954 period (grey box 5) shows increasing lake level with increasing precipitation. The increase in precipitation can be attributed to the moderate La Niña episodes.

During the mid-1950s–1960s, the LC level–rainfall relationship was in transition with a time lag in preparation for a drastic decrease following precipitation. During the lag time, there was a decrease in precipitation in the region, but a marked increase in LC level (grey box 6; Figure 5). The question is where does the inflow to the lake come from when precipitation is decreasing? This apparent physical disconnect in the LC level response to decreasing precipitation can be attributed to groundwater recharge, as there is no other plausible physical mechanism that can explain it. Our result thus suggests that LC must have been replenished through groundwater seepage during these years of relatively decreasing precipitation. According to Geerken, Vassolo, and Bila (2010), LC is linked to groundwater even as a closed basin. The physical terrain of the LCB with mountainous ranges at the borders (Figure 1(a)) means that precipitation ultimately reaches the lake either through surface flows by way of the tributaries or through groundwater discharge (Okonkwo & Demoz, 2014).

The period between the late 1960s and early 1980s (grey box 6 to grey box 8, Figure 5) marked a turning point in the response of the LC level to climatic (and possibly anthropogenic) drivers in the region. This is the most striking feature in the historical changes in the LC size. It also shows a marked departure in both rainfall and LC level variability compared to the period of 1900s–early 1950s. Trenberth et al. (2007) qualified this outstanding decrease in precipitation as the largest climate change event observed thus far. This period also corresponds to the beginning of the last cold phase of AMO known to have caused a reduction in the Sahel precipitation due to a southward displacement of the Inter-tropical Convergence Zone (e.g. Zhang & Delworth, 2006). The negative trend in precipitation in the 1970s also coincided with La Niña episodes that probably resulted in increased precipitation across the Sahel region. Indeed, this La Niña effect is represented by period 7, which shows a mild recovery in the LC level due to increased precipitation in the region in the early 1970s.

However, that mild recovery did not last long, as the region witnessed a prolonged drought period from the late 1970s to the early 1980s. The period from 1978 to the mid-1990s (grey box 8 Figure 5) coincides with the combined cold phase of AMO and strong El Niño phase of ENSO. The decreased precipitation during this cold phase of AMO has been attributed to decreased African easterly wave activity and southward displacement of the ITCZ (e.g. Martin & Thorncroft, 2013). El Niño on the other hand weakens the WAM flow, creating a dry condition across the Sahel region (e.g. Janicot et al., 2010). The drop in the size of the LC level from 282.5 m in the early 1960s to about 278.1 m in 1983/1984 was the largest to occur during the period of study and accompanied one of the strongest El Niño events of 1982/1983. Since the AMO had started warming up from the early 1990s, the LC level appears to be on the path to recovery (grey box 9 Figure 5). However, since the early 2000s (grey box 10 Figure 5), despite the prevailing warm phase of AMO, precipitation appears to be in a steady state. As a result, the LC level is still below the 1900–2010 mean of 281.1 m above sea level (shaded blue dotted line; Figure 5).

We also used quantitative analytical techniques based on correlation to assess the relationships between the LC level variability and rainfall. From our results, the lake level at Bol has a statistically significant correlation with precipitation ($R^2 = 0.6$, at the 99.5% confidence level).

A number of additional correlation analyses were performed to identify the relationships between precipitation, AMO, LC level, and ENSO. The result is summarized in Table 2. The correlation between LC level and the precipitation is 0.52, 0.74, 0.96, and 0.86 for 1900–1929 (−AMO), 1930–1959 (+AMO), 1960–1989 (−AMO), and 1990–2010 (+AMO), respectively. These correlations are high and statistically significant at a 99% confidence level. On the other hand, the correlation between LC level and the ENSO is only 0.21, 0.08, −0.14, and −0.65 for 1900–1929 (−AMO), 1930–1959 (+AMO), 1960–1989

Table 1. Summary of the periods, LC level, and signs of AMO, ENSO, and precipitation

Period	Years	AMO	ENSO	Precipitation	LC level*
1	1904–1910	–	–	–	Mean
2	1920–1926	–	–	+	Above
3	1928–1932	+	–	+	Above
4	1932–1944	+	+	–	Below
5	1944–1954	+	–	+	Above
6	1956–1968	+	+	+	Above
7	1972–1975	–	–	–	Below
8	1978–1988	–	–	–	Below
9	1988–2000	–	+	–	Below
10	2000–2010	+	Neutral	–	Below

Notes: Negative are for cold phase of AMO and La Niña phase of ENSO, while positive signs are for warm phase of AMO and El Niño phase of ENSO. The precipitation anomalies are with respect to 1950–1979 mean.

*The LC level (classified as above or below mean) is with reference to the 1900–2010 mean.

Table 2. Correlations between LC level, precipitation, AMO, and ENSO

Variables	Period (AMO phase)			
	1900–1929 (−)	1930–1959 (+)	1960–1989 (−)	1990–2010 (+)
Lake level–precipitation	0.52*	0.74*	0.96*	0.86*
Lake level–ENSO	0.21	0.08	−0.14	−0.65*
ENSO–precipitation	0.84*	−0.18	−0.15	−0.62*
Lake level–AMO	−0.01	−0.17	0.53*	0.88*
AMO–precipitation	0.73*	−0.18	0.63*	0.80*

*Significant level at 99%.

(-AMO), and 1990–2010 (+AMO), respectively. These correlations are low and statistically non-significant except for the 1990–2010 positive phase of AMO.

Similarly, the correlation between AMO and the precipitation is 0.73, -0.18, 0.63, and 0.80 for 1900–1929 (-AMO), 1930–1959 (+AMO), 1960–1989 (-AMO), and 1990–2010 (+AMO), respectively. These are all statistically significant at a 99% confidence level except for 1930–1959 positive phase of AMO. On the other hand, the correlation between LC level and the AMO is -0.01, -0.17, 0.53, and 0.88 for 1900–1929 (-AMO), 1930–1959 (+AMO), 1960–1989 (-AMO), and 1990–2010 (+AMO), respectively.

There are several possible explanations for these different quantitative results. First, the LC level and size are derived predominantly from precipitation during the short period of Boreal Summer (July, August, and September). As such, the correlation between LC level and precipitation is consistently much stronger when compared to LC-AMO or LC-ENSO (Table 2). The relatively high correlation of the AMO index with Sahel precipitation (except for the 1930–1959 phase, Table 2) is a confirmation of its effectiveness. In particular, the high positive correlation between LC level and precipitation (0.63) and LC level and AMO (0.53) suggests that the expected precipitation outcome is not so much dependent on Niño 3.4 sign, but on the AMO phase as indicated by the composites (Figures 2 and 3). Interestingly also, the small negative correlation between LC level and precipitation (LC level and ENSO) during the 1960–1989 negative phase of AMO is -0.15 (-0.14) and not statistically significant.

Figure 6 shows the scatter plot between the AMO index and precipitation anomalies. The fit between dry conditions and -AMO (1960–1989 and 1900–1920 phases) is around 0.5 which is weaker than that between wet conditions and +AMO phase of 1990–2010 which is 0.7. On the other hand, however, the fit between dry conditions and -AMO (1960–1989 and 1900–1920 phases) is stronger than that between wet conditions and +AMO phase of 1930–1959. A possible explanation of this conflicting signal could be linked to the combined AMO-ENSO and other oceanic indices' (Mediterranean and Indian) modulation of upper air circulation features, especially the TEJ and African easterly jet (AEJ) over the Sahel region during the WAM season.

Figure 6. Scatter plot between the AMO Index and precipitation anomalies for the cold and warm phases of AMO.

Note: R^2 denotes the goodness of fit.

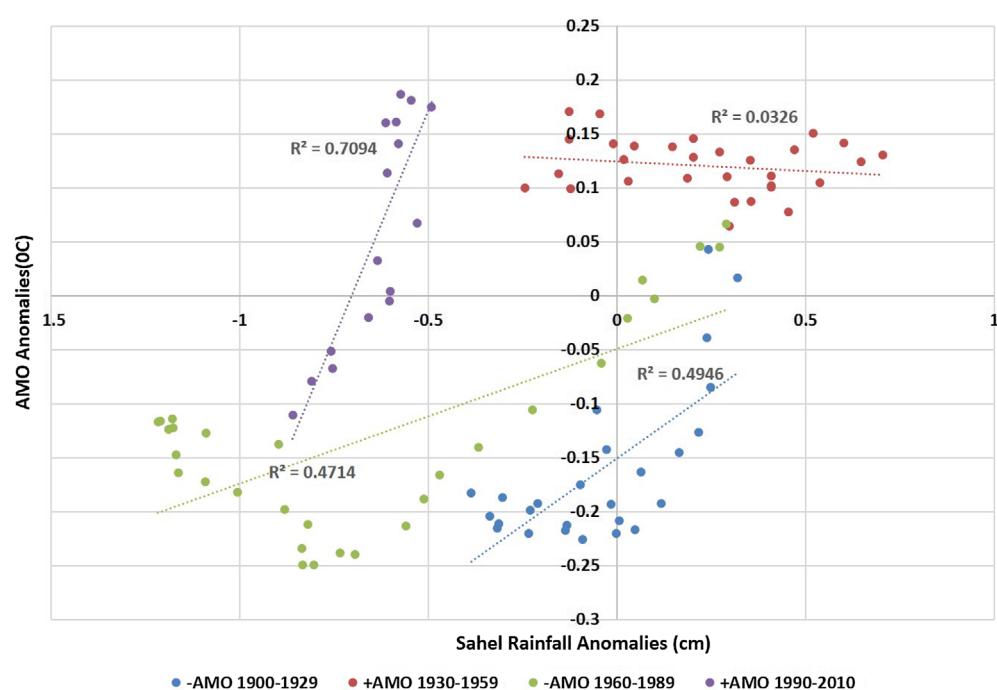
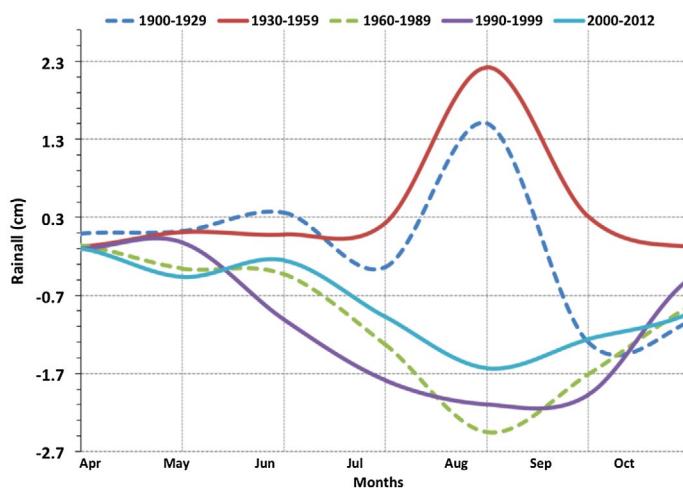


Figure 7. Plot of April–October western Sahel rainfall (cm) anomalies for different cold and warm periods of AMO.

Notes: The dotted lines are for cold phase of AMO while the solid lines are for warm phase of AMO.



According to Druyan (2011), one of the conclusions that came out of multiple studies on the inter-annual to decadal variability in precipitation over West Africa is that it is controlled by competing physical mechanisms. We thus note here that an analysis of the physical mechanisms and interaction of how AMO modulates ENSO rainfall in the Sahel is complicated and beyond the scope of this paper. However, we will address the possible physical mechanism during the combined phase of AMO and ENSO in Section 3.3.

Overall, the response of the LC level mirrors precipitation from the 1970s to the late 2000. It is therefore important to take a closer look at the temporal variability of rainfall averaged over different warm and cold phases of AMO for the period under study. The choice of AMO is predicated on the multi-decadal variability of Sahel rainfall suggested by our result similar to some recent studies (e.g. Martin & Thorncroft, 2013; Ting, Kushnir, Seager, & Li, 2009).

Figure 7 shows April–October western Sahel rainfall anomalies (cm) with respect to the 1950–1979 climatology, for different cold and warm periods of AMO. For the period of this study, the wettest (driest) period is 1958–1974. We thus used the 1950–1979 climatological mean to capture these two extremes of the wet and dry periods in this region.

The Sahel region experienced a much wetter condition during the cold phase of AMO in the 1900s–1920s compared to the 1960s–1980s (Figure 7). This can be attributed to the moderate La Niña events in the 1900s–1920s (Figure 5), known to favor precipitation in the region through the enhancement of the Walker circulation (Nicholson, 2001). On the other hand, the high intensity of El Niño events together with the cold phase of AMO during the 1970s and 1980s (Figure 5) appears to have a combined effect of reducing precipitation to produce the most significant drought thus far seen in the region. This is reflected in the negative rainfall anomalies for that period in contrast to positive rainfall anomalies for the 1900s–1920s (Figure 7).

In the 2013 State of the Climate report (Sima, Kamga, Raiva, Dekaa, & James, 2013), the 2012 extensive flooding in the Sahel was used as a pointer to a full return to “wet” periods across the region. Our analysis however shows negative rainfall anomalies for the 1990–1999 and 2000–2012 periods compared to 1900–1950 (Figure 7). This difference could be explained by the use of the 1981–2010 climatology in the State of the Climate report—a period characterized by severe drought in the Sahel region. Our results thus suggest that while the Sahel regional rainfall appears to be on a path to recovery, it is still short of the “wet” period experienced in the 1930–1950s era. This shortfall in precipitation in the Sahel region could also explain the slow recovery of the LC level as shown in Figure 5.

3.3. Physical mechanisms and impacted hydrological parameters

To discuss the physical mechanism of the combined AMO-ENSO phase on the atmospheric conditions that control precipitation in the region, we first identify two distinct cases: (Case 1), a warm phase of AMO and moderate El Niño phase represented by 1959–1962 and (Case 2), a cold phase of AMO and strong El Niño period represented by 1982–1985 (see Figure 2). The criteria for selecting these periods are the availability of relevant atmospheric fields from NCEP–NCAR reanalysis data from 1948 onwards and the representative composites for each case from the smoothed time series (Figure 5). This discussion will focus on the key physical mechanisms at play during WAM systems, namely the AEJ, TEJ, and “rainbelt” (Nicholson, 2009).

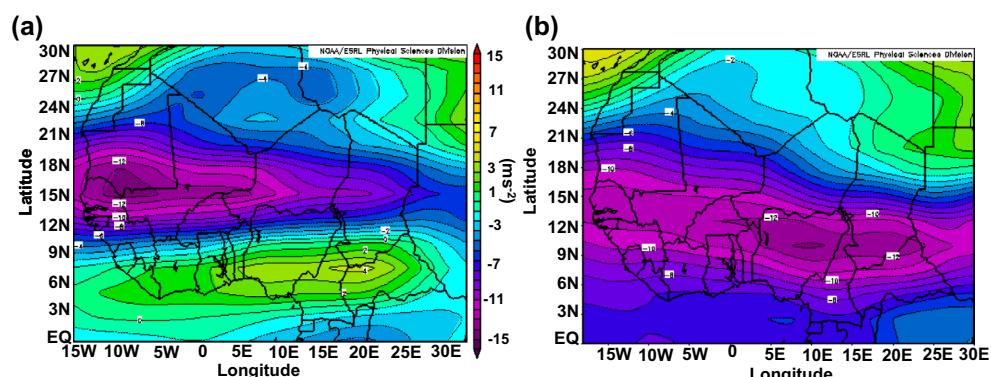
Figure 8 shows the location of the AEJ for the combined warm and cold AMO and ENSO phases. According to Nicholson (2013), AEJ plays a role in generating African Easterly Waves which contribute to the WAM precipitation. The development of the AEJ is due to the latitudinal temperature gradient between the Atlantic Ocean and the Sahara (Nicholson & Grist, 2003). The combined AMO warm phase and El Niño shows a clear northward location of the core of the AEJ (Figure 8(a)) in contrast to the southward location during the cold AMO phase and El Niño (Figure 8(b)). The AEJ has a core over the West African Sahel centered over 15°N–18°N for case (1) and around 9°N–12°N for case (2). This northward displacement in case (1) boosts the vertical and horizontal wind shear (Grist & Nicholson, 2001), which favors precipitation formation.

There is also latitudinal variability in the structure of the AEJ in agreement with Hall, Kiladis, and Thorncroft (2006). Okonkwo, Demoz, and Tesfai (2014) have also documented that AEJ correlates better with rainfall north of the equator in the summer months. The high intensity of El Niño events together with the cold phase of AMO in the 1980s (Figure 5) appears to have a combined effect on suppressing precipitation to produce the most significant drought in the region.

The TEJ (not shown) develops from the north-south temperature gradient between the Indian Ocean and Himalayan plateau (Nicholson, 2013). The TEJ has been reported to be weaker in the El Niño years (Chen & van Loon, 1987), which translates to more tropospheric convergence and less tropospheric divergence (e.g. Janicot, 1997). During an El Niño event, there are anomalous warm SSTs in the eastern tropical Pacific and anomalous cold SSTs in the western tropical Pacific due to equatorial wave forcing (Laing & Evans, 2011). The heating (atmospheric and oceanic in nature) in the western flanks of the equatorial circulations drives the Walker circulation (Laing & Evans, 2011). The weakening (or breakdown) of the Walker circulation during El Niño is propagated eastward to the Atlantic and West African region. As a result, the TEJ that is energetically maintained by the tropical divergent circulations associated with the east-west Walker circulation (Nicholson, 2013) is diminished.

The meridional component of the TEJ may be playing an important role in rainbelt development over this region during the monsoon season by creating a convergence zone around the 200-mb pressure level (Martin & Thorncroft, 2013; Nicholson & Grist, 2003). These conditions are unfavorable

Figure 8. The AEJ (ms^{-1}) at 600 mb for (a) 1959–1962 and (b) 1982–1985 zonally averaged between longitudes 18°W and 10°E.



for convective precipitation systems in the region (Jenkins et al., 2005) and could further explain the reduced precipitation during the combined warm phase of AMO and strong El Niño. This physical mechanism is similar to Slingo and Annamalai (2000) that linked the unexpected impact of the 1997/1998 El Niño on Indian monsoons to teleconnections to the Walker and Hadley (lateral monsoon) circulations. Also, Okonkwo et al. (2014) reported a statistically significant association between the TEJ and the El Niño events of the 1980s that led to intense drought in western Sahel.

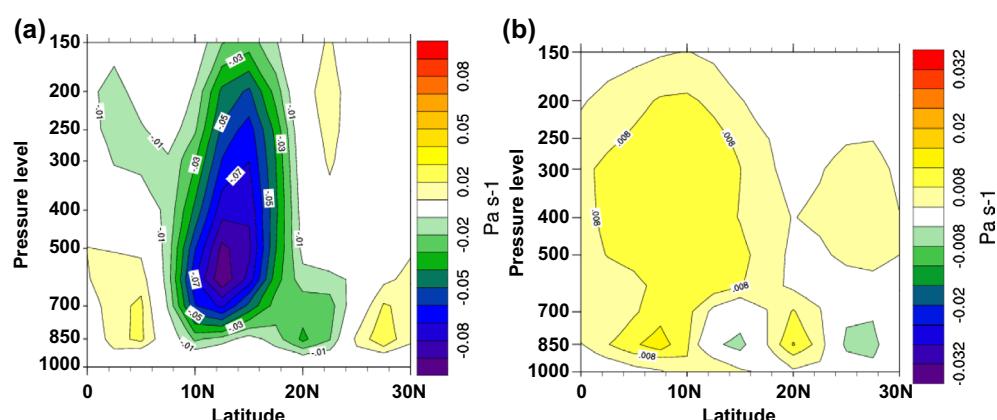
Figure 9(a) and (b) shows rising air in terms of vertical velocity in Pa s^{-1} for July, August, and September (JAS) for cases (1) and (2), respectively. Negative signs (shades of blue to purple colors) represent upward vertical velocities while positive signs (green to red) denote downward velocities. There is upward vertical velocity for case (1) during the summer months characterized by a very strong core (Figure 9(a)) between the TEJ and AEJ in the region classified as tropical rainbelt (Nicholson, 2009) or “rain band” (Zhang & Delworth, 2006). This region is a key feature in the revised picture of WAM and associated rainfall in relation to the column of rising air bounded on the north by the AEJ axis and on the south by the TEJ axis (Nicholson, 2009). The strong region of rising air stretches from the lower troposphere to the upper troposphere.

According to Nicholson (2009), not only is humid air limited to this region of ascent, but also the juxtaposition of TEJ and AEJ suggests that this vertical motion within the rainbelt acts as a vehicle for moisture transport from the surface to the troposphere. There is a strong coupling between the moisture available for convection and its uplift strength controlled by the characteristics of AEJ and TEJ (Nicholson, 2009). It is also possible that the convective processes may have contributed to the enhanced vertical motion. This possibility according to Nicholson (2009) was a big factor in the overall wet 1950 West African rainy season. The impact of this rising air is the increased precipitation in the region in the 1950s–early 1960s. In contrast to case (1), case (2) shows downward velocities in the summer months between the Equator and 20°N (Figure 9(b)). This region of downward vertical velocity is in the rainbelt and could explain the difference in anomalous precipitation in the region between the cases.

The link between the LCB hydro-climatic system and the physical mechanisms from the results above will be discussed next. Figure 10 shows the regional precipitation and run-off anomalies (based on the 1982–2010 climatology) for the two cases. The LCB is indicated by the red box. There is clearly an increased precipitation (positive anomaly) of 20–80 mm in the LC region in case 1 (Figure 10(a)) compared to decreased precipitation (negative anomalies) of –20–80 mm for case 2 (Figure 10(b)). The positive precipitation anomaly is most pronounced across the Sahel region, particularly the LC region (red box, Figure 10(a)). Rainfall for the entire West African region is also well above normal for case 1. Similarly, the Sahel region was most impacted by rainfall anomalies in case 2 (Figure 10(b)). The anomalies are also apparently more severe around the LCB (Figure 10(b)), particularly within lat. 10°N–13°N in southern LCB. Rainfall anomaly was in the range of 40–60 mm in southern LCB in case 2.

Figure 9. Anomalies of vertical velocity (Pa s^{-1}) (a) 1959–1962 and (b) 1982–1985, zonally averaged between longitudes 18°W and 10°E.

Notes: Negative signs (shades of blue to purple colors) represent upward vertical velocities, while positive signs (green to red) denote downward velocities.



This analysis of precipitation and run-off incorporates an important parameter—river discharge—in a hydrological closed basin like LC. As expected, locations of increased precipitation result in increased run-off (Figure 10(a) and (c)). On the other hand, lower precipitation across the Sahel region led to a run-off that is below climatological mean based on the 1981–2010 mean (Figure 10(b) and (d)). The positive anomaly for case 1 from reanalysis data shows that the LC region is indeed significantly affected. As such, the relationship between these anomalies in precipitation and run-off is reflected in the LC level variability shown in Figure 5.

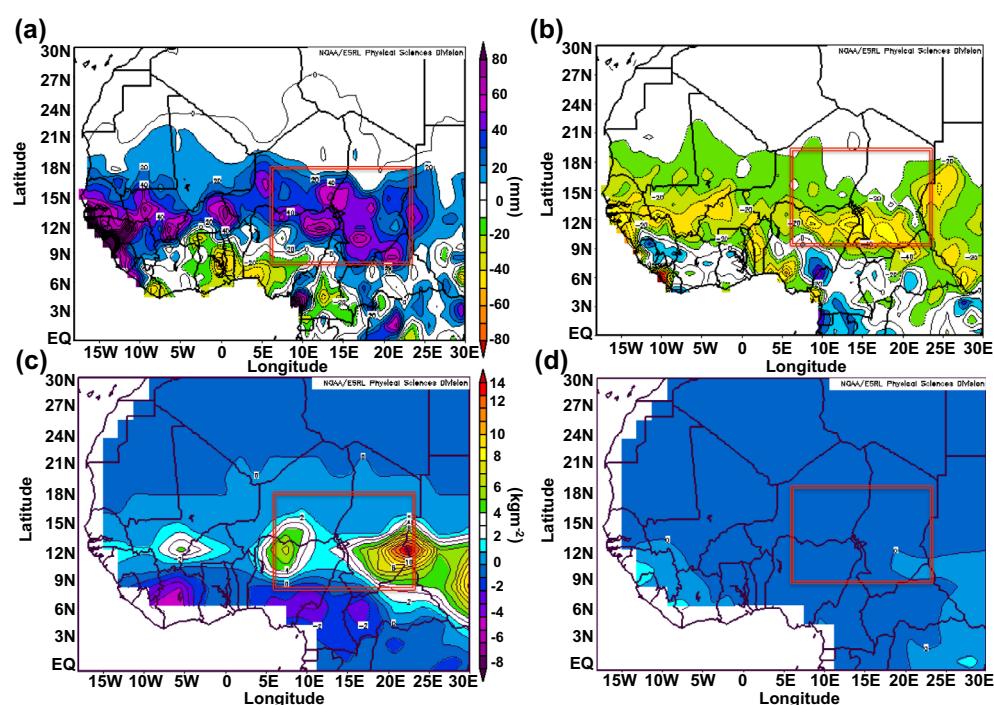
Our results thus suggest that LC level variability is very sensitive to rainfall variability and surface hydrological processes. A combination of increased rainfall and run-off is mirrored by LC level increase from 281 m in the late 1950s to about 281.5 m in the early 1960s (Figure 5). On the other hand, the immediate hydrological impact of decreased precipitation and run-off includes a decrease in the LC level from about 280.6 m in the mid-1970s to about 278.1 m in the early 1990s (Figure 5). Also, a decrease in the inundated areas of LC during that period has been reported (e.g. Lemoalle et al., 2012). Our result is also in agreement with similar studies on the relationship between the variability of water levels in continental lakes and global climate changes (Mercier, Cazenave, & Maheu, 2002). It also confirms the role of river discharge and precipitation in modulating lake levels, as reported by Hwang, Peng, Ning, Luo, and Sui (2005). We should however state that although these results are very convincing, they are not yet fully confirmed due to the absence of evaporation and LC surface inflow data for the study period. Full LC water balance modeling is needed to further confirm some of these results.

3.4. Possible anthropogenic effects

With LC as a closed lake, we should expect corresponding increases in the LC level with increasing precipitation in the absence of anthropogenic impact. But, we know that the population per unit area in southern LCB has been increasing steadily over the last several decades (Okonkwo & Demoz, 2014). The role of anthropogenic pressure through water abstraction and diversion for irrigation and agriculture has thus been suggested as a possible explanation for the physical disconnect between the decreasing LC level and increasing precipitation shown in Figure 5 (grey box 10) (e.g. Gao et al., 2010; Lemoalle et al., 2012; Okonkwo & Demoz, 2014).

Figure 10. Precipitation anomalies (mm) for (a) 1959–1962 and (b) 1982–1985 and run-off anomalies (kg s^{-2}) for (c) 1959–1962 and (d) 1982–1985.

Notes: The anomalies are with respect to the 1981–2010 mean. The red square is the LCB.



The lake, according to Gao et al. (2010), has not returned to its pre-1960 levels even as precipitation has been increasing since the 2000s due to a tremendous increase in irrigation diversions. This anthropogenic pressure has tilted the natural dynamics and is expected to dominate the LCB in the near future (Okonkwo & Demoz, 2014). The combined effect of decreased precipitation during the next cold phase of AMO and increase in anthropogenic pressure on water resources thus points to a potential mega-water crisis that could pose agricultural, environmental, and health problems to the regional population.

4. Discussion and conclusions

An understanding of the past and present influences of the combined AMO and ENSO modes in modulating regional climate dynamics in the Sahel region of Africa is critical to the regional seasonal rainfall forecasting. We have shown that the severe drought of the 1970s and 1980s is tied to the combined effect of the cold phase of AMO and warm phase of ENSO. We also established that the fluctuations in the LC level since the early 1970s have followed precipitation in the region. Further analyses show that the current warm phase of AMO appears to be playing a major role in the increased precipitation in the Sahel region. The LC level is responding to this increase in precipitation by a gradual recovery, though still below the pre-1960s level.

These results have thus improved our understanding of the combined effect of change in phase of the AMO from warm to cold in the mid-1960s and increases in El Niño intensity in the 1980s on atmospheric dynamics that are unfavorable to precipitation formation. Of particular importance are the temporal characteristics of the hydrological impacts. Our results thus suggest that the negative impact of the cold phase of AMO on the Sahel precipitation masked and superseded the positive effect of La Niña in the early 1970s. This is in line with Mohino, Janicot, and Bader (2011), which reported that the change to a negative phase of the AMO explains about 50% of the SST-driven Sahel drought in the 1980s. The marked decrease in LC in the 1970s and 1980s due to decreased precipitation can thus be attributed to the cold phase of AMO. The decrease in precipitation was further enhanced by the warm phase of ENSO. This understanding of the AMO-ENSO-rainfall-LC level association will enhance the predictability of the impacts of similar combined episodes in the future.

The understanding of this multi-decadal variability of Sahel precipitation also has some societal implications. For example, if this association had been predicted before the devastating drought years of the 1970s and 1980s, perhaps the region could have prepared more adequately. There are however two questions that should be receiving greater attention from researchers and policy-makers in this region: (1) Why is the precipitation for the warm phase of AMO (2000–2012; Figure 7) still below the 1900–1950s level? and (2) Why does the LC level increase (recovery) appear to be slowing down with increasing precipitation from the mid-2000s (Figure 5)? It is still unclear whether these apparent physical disconnects between rainfall and AMO, on one hand, and LC level variability and precipitation, on the other hand, are a trend of long-term change or an abrupt change. The negative rainfall anomaly for the 2000–2012 compared to the positive rainfall anomaly for the 1930–1959 (with respect to the 1950–1979 mean) should be a concern to water resources managers in the LC region. This is because we are already about mid-way into the warm phase of AMO; in about 15–20 years, the cold phase of AMO will return and with projected increase in the intensity of El Niño, the prospect of another round of severe drought in this region is increasingly likely.

Finally, while emphasizing the importance of AMO and ENSO in influencing the LC level variability, it should be pointed out that a much deeper understanding of the effect of other oceanic conditions like the Indian Ocean and the Mediterranean on the Sahel precipitation at different timescales is needed for a complete picture. Also, the absence of consistent long-term observations of LC level variability and inflow has made it difficult to be exact in studies like this. However, the recognition of the negative response of the LC level to the combined effect of cold phase of AMO and El Niño event established in this study will be beneficial to forecasters and help the LC region adequately prepare for future dry periods.

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References

- Birkett, C. M. (2000). Synergistic remote sensing of Lake Chad: Variability of basin inundation. *Remote Sensing of Environment*, 72, 218–236.
- Bjerknes, J. (1969). Atmospheric teleconnections from the equatorial Pacific. *Monthly Weather Review*, 97, 163–172.

[http://dx.doi.org/10.1175/1520-0493\(1969\)097<0163:ATFTEP>2.3.CO;2](http://dx.doi.org/10.1175/1520-0493(1969)097<0163:ATFTEP>2.3.CO;2)

Bunge, L., & Clarke, A. J. (2009). A verified estimation of the El Niño Index Niño-3.4 since 1877. *Journal of Climate*, 22(14), 3979–3992. <http://dx.doi.org/10.1175/JCLI2724.1>

Caminade, C., & Terray, L. (2010). Twentieth century Sahel rainfall variability as simulated by the ARPEGE AGCM, and future changes. *Climate Dynamics*, 35, 75–94. <http://dx.doi.org/10.1007/s00382-009-0545-4>

Chen, T. C., & van Loon, H. (1987). Interannual variation of the tropical easterly jet. *Monthly Weather Review*, 115, 1739–1759.

Chiang, J. C., & Sobel, A. H. (2002). Tropical tropospheric temperature variations caused by ENSO and their influence on the remote tropical climate. *Journal of Climate*, 15, 2616–2631.

Chung, C. E., & Ramanathan, V. (2006). Weakening of North Indian SST gradients and the monsoon rainfall in India and the Sahel. *Journal of Climate*, 19, 2036–2045. <http://dx.doi.org/10.1175/JCLI3820.1>

Climate Prediction Center. (2013). *Description of changes to oceanic Niño indexes*. Retrieved December 12, 2013, from http://www.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ONI_change.shtml

Contoux, C., Jost, A., Ramstein, G., Sepulchre, P., Krinner, G., & Schuster, M. (2013). Megalake Chad impact on climate and vegetation during the late Pliocene and the mid-Holocene. *Climate of the Past*, 9, 1417–1430.

Druyan, L. M. (2011). Studies of 21st-century precipitation trends over West Africa. *International Journal of Climatology*, 31, 1415–1424. <http://dx.doi.org/10.1002/joc.v31.10>

Enfield, D. B., Mestas-Nuñez, A. M., & Trimble, P. J. (2001). The Atlantic multidecadal oscillation and its relation to rainfall and river flows in the continental US. *Geophysical Research Letters*, 28, 2077–2080. <http://dx.doi.org/10.1029/2000GL012745>

Gao, H., Bohn, T. J., Podest, E., McDonald, K. C., & Lettenmaier, D. P. (2010). On the causes of the shrinking of Lake Chad. *Environmental Research Letters*, 6, 034021.

Geerken, R., Vassolo, S., & Bila, M. (2010, December 6–9). Impacts of climate variability and population pressure on water resources in the Lake Chad Basin. In The global dimensions of change in river basins. Conference of the Global Catchment Initiative (GCI) (p. 116). Bonn.

Giannini, A., Biasutti, M., & Verstraete, M. M. (2008). A climate model-based review of drought in the Sahel: Desertification, the re-greening and climate change. *Global and Planetary Change*, 64, 119–128. <http://dx.doi.org/10.1016/j.gloplacha.2008.05.004>

Giannini, A., Saravanan, & Chang, P. (2003). Oceanic forcing of Sahel rainfall on interannual to interdecadal time scales. *Science*, 302, 1027–1030. <http://dx.doi.org/10.1126/science.1089357>

Goddard, L., DeWitt, D. G., & Reynolds, R. W. (2009). Practical implications of uncertainty in observed SSTs. *Geophysical Research Letters*, 36, L09710.

Goddard, L., Mason, S., Zebiak, S., Ropelewski, C., Basher, R., & Cane, M. (2001). Current approaches to seasonal-to-interannual climate predictions. *International Journal of Climatology*, 21, 1111–1152. [http://dx.doi.org/10.1002/\(ISSN\)1097-0088](http://dx.doi.org/10.1002/(ISSN)1097-0088)

Grist, J. P., & Nicholson, S. E. (2001). A study of the dynamic factors influencing the rainfall variability in the West African Sahel. *Journal of Climate*, 14, 1337–1359.

Hall, N. M., Kiladis, J. G. N., & Thorncroft, C. D. (2006). Three-dimensional structure and dynamics of African easterly waves. Part II: Dynamical modes. *Journal of the Atmospheric Sciences*, 63, 2231–2245.

Hwang, Y. T., Frierson, D. M., & Kang, S. M. (2013). Anthropogenic sulfate aerosol and the southward shift of tropical precipitation in the late 20th century. *Geophysical Research Letters*, 40, 2845–2850.

- Hwang, C., Peng, M. F., Ning, J., Luo, J., & Sui, C. H. (2005). Lake level variations in China from TOPEX/Poseidon altimetry: Data quality assessment and links to precipitation and ENSO. *Geophysical Journal International*, 161(1), 1–11.
- IPCC. (2007). Contribution of working groups I, II and III to the fourth assessment report of the intergovernmental panel on climate change (Climate change (2007) synthesis report (IPCC, Geneva), p. 104). Geneva: Author.
- Janicot, S. (1997). Impact of warm ENSO events on atmospheric circulation and convection over the tropical Atlantic and West Africa. *Annals in Geophysics*, 15, 471–475.
- Janicot, S., Mounier, F., Gervois, S., Sultan, B., & Kiladis, G. N. (2010). The dynamics of the West African monsoon. Part V: The detection and role of the dominant modes of convectively coupled equatorial Rossby waves. *Journal of Climate*, 23, 4005–4024.
- Janowiak, J. E. (1988). An investigation of interannual rainfall variability in Africa. *Journal of Climate*, 1, 240–255.
[http://dx.doi.org/10.1175/1520-0442\(1988\)001<0240:AIORIV>2.0.CO;2](http://dx.doi.org/10.1175/1520-0442(1988)001<0240:AIORIV>2.0.CO;2)
- Jenkins, G. S., Gaye, A. T., & Sylla, B. (2005). Late 20th century attribution of drying trends in the Sahel from the Regional Climate Model (RegCM3). *Geophysical Research Letters*, 32(22), pL22705.
- Joly, M., Volodire, A., Douville, H., Terray, P., & Royer, J. F. (2007). African monsoon teleconnections with tropical SSTs: Validation and evolution in a set of IPCC4 simulations. *Climate Dynamics*, 29(1), 1–20.
<http://dx.doi.org/10.1007/s00382-006-0215-8>
- Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., ... Joseph, D. (1996). The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American Meteorological Society*, 77, 437–471.
[http://dx.doi.org/10.1175/1520-0477\(1996\)077<0437:TNYRP>2.0.CO;2](http://dx.doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2)
- Kerr, R. A. (2000). A north Atlantic climate pacemaker for the centuries. *Science*, 288, 1984–1985.
<http://dx.doi.org/10.1126/science.288.5473.1984>
- Knight, J. R., Allan, R. J., Folland, C. K., Vellinga, M., & Mann, M. E. (2005). A signature of persistent natural thermohaline circulation cycles in observed climate. *Geophysical Research Letters*, 32, L20708. doi:[10.1029/2005GL024233](https://doi.org/10.1029/2005GL024233)
- Laing, A., & Evans, J. L. (2011). *Introduction to tropical meteorology*. In *Educational material from The COMET Program*. Boulder, CO: University Corporation for Atmospheric Research.
- Lauwaet, D., van Lipzig, N. P. M., Van Weverberg, K., De Ridder, K., & Goyens, C. (2011). The precipitation response to the desiccation of Lake Chad. *Quarterly Journal of Royal Meteorological Society*, 138, 707–719. doi:[10.1002/qj.942](https://doi.org/10.1002/qj.942)
- Leblanc, M., Lemoalle, J., Bader, J. C., Tweed, S., & Mofor, L. (2011). Thermal remote sensing of water under flooded vegetation: New observations of inundation patterns for the “small” Lake Chad”. *Journal of Hydrology*, 404, 87–98.
- Lemoalle, J., Bader, J. C., Leblanc, M., & Sedick, A. (2012). Recent changes in Lake Chad: Observations, simulations and management options (1973–2011). *Global and Planetary Change*, 80–81, 247–254.
<http://dx.doi.org/10.1016/j.gloplacha.2011.07.004>
- Lu, J. (2009). The dynamics of the Indian Ocean sea surface temperature forcing of Sahel drought. *Climate Dynamics*, 33, 445–460.
- Lu, J., & Delworth, T. L. (2005). Oceanic forcing of the late 20th century Sahel drought. *Geophysical Research Letters*, 32(22), L22706.
<http://dx.doi.org/10.1029/2005GL023316>
- Martin, E. R., & Thorncroft, C. D. (2013). The impact of the AMO on the West African monsoon annual cycle. *Quarterly Journal of the Royal Meteorological Society*, 140, 31–46.
- Martin, E. R., Thorncroft, C., & Booth, B. B. (2014). The multidecadal Atlantic SST–Sahel rainfall teleconnection in CMIP5 simulations. *Journal of Climate*, 27, 784–806.
- Mercier, F., Cazenave, A., & Maheu, C. (2002). Interannual lake level fluctuations (1993–1999) in Africa from Topex/Poseidon: Connections with ocean–atmosphere interactions over the Indian Ocean. *Global and Planetary Change*, 32, 141–163.
[http://dx.doi.org/10.1016/S0921-8181\(01\)00139-4](http://dx.doi.org/10.1016/S0921-8181(01)00139-4)
- Mohino, E., Janicot, S., & Bader, J. (2011). Sahel rainfall and decadal to multi-decadal sea surface temperature variability. *Climate Dynamics*, 37, 419–440.
<http://dx.doi.org/10.1007/s00382-010-0867-2>
- Nicholson, S. E. (2001). Climatic and environmental change in Africa during the last two centuries. *Climate Research*, 17, 123–144.
- Nicholson, S. E. (2008). The intensity, location and structure of the tropical rainbelt over West Africa as factors in interannual variability. *International Journal of Climatology*, 28, 1775–1785.
- Nicholson, S. E. (2009). On the factors modulating the intensity of the tropical rainbelt over West Africa. *International Journal of Climatology*, 29, 673–689.
<http://dx.doi.org/10.1002/joc.v29.5>
- Nicholson, S. E. (2013). The West African Sahel: A review of recent studies on the rainfall regime and its interannual variability. *Meteorology*, 2013, 453521, 32 p. doi:[10.1155/2013/453521](https://doi.org/10.1155/2013/453521)
- Nicholson, S. E., & Grist, J. P. (2003). The seasonal evolution of the atmospheric circulation over West Africa and equatorial Africa. *Journal of Climate*, 16, 1013–1030.
[http://dx.doi.org/10.1175/1520-0442\(2003\)016<1013:TEOTA>2.0.CO;2](http://dx.doi.org/10.1175/1520-0442(2003)016<1013:TEOTA>2.0.CO;2)
- Nicholson, S. E., Some, B., & Kone, B. (2000). An analysis of recent rainfall conditions in West Africa, including the rainy seasons of the 1997 El Niño and the 1998 La Niña years. *Journal of Climate*, 13, 2628–2640.
[http://dx.doi.org/10.1175/1520-0442\(2000\)013<2628:AAORRC>2.0.CO;2](http://dx.doi.org/10.1175/1520-0442(2000)013<2628:AAORRC>2.0.CO;2)
- Okonkwo, C. (2013). Integrating science and policy in the management of Lake Chad. *International Journal of Environmental Studies*, 70, 844–845.
<http://dx.doi.org/10.1080/00207233.2013.864750>
- Okonkwo, C., & Demoz, B. (2014). Identifying anthropogenic ‘hotspots’ and management of water resources in Lake Chad Basin using GIS. *Journal of Natural Resources Policy Research*, 6, 135–149.
- Okonkwo, C., Demoz, B., & Tesfai, S. (2014). Characterization of West African jet streams and their association to ENSO events and rainfall in ERA-Interim 1979–2011. *Advances in Meteorology*, 2014, 405617, 12 p. doi:[10.1155/2014/405617](https://doi.org/10.1155/2014/405617)
- Olivry, J. C., Chouaret, A., Vuillaume, G., Lemoalle, J., & Bricquet, J. P. (1996). *Hydrologie du lac Chad, Monographies Hydrologiques 12 [Hydrology of Lake Chad, ORSTOM, Monographies Hydrologiques 12]* (ORSTOM ed., Vol. 266). Paris.
- Paeth, H., & Stuck, J. (2004). The West African dipole in rainfall and its forcing mechanisms in global and regional climate models. *Mausam*, 55, 561–582.
- Polo, I., Ullmann, A., Roucou, P., & Fontaine, B. (2011). Weather regimes in the Euro-Atlantic and Mediterranean sector, and relationship with West African Rainfall over the 1989–2008 period from a self-organizing maps approach. *Journal of Climate*, 24, 3423–3432.
<http://dx.doi.org/10.1175/2011JCLI3622.1>
- Polo, I. B. R.-F., Rodriguez-Fonseca, B., & Losada, T. (2008). Tropical Atlantic variability modes (1979–2002). Part I: Time-evolving SST modes related to West African rainfall. *Journal of Climate*, 21, 6457–6475.
<http://dx.doi.org/10.1175/2008JCLI2607.1>
- Rose, M. A., & Darrell, R. M. (2009). *Distinguishing effects of El Niño Southern oscillation from short term climate variability* (Technical Memorandum NWS SR-231). Fort Worth, TX: US Department of Commerce, National Oceanic and

- Atmospheric Administration, National Weather Service.
- Schlesinger, M. E., & Ramankutty, N. (1994). An oscillation in the global climate system of period 65–70 years. *Nature*, 367, 723–726. <http://dx.doi.org/10.1038/367723a0>
- Sima, F., Kamga, A., Raiva, I., Dekaa, F. S., & James, A. I. (2013). State of the climate 2012: [Africa] West Africa. *Bulletin of the American Meteorological Society*, 94, S161–S166.
- Singh, A., Diop, S., & M'mayi, P.L. (2006). *Africa's lakes: Atlas of our changing environment*. Nairobi: UNEP.
- Slingo, J. M., & Annamalai, H. (2000). 1997: The El Niño of the century and the response of the Indian summer monsoon. *Monthly Weather Review*, 128, 1778–1797.
- Tarhule, A., & Lamb, P. J. (2003). Climate research and seasonal forecasting for West Africans: Perceptions, dissemination, and use? *Bulletin of the American Meteorological Society*, 84, 1741–1759. <http://dx.doi.org/10.1175/BAMS-84-12-1741>
- Thornicroft, C. D., Nguyen, H., Zhang, C., & Peyrillé, P. (2011). Annual cycle of the West African monsoon: Regional circulations and associated water vapour transport. *Quarterly Journal of the Royal Meteorological Society*, 137, 129–147. <http://dx.doi.org/10.1002/qj.728>
- Ting, M. F., Kushnir, Y., Seager, R., & Li, C. H. (2009). Forced and internal twentieth-century SST trends in the North Atlantic. *Journal of Climate*, 22, 1469–1481. [http://dx.doi.org/10.1175/2008JCLI2561.10442\(2001\)014<1697:LIOENO>2.0.CO;2](http://dx.doi.org/10.1175/2008JCLI2561.10442(2001)014<1697:LIOENO>2.0.CO;2)
- Trenberth, K. E., & Hoar, T. J. (1997). El Niño and climate change. *Geophysical Research Letters*, 24, 3057–3060. <http://dx.doi.org/10.1029/97GL03092>
- Trenberth, K. E., Jones, P. D., Ambenje, P., Bojariu, R., Easterling, D., Klein Tank, A., ... Zhai, P. (2007). Observations: Surface and atmospheric climate change. In S. Solomon, D. Qin, M. Manning, Z. Chen, M. Marquis, K. B. Averyt, ... H. L. Miller (Eds.), *Climate change 2007: The physical science basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* (pp. 235–336). Cambridge: Cambridge University Press.
- Trenberth, K. E., & Stepaniak, D. P. (2001). Indices of El Niño evolution. *Journal of Climate*, 14, 1697–1701. <http://dx.doi.org/10.1175/1520-2523.14.11.1697>
- Vallet-Coulomb, C., Legesse, D., Gasse, F., Travi, Y., & Chernet, T. (2001). Lake evaporation estimates in tropical Africa (Lake Ziway, Ethiopia). *Journal of Hydrology*, 245(1–4), 1–18. [http://dx.doi.org/10.1016/S0022-1694\(01\)00341-9](http://dx.doi.org/10.1016/S0022-1694(01)00341-9)
- Vizy, E. K., & Cook, K. H. (2001). Mechanisms by which Gulf of Guinea and eastern north Atlantic sea surface temperature anomalies can influence African rainfall. *Journal of Climate*, 14, 795–821. [http://dx.doi.org/10.1175/1520-0442\(2001\)014<0795:MBWGOG>2.0.CO;2](http://dx.doi.org/10.1175/1520-0442(2001)014<0795:MBWGOG>2.0.CO;2)
- Ward, M. N. (1998). Diagnosis and short-lead time prediction of summer rainfall in tropical North Africa at interannual and multidecadal timescales. *Journal of Climate*, 11, 3167–3191. [http://dx.doi.org/10.1175/1520-0442\(1998\)011<3167:DASLTP>2.0.CO;2](http://dx.doi.org/10.1175/1520-0442(1998)011<3167:DASLTP>2.0.CO;2)
- Yeh, S. W., & Kirtman, B. P. (2005). Pacific decadal variability and decadal ENSO amplitude modulation. *Geophysical Research Letters*, 32(5), L05703. doi:10.1029/2004GL021731
- Zhang, R., & Delworth, T. L. (2006). Impact of Atlantic multidecadal oscillations on India/Sahel rainfall and Atlantic hurricanes. *Geophysical Research Letters*, 33(17), L17712. doi:10.1029/2006GL026267



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