Interdecadal Changes in the Leading Ocean Forcing of Sahelian Rainfall Interannual Variability: Atmospheric Dynamics and Role of Multidecadal SST Background

ROBERTO SUÁREZ-MORENO AND BELÉN RODRÍGUEZ-FONSECA

Department of Geophysics and Meteorology, Universidad Complutense de Madrid, and Instituto de Geociencias, Universidad Complutense de Madrid-CSIC, Madrid, Spain

JESÚS A. BARROSO

Department of Geophysics and Meteorology, Universidad Complutense de Madrid, Madrid, Spain

ANDREAS H. FINK

Institute of Meteorology and Climate Research, Karlsruhe Institute of Technology, Karlsruhe, Germany

(Manuscript received 31 May 2017, in final form 6 April 2018)

ABSTRACT

The atmospheric response to global sea surface temperatures is the leading cause of rainfall variability in the West African Sahel. On interannual periodicities, El Niño–Southern Oscillation, the Atlantic equatorial mode, and Mediterranean warm/cold events primarily drive variations of summer rainfall over the Sahel. Nevertheless, the rainfall response to these modes of interannual SST variability has been suggested to be unstable throughout the observational record. This study explores changes in the leading patterns of co-variability between Sahel rainfall and SSTs, analyzing the dynamical mechanisms at work to explain the nonstationary relationship between anomalies in these two fields. A new network of rain gauge stations across West Africa is used for the first time to investigate these instabilities during the period 1921–2010. A hypothesis is raised that the underlying SST background seems to favor some interannual teleconnections and inhibit others in terms of the cross-equatorial SST gradients and associated impacts on the location of the intertropical convergence zone. Results of this study are relevant for improving the seasonal predictability of summer rainfall in the Sahel.

1. Introduction

The West African Sahel is the semiarid transition region between the Sahara Desert to the north and the humid savannas to the south (e.g., [Nicholson 2000](#)). It stretches across northern tropical Africa from Senegal to Niger (10°–18°N, 15°W–15°E). The West African monsoon (WAM) characterizes the hydrological cycle in this region, being driven by a south–north–south

displacement of the intertropical convergence zone (ITCZ; e.g., [Sultan and Janicot 2000](#)), a band of deep convective clouds in the tropics that defines the region of maximum precipitation. Recent studies demonstrate that the ITCZ tends to shift in the opposite direction to the atmospheric energy transport across the equator, which in turn is generally reinforced toward the winter hemisphere, to partially compensate for this hemisphere's radiative cooling (e.g., [Frierson and Hwang 2012](#); [Bischoff and Schneider 2014](#)). On average, the ITCZ over West Africa abruptly shifts from the Guinea coast (4°–6°N) to about 10°–12°N at the end of June ([Sultan and Janicot 2000](#); [Okumura and Xie 2004](#)), determining the peak of rainfall in the Sahel during July–September (JAS), over which the variability of precipitation is also maximum ([Fig. 1a](#)).

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-17-0367.s1>.

Corresponding author: Roberto Suárez-Moreno, roberto.suarez@fis.ucm.es

FIG. 1. Sahel rainfall seasonal cycle and indices of variability. (a) Latitude–time diagram of rainfall amounts longitudinally averaged (15°W – 15°E) (shaded) and its standard deviation (contours) using the CRU database (see section 2). Solid green horizontal lines indicate the Sahelian latitudinal domain (10° – 18°N). Solid red vertical lines denote the JAS season. The axis with the months indicates the central day of each month. (b) The top panel shows the standardized index of anomalous rainfall (mm day^{-1}) in the Sahel (red line), together with its low-frequency (green line) and high-frequency (blue bars) components obtained by applying a Butterworth low-pass (cutoff frequency of $2/13 \text{ yr}^{-1}$) and high-pass (cutoff frequency of $2/7 \text{ yr}^{-1}$) filter, respectively. The seasonal anomalies are calculated for the Sahel spatial domain (15°W – 15°E , 10° – 18°N) for JAS with respect to the 1902–2013 climatology using the CRU database. Shown in the bottom panel is the variability of standard deviation calculated for the high-frequency component of the standardized rainfall (mm day^{-1}) index. The anomalous standard deviation (black line) is calculated as 21-yr running means of the high-frequency index (blue bars) minus the averaged standard deviation for the whole time series of the same index.

Low-frequency rainfall variability in the Sahel ranges from interannual to decadal time scales, with periods of abrupt year-to-year fluctuations and decades of pronounced trends toward either drought or humid conditions. The pronounced variability of the WAM system makes the population in the Sahel highly vulnerable to climate-related impacts (e.g., Sultan and Gaetani 2016). Changes in the interannual and decadal Sahel rainfall

variability are depicted in Fig. 1b. Noticeably, the standard deviation of interannual variability changes in the observational record, with some decades of reduced amplitude of the anomalies and some others in which the amplitude increases (Fig. 1b, bottom panel). A trend toward an increased amplitude of interannual variability is observed. Beyond the ITCZ, the leading circulation features associated with the variability of Sahelian

rainfall on interannual and decadal time scales are the Saharan heat low (SHL; e.g., Lavaysse et al. 2009, 2010), the African easterly jet (AEJ), the tropical easterly jet (TEJ), and the low-level westerly jets over the Atlantic (e.g., Grist and Nicholson 2001; Nicholson 2009a; Lafore et al. 2011; Fink et al. 2017). If any of these key WAM components is perturbed, the seasonal cycle of rainfall in the Sahel can be disrupted (e.g., Nicholson and Webster 2007; Nicholson et al. 2008; Nicholson 2009b).

Oceanic forcing is a dominant driver of interannual-to-decadal WAM variability (e.g., Folland et al. 1986; Rowell et al. 1995), altering the aforementioned WAM atmospheric components. Oceans are able to store large amounts of heat in the surface layer to be transferred to the atmosphere, affecting the atmospheric circulation and stability with resultant changes in precipitation patterns (Polo et al. 2008; Stockdale et al. 2010; Rodríguez-Fonseca et al. 2011, 2015; Rowell 2013). Sea surface temperature anomalies (SSTs) are organized in so-called modes of variability, which are specific for each ocean basin and are used to explain large-scale patterns of mostly intrabasin sea surface temperature (SST) variability (e.g., Moron et al. 1998; Cai and Whetton 2001; von Storch and Zwiers 2002). When it comes to Sahel rainfall, the influence of global SSTs is often split into interannual and decadal periodicities (e.g., Giannini et al. 2003; Diatta and Fink 2014). On interannual time scales, El Niño–Southern Oscillation (ENSO; Janicot et al. 2001; Rowell 2001; Giannini et al. 2005; Joly and Voldoire 2009; Mohino et al. 2011a), the Atlantic equatorial mode (AEM, also known as Atlantic Niño; Janicot et al. 1998; Giannini et al. 2003; Polo et al. 2008; Joly and Voldoire 2010; Losada et al. 2010), and SSTs in the Mediterranean Sea (Rowell 2003; Jung et al. 2006; Fontaine et al. 2010, 2011; Gaetani et al. 2010) have been shown to consistently impact the WAM dynamics. On longer time scales, meridional migrations of the ITCZ respond to changes of interhemispheric temperature contrasts (e.g., Broccoli et al. 2006; Kang et al. 2008), with the global ocean SST background playing a prominent role (Chiang et al. 2000).

Despite being broadly described in the literature, the atmospheric dynamics of interannual teleconnections between summer Sahel rainfall and the aforementioned patterns of SST variability is far from stable throughout the observational record. This feature has been described in terms of strengthened or weakened teleconnections and associated impacts depending on the sequence of decades under study (Janicot et al. 1996; Fontaine et al. 1998; Mohino et al. 2011c; Rodríguez-Fonseca et al. 2011, 2015, 2016; Losada et al. 2012; Diatta and Fink 2014). In this context, Janicot et al. (1996) suggested changes in the association between

eastern tropical Atlantic and Pacific Ocean basins and rainfall in West Africa after the 1970s. Later, Fontaine et al. (1998) pointed out a time evolution in the SSTA–rainfall links due to interactions between modes of tropical SST variability obtained from discriminant analysis techniques. Moreover, Rodríguez-Fonseca et al. (2011) discussed changes in the interannual teleconnection patterns between tropical SSTs and precipitation in West Africa. Similarly, Mohino et al. (2011b) found differences in the interannual SST-forced response of rainfall in the Sahel before and after the 1970s, suggesting that, in turn, this could be the result of the nonstationary behavior in the teleconnection between tropical Atlantic and Pacific SSTA covariability modes (Rodríguez-Fonseca et al. 2009). Nevertheless, the evidence of an unstable ocean impact on the WAM remains almost entirely observational, with the dynamical factors causing the varying strengths of interannual teleconnections not having been analyzed so far.

This work investigates the instability (nonstationarity) of the SSTA-forced responses of anomalous Sahelian rainfall from the tropical Atlantic, tropical Pacific, and Mediterranean Sea on interannual time scales. In this framework, the study is focused on analyzing (i) interdecadal changes in the leading modes of interannual SSTA–anomalous rainfall variability in each of the above-mentioned ocean basins and related impacts on the key atmospheric features of the WAM and (ii) the influence of the varying global-scale SST background on decadal time scales on the general atmospheric circulation in terms of the SSTA-induced migrations of the ITCZ. The nonstationarities are statistically assessed for the period 1921–2010 by using the SST-based statistical seasonal forecast (S^4 CAST; Suárez-Moreno and Rodríguez-Fonseca 2015, hereinafter SR15) tool, which is described together with data in section 2. The results are presented in sections 3 and 4, where the respective dynamical mechanisms associated with points i and ii are explored. Section 5 covers the discussion and conclusions.

2. Data and methodology

This work is based on the use of observations from individual stations and gridded surface data, as well as three-dimensional reanalysis data. Observations serve to explore covariability patterns between anomalous SSTs and rainfall, whereas reanalysis data are used to examine the dynamical mechanisms at work. To validate the results, a network of in situ rainfall measurements is used together with gridded precipitation data.

Rain gauge measurements across West Africa used in this study comprised 167 stations with daily data and

254 stations with monthly records. Long-term monthly rainfall time series between 1921 and 2010 were extracted from 67 stations with at least 90% data availability. These observations were provided from the database of the Institute of Geophysics and Meteorology of the University of Cologne (Germany). Monthly data can be downloaded from <http://www.geomet.uni-koeln.de/en/the-institute/data/>. Further description of the station data is provided in [Sanogo et al. \(2015\)](#). Stations with 100% data availability were selected for this study, yielding a total of 53 stations. As an alternative source of precipitation data, the gridded dataset from the Climatic Research Unit (CRU TS3.22) was used for the period 1901–2013 on a $0.5^\circ \times 0.5^\circ$ latitude–longitude grid ([Harris et al. 2014](#)). The statistical–observational analysis was carried out based on the longest period (1921–2010) common to both datasets.

The region of study is the Sahel, with the spatial domain corresponding to that defined in [Diallo et al. \(2013\)](#) (10° – 18° N, 15° W– 15° E). This area is a core study region within the African Monsoon Multidisciplinary Analysis (AMMA) project ([Redelsperger et al. 2006](#)) and contains the highest number of rainfall stations (30) with 100% data availability.

The Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST; [Rayner et al. 2003](#)) is used for the SST field. This dataset consists of monthly means of SSTs with a resolution of $1.0^\circ \times 1.0^\circ$ spanning the period of January 1870–September 2013. The HadISST data are reconstructed using a two-stage reduced-space optimal interpolation procedure, followed by superposition of quality-improved gridded observations onto the reconstructions to restore local detail. Extended information about the data and a link for downloading can be found at <http://www.metoffice.gov.uk/hadobs/hadisst/>.

The selected spatial domains considered for analyzing the SST's impact on rainfall correspond to those ocean basins known to consistently impact WAM variability on interannual time scales (e.g., [Rodríguez-Fonseca et al. 2011, 2015](#)). These regions are the tropical Pacific Ocean (tPAC; 170° E– 95° W, 30° S– 20° N), the tropical Atlantic Ocean (tATL; 60° W– 20° E, 20° S– 10° N), and the eastern Mediterranean Sea (eMED; 16° – 38° E, 30° – 40° N). Specifically, the eastern Mediterranean Sea is the region for which a robust teleconnection with Sahelian rainfall has been defined on interannual time scales (e.g., [Fontaine et al. 2010](#); [Gaetani et al. 2010](#); [Diatra and Fink 2014](#)).

For the dynamical analysis, the atmospheric reanalysis of the twentieth century (ERA-20C) from the European Centre for Medium-Range Weather Forecasts (ECMWF) was used ([Poli et al. 2016](#)). This

database provides a long gridded dataset of climate variables at different pressure levels, spanning the period 1900–2010. A coupled atmosphere–land surface–ocean wave model is used to reanalyze the weather by assimilating surface observations. The observations assimilated in ERA-20C include surface pressures and mean sea level pressures. In particular, the zonal, meridional, and vertical wind components (u , v , and w , respectively), geopotential height z , low-level specific humidity q , temperature T , and mean sea level pressure (slp) fields were used. All these data can be downloaded from the ECMWF website: <http://www.ecmwf.int/en/research/climate-reanalysis/era-20c>. Quality concerns for ERA-20C in the first part of the twentieth century can be raised, even though it is assumed that the planetary-scale, JAS mean anomalies are reasonably reflected. The remote impact of the eastern equatorial Pacific on the Sahel involves alterations in the upper-level atmospheric circulation, being further explored here by means of the large-scale upper-level divergent and rotational circulations represented by the velocity potential χ and streamfunction ψ , respectively. These global-scale potentials are computed from horizontal wind components (u , v) (e.g., [Bijlsma et al. 1986](#)) at different pressure levels.

The anomalous fields are calculated by subtracting the seasonal long-term mean from the three-month season (July–September). Rainfall anomalies are standardized since the seasonal distribution of precipitation follows a non-normal distribution. A high-pass Butterworth filter of tenth order is applied with a cutoff frequency given by $f_c = 2dt/T$, where dt is the sample interval and T is the periodicity to be filtered. Thus, $f_c = 2/7 \text{ yr}^{-1}$ is applied to the time series of rainfall and SSTA in order to extract the purely interannual variability, isolating the high-frequency influence of SSTA on anomalous rainfall to explore the associated leading covariability patterns. The cutoff period of seven years, chosen here for the Butterworth high-pass filter, is consistent with other studies that tried to isolate the leading modes of interannual SST variability such as the AEM (e.g., [Polo et al. 2008](#)) or ENSO (e.g., [Clarke 2014](#); [Wang et al. 2012](#)), with a periodicity rarely exceeding that period.

A series of simulations with the S⁴CAST model were conducted to explore the leading modes of covariability between anomalous summer Sahel rainfall and SSTA in eMED, tATL, and tPAC. The associated predictability is evaluated applying a cross-validated hindcast. The S⁴CAST tool is based on the maximum covariance analysis (MCA) method, a discriminant analysis technique widely used to explore the covariability patterns between two climate fields (e.g., [Bretherton et al. 1992](#); [Wallace et al. 1992](#); [Lau and Nath 1994](#); [Cherry 1996](#);

Widmann 2005). The MCA calculates the principal directions in which the covariance matrix $\mathbf{C}$ between a given predictor field $\mathbf{Y}$ and a field to be predicted $\mathbf{Z}$ is maximized. Both fields are defined on time n_t and space domains (n_Y and n_Z for $\mathbf{Y}$ and $\mathbf{Z}$, respectively), and $\mathbf{C}$ is a nonsquare covariance matrix calculated as follows:

$$\mathbf{C} = \frac{\mathbf{Y}\mathbf{Z}^T}{(n_t - 1)}. \quad (1)$$

The matrix $\mathbf{C}$ is diagonalized and represented in terms of covariability modes between $\mathbf{Z}$ and $\mathbf{Y}$ (e.g., anomalous rainfall and SSTA, respectively, in the present study). In a forecasting framework, a linear model is used to obtain an estimation of $\mathbf{Z}$, denoted as $\hat{\mathbf{Z}}$, which is expressed as follows:

$$\hat{\mathbf{Z}} = \phi \mathbf{Y}, \quad (2)$$

where the regression coefficient ϕ is calculated according to the expression

$$\phi = \mathbf{R}(\mathbf{U}\mathbf{U}^T)^{-1}\mathbf{U}\mathbf{Z}^T. \quad (3)$$

In (3), $\mathbf{U}$ represents the time series of the expansion coefficients of the predictor $\mathbf{Y}$, computed as the singular value for a given mode of covariability n , whereas its corresponding singular vector $\mathbf{R}$ represents a spatial pattern obtained by regressing $\mathbf{U}$ onto its respective field $\mathbf{Y}$. The results of this study are focused on the leading MCA mode (i.e., $n = 1$).

The linear model uses ϕ to obtain cross-validated hindcasts $\hat{\mathbf{Z}}$. Cross validation is applied following the leave-one-out method (e.g., Dayan et al. 2014), according to which, for a given time step i within the temporal dimension n_t , defined as $i = 1, 2, 3, \dots, n_t$, the data corresponding to the observed predictand $\mathbf{Z}_i = \mathbf{Z}(n_t, n_Z)$ and predictor $\mathbf{Y}_i = \mathbf{Y}(n_t, n_Y)$ are removed from the whole time series. Next, the MCA is applied using the remaining data from both predictor and predictand fields, obtaining the so-called cross-covariance matrix as in (1). Next, the regression coefficients ϕ_i are calculated as in (3), being introduced in (2) to obtain an estimate $\hat{\mathbf{Z}}_i$ of the removed predictand $\mathbf{Z}_i$ from a predictor $\mathbf{Y}_i$ that was also initially removed. This process is applied successively for each time step, resulting in cross-validated hindcasts for the whole time series. Then, the skill score of the model is calculated by means of Pearson correlation coefficients ρ between observations and hindcasts to assess qualitatively the predictability of $\hat{\mathbf{Z}}$:

$$\rho_{\mathbf{ZZ}} = \frac{\text{cov}(\mathbf{Z}, \hat{\mathbf{Z}})}{\sigma_{\mathbf{Z}} \sigma_{\hat{\mathbf{Z}}}}, \quad (4)$$

where cov denotes covariance between $\mathbf{Z}$ and $\hat{\mathbf{Z}}$, and σ is the standard deviation.

As pointed out in SR15, the S⁴CAST model uses the methodology described above in a novel way to statistically evaluate periods of potential stability (stationarity) in the relationship between $\mathbf{Y}$ and $\mathbf{Z}$. These periods are determined by means of 21-yr sliding window correlations between the time series of $\mathbf{Y}$ and $\mathbf{Z}$ ($\mathbf{U}$ and $\mathbf{V}$, respectively) for an isolated mode of covariability. Stationary (stable) periods correspond to those years in which the association between $\mathbf{U}$ and $\mathbf{V}$ remains invariant in terms of significant or nonsignificant correlation scores between time series. In this study, the correlation index (herein COI) between $\mathbf{U}$ and $\mathbf{V}$ is defined as in (4) for each time step i within the temporal dimension n_t :

$$\text{COI}(i) = \frac{\text{cov}[\mathbf{U}(i - 20:i), \mathbf{V}(i - 20:i)]}{\sigma_{\mathbf{U}(i-20:i)} \sigma_{\mathbf{V}(i-20:i)}}. \quad (5)$$

A significance test is applied to determine the significant correlation (SC) period over which COI scores are statistically significant at a given confidence level. Likewise, the nonsignificant correlation (NSC) period corresponds to the remaining COI scores below the same significance level. To better assess the current period, 21-yr-delayed (1 year and the 20 previous years) windows correlation has been selected as pointed out in (5). From (5), the transition between SC and NSC periods occurs for a given time step i considering that COI is statistically significant for its corresponding window ($i - 20:i$) but not for the previous one ($i - 21:i - 1$).

The S⁴CAST model performs the significance tests according to the nonparametric Monte Carlo method under a given number of random permutations. In the present study, the significance level has been set to 95% ($\alpha = 0.05$). A full description of mathematical developments as well as different possibilities of application of the model and two detailed benchmark cases can be found in SR15.

The S⁴CAST model is used to explore the potential stationarity (stability) in the leading mode of covariability between SSTA in different oceanic regions and anomalous Sahelian rainfall during JAS. In this framework, the model is applied to those oceanic regions that mainly impact rainfall on interannual time scales: eMED, tATL, and tPAC. The results are validated using both gridded and station data as rainfall fields in each simulation, performing a total of six configurations. For each configuration, the statistical methodology is applied under two assumptions:

- (i) During the extended period (1921–2010), the $\mathbf{Y}$ – $\mathbf{Z}$ covariability is nonstationary (nonstable).

- (ii) During SC and NSC periods, the **Y–Z** covariability is significant and nonsignificant, respectively, yet stationary (stable).

As stated above, the forecast period is selected to cover the peak of the monsoon season in the Sahel (JAS).

The atmospheric dynamics underlying the different modes of covariability are explored by means of seasonal composites from ERA-20C reanalysis atmospheric variables. Composite maps are calculated by subtracting the averaged JAS anomalous fields for a set of years during which the expansion coefficient exceeds a given threshold. In this context, high (denoted by H) refers to those years with the predictor expansion coefficient **U** exceeding one positive standard deviation, whereas low (L) corresponds to **U** scores below one negative standard deviation. Composites are independently calculated for SC and NSC periods to compare the changing atmospheric response under a potentially different SSTA pattern. To evaluate the statistical significance of composite maps, a parametric *t* test of the difference between two means is applied, setting the confidence level at 90% ($\alpha = 0.10$).

3. Stationarity of the leading interannual covariability modes

In this section, the leading covariability modes of the “SSTA–anomalous Sahel rainfall” relation are analyzed for eMED, tATL, and tPAC from the S⁴CAST simulations. These modes are analyzed considering the SC and NSC periods statistically evaluated from COI. The dynamical mechanisms underlying the covariability patterns in both periods are subsequently addressed.

The expansion coefficients time series (**U**, **V**) along with COI indices are shown in Fig. 2 corresponding to the “HadISST CRU” and “HadISST station” configurations. The COI associated with eMED (Fig. 2a) presents SC periods before the 1960s and after the 1990s. By construction, the complementary NSC period lies in between these decades. For the tATL (Fig. 2b), the SC period extends up to the end of the 1980s, with the NSC period lying in recent decades. For the tPAC (Fig. 2c), SC spans from the 1960s onward until 2010, so that NSC is restricted to the 1940s and 1950s. Additionally, an Ebisuzaki test (Ebisuzaki 1997) was applied to assess the autocorrelation (i.e., to account for serial correlations) in the calculation of SC and NSC periods (Fig. S1 in the online supplemental material). In this context, the COI indices and their significant correlation appear to be consistently calculated. To assess the impact of restricting our investigations to the leading MCA covariability modes, thereby neglecting higher modes in the

determination of SC and NSC periods, an additional calculation has been performed between the index of anomalous precipitation in the Sahel and spatially averaged SSTA indices. The resultant SC and NSC periods are very similar to those introduced for the COI index (see Fig. S2).

The leading SSTA–anomalous rainfall covariability mode for eMED is depicted in Fig. 3. Positive (negative) SSTA in the Mediterranean Sea is related to increased (decreased) Sahel rainfall during SC (Fig. 3a). Note that station data indicate less coherence that is unlikely to be explicable by errors in the station data, but could be caused by smoothing effects and fewer or different stations used in the CRU analysis. Under a warmer eMED, the overall result is consistent with previous works that relate increased Sahelian rainfall to enhanced southward moisture advection by the mean flow from the Mediterranean across the Sahara Desert to the Sahel. This anomalous moisture advection appears to intensify the low-level convergence over the central-eastern Sahel (Rowell 2003; Fontaine et al. 2010, 2011; Gaetani et al. 2010). On the contrary, a Mediterranean Sea warming (cooling) is not associated with remarkable rainfall anomalies over the Sahel during the NSC period (Fig. 3b), which could be related to the effect of significant negative SSTA in the eastern North Atlantic Ocean.

For tATL (Fig. 4), the SC period is characterized by a positive (negative) AEM-like pattern related to negative (positive) rainfall anomalies in the Sahel, being opposite in the Guinea coast (Fig. 4a). A similar WAM response to the positive AEM has been found in previous studies, showing a southward shift of the ITCZ in response to a reduced land–ocean thermal contrast (e.g., Chiang et al. 2002) during a warm event of the AEM. This equatorward location of the ITCZ is associated with increased convective activity over the Guinea coast and dry conditions over the Sahel, resulting in a rainfall dipole over West Africa (Janowiak 1988; Fontaine and Janicot 1996; Janicot et al. 1998; Giannini et al. 2003; Joly and Voldoire 2010; Losada et al. 2010; Mohino et al. 2011b; Rodríguez-Fonseca et al. 2011). A different covariability pattern is observed during the NSC period (Fig. 4b). No dipolar rainfall structure appears in response to a tropical Atlantic warming; this result is consistent under the use of both gridded and station data. As for the eMED case, the CRU signal suggests a “simpler” response than the station network. In any case, the NSC results point to the possibility that other mechanisms may be undermining the influence of the tropical Atlantic in the last 20 years. Notably, a significant, opposite SSTA pattern appears in the eastern equatorial Pacific. The counteracting effect between the

FIG. 2. Expansion coefficients time series and COI indices. SST and rainfall expansion coefficients ($\mathbf{U}$ and $\mathbf{V}$, respectively) for the leading MCA mode between observed SSTA (HadISST) and observed standardized anomalous JAS rainfall calculated from CRU (blue bars and dashed green line, respectively) and stations (red bars and green solid line, respectively) in the Sahel (15°W – 15°E , 10° – 18°N). COI indices are calculated between $\mathbf{U}$ and $\mathbf{V}$ as indicated in the text for the HadISST–CRU analysis (blue line) and the HadISST–station analysis (red line). The expansion coefficients and COIs are presented for (a) eMED, (b) tATL, and (c) tPAC. Statistical significance for COI indices is set at 95% under the Monte Carlo method (1000 permutations) for the HadISST–CRU analysis (blue contoured boxes) and the HadISST–station analysis (red contoured circles). The correlation score between $\mathbf{U}$ and rainfall expansion coefficients ($\mathbf{V}$, not shown) is included in figure titles for the CRU (ruv_CRU) and stations (ruv_stations) analysis.

FIG. 3. Regression maps for the leading MCA mode between eMED SSTA and anomalous Sahel rainfall for (a) the SC period and (b) the NSC period. (left) Homogeneous SSTA (K std^{-1}) maps obtained by regression of SST expansion coefficient U onto global SSTa. Colored-shaded values correspond to HadISST-station analysis; contoured values denote results from HadISST-CRU analysis. (right) Heterogeneous anomalous rainfall ($\text{mm day}^{-1} \text{std}^{-1}$) maps obtained by regression of SST expansion coefficient U onto regional anomalous rainfall. Circles and squares denote significant and nonsignificant values, respectively, for HadISST-station analysis. Colored-shaded values correspond to HadISST-CRU analysis with the significant interval in stippling. The squared covariance fraction (SCF) is shown in figure titles for both (CRU and stations) analyses. Significance level is set at 95% under the Monte Carlo method (1000 permutations).

Atlantic and Pacific tropical ocean basins on Sahel rainfall has been addressed in observational and modeling studies (Rodríguez-Fonseca et al. 2011, 2015; Losada et al. 2012).

Regarding tPAC (Fig. 5), in both SC and NSC periods (Figs. 5a and 5b, respectively), ENSO-like warm (cold) events are related to decreased (increased) rainfall in the Sahel. However, the rainfall response is shown to be more robust in SC compared to NSC in terms of significant anomalies. Previous studies suggest how the upper-tropospheric heating over the tropical Pacific generates an atmospheric stationary Kelvin wave that propagates eastward within the equatorial Atlantic sector. Consequently, the upper-level zonal atmospheric circulation is altered, inducing anomalous subsidence over West Africa with the associated decrease in rainfall (Janicot et al. 2001; Rowell 2001; Giannini et al. 2005;

Joly and Voldoire 2009; Mohino et al. 2011a; Rodríguez-Fonseca et al. 2015). Accordingly, the different amplitudes of the SSTAs when comparing SC to NSC are related to the observed changes in the rainfall response, and hence are responsible for the nonstationary teleconnection pattern.

The findings for each oceanic predictor, specifically those for the SC periods, are mostly consistent with what is known about the leading SST-forced teleconnections driving interannual variability of rainfall in the Sahel. However, observational evidence is shown to conclude that these teleconnections are potentially unstable within the 90-yr study period. The results presented in the following subsections will shed light on these nonstationary links by analyzing the dynamical factors associated with SC and NSC covariability patterns. The years involved in the

FIG. 4. As in Fig. 3, but for the tATL.

calculations of seasonal composites are depicted in Fig. S3.

a. eMED–Sahel

Regarding eMED, a basinwide warming (cooling) is related to increased (decreased) rainfall in the Sahel during the SC period (see Fig. 3a). Positive rainfall anomalies occur in the central-eastern part of the study region (10°–18°N, 5°W–12°E). They appear to be related to enhanced northeasterly low-level moisture transport from a warmer eMED. This moisture flow emanating from the Mediterranean feeds convergence over the Sahel in association with the anomalous southwesterly monsoon flow (Fig. 6b). This mechanism is in agreement with previous studies dealing with the eMED–Sahel teleconnection (e.g., Rowell 2003; Gaetani et al. 2010). The anomalous northeasterly moisture flow from the eMED is part of the cyclonic circulation as a response to negative slp anomalies over Egypt and Libya (Fig. 6c). Enhanced convergence over the Sahel is shown in terms of low-level tropospheric moisture flux convergence (MFC) over the region (Fig. 6d), suggesting a northward migration of the ITCZ, as indicated by the anomalous southerly monsoon flow inland. The MFC is calculated

following the classical definition of mass convergence plus moisture advection (e.g., Banacos and Schultz 2005).

For the NSC period, the significant response of rainfall to the eMED warming vanishes (Fig. 7a). The anomalous low-level moisture transport from the Mediterranean Sea disappears, leading to lower, though nonsignificant, values of specific humidity over the Sahel (Fig. 7b). The anomalous moisture transport from the North Atlantic inland establishes as a result of the geostrophic low-level wind associated with an slp dipole of anomalously high pressures over the eastern subtropical North Atlantic Ocean and low pressures over North Africa (Fig. 7c). This anomalous northerly flow may be responsible for inhibiting the northward migration of the Atlantic ITCZ and associated southwesterly monsoon flow, keeping maximum precipitation to the south. As a result of this atmospheric configuration, the MFC weakens over the Sahel (Fig. 7d) compared to the pattern for the SC period.

b. tATL–Sahel

Significant changes are evident between NSC and SC periods concerning tATL. The SSTA in the SC period is

FIG. 5. As in Fig. 3, but for the tPAC.

characterized by a positive AEM-like pattern (see Fig. 4a). Because of its similarity with ENSO, the AEM is also known as the Atlantic Niño (e.g., Polo et al. 2008). The AEM peaks during boreal summer, when the anomalous conditions of the eastern equatorial upwelling can modify the zonal pressure gradient. Some authors point out a dipolar rainfall pattern in response to positive SSTAs (positive AEM phase) during some particular decades coinciding with the SC period identified in this study (e.g., Mohino et al. 2011b; Rodríguez-Fonseca et al. 2011; Losada et al. 2012). Positive SSTAs are also observed over the tATL during the NSC period (see Fig. 4b), in this case associated with a monopole response of increased rainfall in both the Gulf of Guinea and Sahel regions. Losada et al. (2012) conducted a set of sensitivity experiments showing this changing rainfall response to a positive AEM-like pattern. In addition, Diatta and Fink (2014) put forward a similar nonstationary relationship by means of statistical–observational analysis.

The AEM influence on the West African monsoon is characterized by an alteration of the monsoon-system components. The positive phase characterizing the SC period results in a well-defined dipole of anomalous precipitation (Fig. 8a), with negative anomalies over the

Sahel and positive ones over the Guinea coast (opposite for the negative AEM phase). The weakening of the meridional pressure gradient induced by the differential north–south land–ocean heating (Fig. 8b) reduces the monsoon flow and keeps the ITCZ equatorward (Chiang et al. 2002). The anomalous zonal wind profile (Fig. 8c) reveals a weakening of both the TEJ and the AEJ and reduced southwesterly monsoon flow related to the weakening of the low-level African westerly jet (AWJ; Pu and Cook 2012). This atmospheric configuration is concomitant with anomalously dry conditions in the Sahel (e.g., Nicholson 2013). This is corroborated by an anomalous sinking motion in the mid-to-upper troposphere over Sahelian latitudes (10° – 18° N), contrasted by enhanced upward motion over the Gulf of Guinea and the adjacent Guinea coast, suggesting a southward displacement of the ITCZ (Fig. 8d). The dry shallow convective cell associated with the SHL is weakened. Enhanced midtropospheric upward motions are in turn observed at Atlantic Ocean and West African longitudes (60° W– 15° E) in a zonal, equatorial cross section of meridional averages of ω (Fig. 8e).

The dynamical features described for the SC period are different during NSC (Fig. 9). The anomalous

FIG. 6. Dynamical mechanism associated with the eMED SSTA–anomalous Sahel rainfall teleconnection for the SC period in terms of high (H) minus low (L) composite maps for different atmospheric variables. The H and L events correspond to values of the SST expansion coefficient U above 1 std dev and below -1 std dev, respectively (see Fig. S3a, left panel). (a) Anomalous rainfall (mm day^{-1}) using CRU (color shaded) and station (circles and squares) databases. Significant values are denoted in stippling and circles. (b) Anomalous horizontal moisture transport ($\text{g kg}^{-1} \text{m s}^{-1}$) at 925 hPa (uvq925; arrows) and anomalous low-level specific humidity (g kg^{-1}) at 925 hPa (q925). Significant values are indicated in stippling and red arrows. (c) Mean slp (hPa) anomalies. (d) Zonally averaged (15°W – 15°E) latitudinal cross section of anomalous low-tropospheric (850–1000 hPa) MFC ($\text{g kg}^{-1} \text{day}^{-1}$). Positive (negative) MFC denotes convergence (divergence). Black vertical lines indicate the northern and southern Sahel limits (18° – 10°N). Significant values in (c) and (d) are denoted in stippling. Statistical significance is set at 90% under a t test.

rainfall dipole disappears (Fig. 9a), presumably because of the counteracting effect between opposite SSTAs in tATL and tPAC (see Fig. 5). The emergence of a colder tPAC induces an anomalous Walker circulation, weakening upward motions in the equatorial Atlantic Ocean including the Gulf of Guinea (Fig. 9e). Related to this, subsidence farther north over the Sahel weakens when compared to SC (Fig. 9d). Significant alterations in the zonal jets are not observed (Fig. 9c). It seems that SSTA forcing from the eastern tPAC counteracts the anomalous southward position of the ITCZ induced by a warmer tATL. The strong northerly anomalies at 10°N over the eastern tATL observed in SC vanish in NSC (Figs. 8b and 9b, respectively). In summary, it is concluded that the negative SSTAs in tPAC override the drying impact of a warmer tATL, thus leading to a relative increase of rainfall in the Sahel (Fig. 9a).

c. tPAC–Sahel

When analyzing the leading MCA mode using the tropical Pacific as a predictor, a positive ENSO-like pattern characterizes the leading SSTA mode in the SC period (see Fig. 5a), accompanying drought conditions in the Sahel (Fig. 10a). As described by Rowell (2001), deep convection is enhanced over a warmer eastern tPAC, transporting heat to the upper troposphere that triggers a stationary equatorial Kelvin wave to the east throughout the African–Atlantic sector and a Rossby wave to the west, resulting in anomalous subsidence over West Africa. Such a mechanism is observed in the present analysis in terms of anomalous upward motions over tPAC, describing an anomalous upper-level circulation that remotely connects the eastern tPAC and West Africa (Figs. 10b,c). This anomalous planetary-scale circulation pattern is obvious in the upper-level

FIG. 7. As in Fig. 6, but for the NSC period. The H and L events correspond to values of the SST expansion coefficient U above 1 std dev and below -1 std dev, respectively (see Fig. S3a, right panel).

divergent and rotational circulations (Figs. 10d and 10e, respectively). The direct response to the tPAC heating in a baroclinic atmosphere results in anomalous upper-level convergence (subsidence) over West Africa (Jin and Hoskins 1995; cf. Fig. 10e), inducing anomalous low-level divergence in the region (not shown). A Gill–Matsuno-type response (Matsuno 1966; Gill 1980) connects the large-scale anomalous circulation between the eastern tPAC and tATL (Fig. 10d), inducing large-scale downward (upward) motions in the West African troposphere under a warming (cooling) in the eastern tPAC. The Sahel rainfall response under these conditions agrees with previous studies on the subject (e.g., Joly and Voldoire 2009; Mohino et al. 2011a; Losada et al. 2012; Rodríguez-Fonseca et al. 2015).

As for the NSC period (Fig. 11), the SSTA forcing from a warmer tPAC resembles that previously described for the SC period. Nevertheless, the significant negative rainfall anomalies observed in SC are barely significant over the Sahel during NSC, being significantly positive in some regions to the south (Fig. 11a). These differences in the rainfall response may be interpreted as a similar ENSO response, yet stronger during SC compared to NSC (Figs. 5a and 5b, respectively). Some differences are apparent, such as in the vertical wind component (Fig. 11b), indicating a

weakening of the anomalous subsidence over the Sahel. When compared to SC, this attenuation of the atmospheric response can also be observed in the remaining variables treated in the analysis (Figs. 11c–e). Despite the similarity in the SSTA mode and associated dynamics characterizing the SC and NSC periods, the remarkable difference in the Sahel rainfall response puts forward the nonstationary behavior of the covariability pattern throughout the study period.

d. Skill scores of the statistical model

The different dynamical mechanisms associated with the SC and NSC leading modes of covariability should be related to changes in the SSTA-driven predictability of anomalous rainfall events in the Sahel. In this section, the skill scores of the S^4 CAST model to reproduce the observed anomalous rainfall are analyzed. Cross-validated hindcasts are calculated for SC and NSC periods following the expression in (2). The skill scores between rainfall hindcasts and observations are assessed by Pearson correlation. Note that in each time step within the SC and NSC periods, the regression coefficients are calculated as in (3), so that correlation skill scores are independently calculated for each period as in (4).

The skill scores of the model are depicted in Fig. 12 for SC and NSC (left and right columns, respectively).

FIG. 8. Dynamical mechanism associated with the tATL SSTA–anomalous Sahel rainfall teleconnection for the SC period in terms of high (H) minus low (L) composite maps for different atmospheric variables. The H and L events correspond to values of the SST expansion coefficient (U) above 1 std dev and below -1 std dev, respectively (see Fig. S3b, left panel). (a) Anomalous rainfall (mm day^{-1}) using CRU (color shaded) and station (circles and squares) databases. Significant values are denoted in stippling and circles. (b) Anomalous horizontal wind (m s^{-1}) at 925 hPa (uv925; arrows) and mean slp (hPa) anomalies. Significant values are indicated in stippling and red arrows. (c) Zonally averaged (15°W – 15°E) latitudinal cross section (100–1000 hPa) of anomalous zonal wind u (m s^{-1}). Positive (negative) values denote westerlies (easterlies). (d) Zonally averaged (15°W – 15°E) latitudinal cross section (100–1000 hPa) of anomalous vertical wind ω ($10^{-2} \text{ Pa s}^{-1}$). Positive (negative) values denote downward (upward) motions. Black vertical lines indicate the northern and southern Sahel limits (18° – 10°N) in (c) and (d). (e) Meridionally averaged (5°S – 5°N) longitudinal cross section (100–1000 hPa) of anomalous vertical wind ω ($10^{-2} \text{ Pa s}^{-1}$). Positive (negative) values denote downward (upward) motions. Significant values in (c)–(e) are indicated in stippling. Statistical significance is set at 90% under a t test.

Regardless of the lengths of the periods, only correlation coefficients exceeding the 95% significance level are plotted. The skill scores of the statistical model to hindcast the anomalous rainfall are qualitatively improved for the SC periods for each oceanic predictor.

Focusing on eMED (Fig. 12a), those changes in the SSTA patterns characterizing the leading modes of covariability in SC and NSC are related to enhanced skill scores for SC, when eMED SSTAs are considered in isolation. Regarding tATL (Fig. 12b), skill scores

FIG. 9. As in Fig. 8, but for the NSC period. The H and L events correspond to values of the SST expansion coefficient U above 1 std dev and below -1 std dev, respectively (see Fig. S3b, right panel).

significantly improve during SC, when the leading covariability mode is described by an isolated SSTA signal in the tropical Atlantic (SC) against the counteracting tATL–tPAC SSTA effect (NSC). Concerning tPAC (Fig. 12c), the skill scores decrease significantly compared to SC, a period during which a better skill could be related to the stronger ENSO-like signal associated with this period.

The improved skill scores are explained in terms of a stronger rainfall response under a certain SSTA pattern in the SC periods compared to the NSC counterpart. In all cases, the changing skill scores between SC and NSC periods are potentially related to the different underlying dynamics. Beyond these changes on interannual

time scales, the SSTA climatological background varies between SC and NSC periods, putting forward the possible influence of the multidecadal modulation of the interannual signal. A potential link between multidecadal SST variability and the nonstationary behavior of interannual teleconnections is analyzed in the next section.

4. Multidecadal SSTA background and ITCZ dynamics

The impacts of global SSTA on Sahelian rainfall variability extend to multidecadal time scales (e.g., Mohino et al. 2011c). The climatological SST background affects the mean ITCZ location (Marshall et al.

FIG. 10. Dynamical mechanism associated with the tPAC SSTA–anomalous Sahel rainfall teleconnection for the SC period in terms of high (H) minus low (L) composite maps for different atmospheric variables. The H and L events correspond to values of the SST expansion coefficient U above 1 std dev and below -1 std dev, respectively (see Fig. S3c, left panel). (a) Anomalous rainfall (mm day⁻¹) using CRU (colored shaded) and station (circles–squares) databases. Significant values are denoted in stippling and circles. (b) Zonally averaged (15°W–15°E) latitudinal cross section (100–1000 hPa) of anomalous vertical wind ω (m s⁻¹). Black vertical lines indicate the northern and southern Sahel limits (18°–10°N). (c) Meridionally averaged (5°S–5°N) longitudinal cross section (100–1000 hPa) of anomalous vertical wind ω (10⁻² Pa s⁻¹). (d) Anomalous streamfunction (10⁶ m² s⁻¹) at 200 hPa. (e) Anomalous velocity potential (10⁶ m² s⁻¹) at 200 hPa. Significant values in (b)–(e) are indicated in stippling. The statistical significance is set at 90% under a *t* test.

2014), with the interhemispheric SST gradients in the Atlantic Ocean basin and tropical Pacific SSTA playing an outstanding role (Chiang et al. 2000; Biasutti 2013). This section focuses on establishing a link

between the above-described significant SSTA–forced responses of anomalous Sahelian rainfall on inter-annual time scales and the multidecadal SSTA background state.

FIG. 11. As in Fig. 10, but for the NSC period. The H and L events correspond to values of the SST expansion coefficient U above 1 std dev and below -1 std dev, respectively (see Fig. S3c, right panel).

Averaging and standardizing the COI indices from Fig. 2 serve to calculate time series of indices varying on decadal time scales. An overview of these indices for the three oceanic predictors under study reveals three different multidecadal stages (Fig. 13a) depending on the skill scores of the statistical model. Overall, an improved, more effective rainfall response is observed according to the SC periods (see Fig. 12, left column): a first period (P1), from 1942 to 1964, which is related to enhanced predictability of anomalous Sahel rainfall associated with interannual variability in the eMED and

tATL regions; a second period (P2), from 1965 to 1987, during which the interannual predictability of anomalous rainfall improves if tropical oceanic predictors (tATL, tPAC) are considered; and a third period (P3), from 1988 to 2010, in which the interannual impact from eMED becomes crucial to enhance seasonal predictability of rainfall anomalies, even though the tPAC influence cannot be neglected. Note that each stage is characterized by the significant impact (SC period) of two oceanic predictors, while the impact of the third one is no longer significant.

FIG. 12. Skill scores between cross-validated hindcasts and observations of rainfall calculated in terms of Pearson correlation coefficients for each grid point in the Sahel (15°W–15°E, 10°–18°N). The analysis is based on the leading MCA modes as indicated in the text. Colored-shaded values correspond to HadISST–CRU analysis. Squares and circles correspond to HadISST–station analysis. Black stippled regions are statistically significant at 95% according to a Monte Carlo test (1000 permutations). Circles (squares) indicate significant (nonsignificant) values for HadISST–station analysis at 95% under a Monte Carlo test (1000 permutations). Cross validation is applied from the MCA between JAS Sahel rainfall and JAS SSTA for (a) eMED, (b) tATL, and (c) tPAC. Skill scores are shown for the (left) SC and (right) NSC periods.

FIG. 13. Multidecadal variability indices and global SSTA patterns. (a) Indices calculated as the 13-yr low-pass filter (solid lines) of the averaged-standardized COIs. SC periods are indicated by colored circles. Three different periods (P1, P2, and P3) are identified. SSTA maps calculated from HadISST as global JAS SST climatologies minus the JAS SST climatology along the whole period for (b) P1, (c) P2, and (d) P3.

Those three different stages are characterized by a specific SSTA background as shown in Figs. 13b–d, which is calculated for each period as the averaged SST minus the climatological SST for the whole study period. The spatial structures of these multidecadal SSTA patterns are characterized by interhemispheric gradients and the remote influence from the tropical Pacific. Differential north–south SST warming is known to impact Sahelian rainfall variability on multidecadal time scales (e.g., Munemoto and Tachibana 2012; Park et al. 2015), inducing meridional shifts of the ITCZ that potentially influence interannual teleconnections. Consequently, the spatial configuration of the SSTA pattern in each period (Figs. 13b–d) suggests a potential link between the changing climatological SSTA background and fluctuations in the ITCZ location.

Recent studies have demonstrated that the mean global position of the ITCZ tends to shift in the opposite direction of the atmospheric energy transport across the equator in response to differential warming between hemispheres (e.g., Kang et al. 2008, 2009; Frierson and Hwang 2012; Donohoe et al. 2013, 2014; Bischoff and Schneider 2014, 2016). The mean global ITCZ position

is explored by means of the latitude at which the vertically integrated moist static energy (MSE) flux [F in (6)] changes its sign (i.e., the energy flux equator; Broccoli et al. 2006), so that the meridional excursions of the ITCZ may be determined by the cross-equatorial energy flux F_0 where F diverges (e.g., Kraus 1977; Donohoe et al. 2013; Adam et al. 2016a,b):

$$F = \frac{1}{g} \int (c_p \bar{T} + g\bar{z} + L_v \bar{q}) \bar{v} dp. \quad (6)$$

In (6), c_p is the specific heat at constant pressure, T is the temperature, g is the gravity acceleration, z is the geopotential height, L_v is the vaporization latent heat, q is the specific humidity, and v is the meridional wind component; overbars denote zonal averages. The cross-equatorial energy flux F_0 is expected to be stronger the farther north the ITCZ is. Global profiles of F indicating meridional MSE fluxes are shown in Fig. 14a. Indeed, a zoomed-in view over subtropical latitudes in Fig. 14a, shown in Fig. 14b, reveals a global mean location of the ITCZ (zero of F ; F_0) around 15°N in P2, whereas it lies

FIG. 14. Atmospheric meridional MSE flux and latitudinal profiles of rainfall averaged in JAS. (a) Latitudinal profile of the vertically integrated MSE flux [petawatts (PW)] for the three periods (P1, P2, and P3) identified in Fig. 13. MSE is zonally and globally averaged between 50°S and 50°N. (b) Zoom in of the squared region marked in (a). (c) Latitudinal profile of climatological rainfall (mm day^{-1}) amounts between 15°W and 15°E for the three periods (P1, P2, and P3) and the EP. (d) Anomalous rainfall amounts between 15°W and 15°E calculated by subtracting the EP rainfall climatology to each one of the three periods (P1, P2, and P3). Each period is identified by a different color as indicated in the figures.

slightly farther north in P1 and P3. Despite the small amplitude ($\sim 0.6^\circ\text{N}$), this difference is found to be statistically significant at 99% under a t test. By contrast, the difference between P1 and P3 ($\sim 0.1^\circ\text{N}$) is not significant. Full information on how to analyze the ITCZ position from meridional energy fluxes as in (6) can be found in Bischoff and Schneider (2014) and Schneider et al. (2014).

The climatological position of the ITCZ is collocated with the maximum tropical rain belt, which is calculated for each period by means of the zonally averaged amount of rainfall. When compared to the entire period (EP), rainfall in the Sahel (Fig. 14c) is shown to increase in P1, decrease in P2, and lie in between both periods in P3. This behavior is further supported by anomalous rainfall in each period taking EP as the reference (Fig. 14d). Positive rainfall anomalies are found during P1, which are negative during P2. The last period (P3) is characterized by dry anomalies, although an apparent recovery trend is observed from the preceding period in terms of seasonal amounts of rainfall. Consequently, the link between the calculated climatological position of the global ITCZ and multidecadal variability of Sahel rainfall seems to be consistent. Moreover, the chronological

evolution P1–P2–P3 agrees with what is known from observations, namely, a wet period, followed by a drought stage in the Sahel, and the apparent trend toward increased seasonal rainfall amounts in the recent period (e.g., Le Barbè et al. 2002; Dai et al. 2004; Hagos and Cook 2008; Sanogo et al. 2015).

Qualitatively, the common warmer Northern Hemispheric SSTs between P1 and P3 (see Figs. 13b,d) are related to an enhanced interannual impact from eMED, as noted by the improved skill scores during the SC period. An intrinsic warming component to eMED is also observed in these periods, which in turn could strengthen the interannual impact from an anomalously warm eMED, whereas the impact from the negative phase (cooling) of the leading interannual covariability mode would be dampened. The weakening in the south–north positive SSTA gradient (during P3), or even the inversion thereof (during P2), goes along with a stronger influence from tPAC. In these cases, the multidecadal warming of the tropical Pacific would reinforce the drying of the Sahel under a positive ENSO-like pattern, the wetting impact associated with the negative ENSO phase being weakened. Concerning tATL, the Northern (Southern) Hemispheric SSTA gradient, characterizing

the underlying SSTA background during P1 (P2) is related to an enhanced interannual impact. The wetting impact from a negative AEM-like pattern would be enhanced (weakened) under a colder (warmer) equatorial Atlantic SSTA background (see [Figs. 13b and 13c](#), respectively). The effect would be the opposite for a positive AEM-like phase.

5. Discussion and conclusions

This study provides evidence that the interannual teleconnection patterns linking sea surface temperature anomalies (SSTAs) in the eastern Mediterranean (eMED), tropical Atlantic (tATL), and tropical Pacific (tPAC) with anomalous summer rainfall in the Sahel are potentially unstable over the observational record. Using observations, the leading covariability patterns between anomalous Sahelian rainfall and SSTA in those three oceanic regions were investigated for the 1921–2010 period. Decades with a significant or nonsignificant SST-forced rainfall response to eMED, tATL, and tPAC were identified from a series of simulations with the sea surface temperature–based statistical seasonal forecast (S^4 CAST) tool. Composites of dynamical and thermodynamical atmospheric fields were constructed for significant (SC) and nonsignificant (NSC) periods. Physical causes for changes in the leading covariability patterns were inferred from these maps. Furthermore, the study reveals a potential link between the nonstationary behavior of interannual teleconnections and large-scale multidecadal variability of the ocean–atmosphere coupled system.

Consistent with earlier studies ([Janicot et al. 1996](#); [Fontaine et al. 1998](#); [Mohino et al. 2011b](#); [Rodríguez-Fonseca et al. 2011, 2015](#); [Losada et al. 2012](#); [Diatta and Fink 2014](#)), the application of the S^4 CAST model corroborates the unstable nature of the eMED, tATL, and tPAC teleconnections with Sahelian rainfall in the 1921–2010 period. From 1942 to 1964 (P1), the tATL and eMED forcing was predominant; from 1965 to 1987 (P2), tPAC and tATL impacts were prevalent; and in the recent 1988–2010 period (P3), the tPAC and, more importantly, the eMED showed a strong link to Sahelian rainfall at year-to-year time scales. The latter has been shown in previous observational studies (e.g., [Fontaine et al. 2010](#)). The unstable teleconnections were linked to interdecadal changes in the leading modes of sea surface temperature (SST) variability related to summer Sahel rainfall. For the eMED, a cooling in the eastern North Atlantic is observed in NSC (i.e., periods when the teleconnection is not significant), being potentially responsible for inducing a more southerly position of the intertropical convergence zone (ITCZ), which is

associated with decreased rainfall in the Sahel despite a warmer Mediterranean Sea. By contrast, the absence of such cooling in SC periods allows the development of the “classical” teleconnection mechanism to unfold in terms of enhanced moisture flux across the Sahara, feeding the convection in the West African ITCZ (e.g., [Rowell 2003](#)). Concerning the tATL, the different mechanisms between SC and NSC periods are based on the previously known damping effect of the tropical Pacific observed in NSC. Regarding the tPAC, the composite analysis suggested that the nonstationary teleconnection is likely due to the varying amplitude of the El Niño–Southern Oscillation (ENSO) signal between SC and NSC.

The changing skill scores of cross-validated hindcasts performed with the S^4 CAST support the relevance of the results presented in this work in a context of rainfall predictability and associated impacts. The increase of skill found for certain periods in relation to different ocean predictors could be used to design socioeconomic adaptation strategies. Previous studies have shown seasonal rainfall totals to be markedly important in fields such as agronomy, energy, and hydrology (e.g., [Sultan et al. 2005, 2010](#); [Sultan and Gaetani 2016](#)). In this context, our results put forward seasonal rainfall anomalies for being more predictable under certain SSTA patterns. During SC, the rainfall response to SSTA over the study region is stronger, defining periods throughout which the predictability is enhanced. Dynamical causes were put forward to support this finding. For NSC periods, the absence of skill is related to weak—if any—teleconnections. The results corroborate the need to consider the nonstationary nature of the interannual teleconnections in order to improve the seasonal predictability of rainfall in the West African Sahel, as previously suggested in [Diatta and Fink \(2014\)](#). In addition, seasonal predictability should be analyzed together with the decadal component in order to obtain a better assessment of impacts over the Sahel.

On decadal time scales, our findings show that SSTA variability plays a major role in driving the interannual teleconnections. Changes in the background climatological SSTA could exert an influence not only on the decadal trends of Sahel rainfall, but also on defining the prevailing interannual SST-driven teleconnections within this region. We have found how a Northern (Southern) Hemispheric differential SST warming is related to a northward (southward) shift of the ITCZ, inducing wetter (drier) conditions in the Sahel whereas different interannual SST-forced teleconnections appear to be enhanced. This feature was evaluated for three decadal stages through an analysis of the meridional moist static energy (MSE) transport, suggesting a

potential link between the climatological position of the global ITCZ and multidecadal changes of the interannual teleconnection patterns. P1 is identified with the northernmost position of the ITCZ, which seems to foster the interannual impacts from the eMED and tATL. Under the SSTA pattern in P2, tropical oceans (tATL, tPAC) seem to interact together, prevailing over the extratropical influence of the eMED, whose impact gets weaker. Conversely, the underlying SST warming trend during P3 appears to promote the interannual teleconnection from the eMED, even though the tPAC impact is also significant. In addition, the SST warming trend appears to be responsible for increasing the Mediterranean Sea impact on the Sahel.

A global warming (GW)-like pattern has been linked to a weakened monsoon circulation, causing drought conditions (Dai 2013) as a response to the GW-induced stabilization of the tropical troposphere (Gaetani et al. 2017). Nevertheless, the moistening impact of anthropogenic Mediterranean Sea warming has been proposed as the leading driver of the recent trend of increasing precipitation in the Sahel (Park et al. 2016). Moreover, the Northern Hemispheric differential warming has been discussed in terms of its role in increasing precipitation, becoming a key factor in the projected Sahel rainfall (Munemoto and Tachibana 2012; Park et al. 2015).

In the current seasonal-to-decadal prediction framework, our results suggest that mixed strategies in which the decadal variability is evaluated as a possible modulator of the interannual variability should be taken into account. Nevertheless, models still fail in correctly reproducing tropical Atlantic and Mediterranean SST variability, with the decadal predictability still under debate. Much work is required in order to reduce biases for a correct dynamical assessment of the predictability. Thus, statistical analysis, longer observational records, and sensitivity experiments are required to better evaluate SST-forced teleconnections with the Sahel and associated modulators.

Acknowledgments. The research leading to these results received funding from the PREFACE-EU project (EU FP7/2007–2013) under Grant Agreement Number 603521 and the MINECO National Project PRE-4CAST (GCL2017-86415R). The fourth author acknowledges support by the MiKlip PROMISA (BMBF Grant 01 LP1520 D) project. All monthly station data are now included and are available from the newly created Karlsruhe African Surface Station Database (KASS-D) from the Institute of Meteorology and Climate Research of the Karlsruhe Institute of Technology, Germany, and can be obtained upon request to the fourth author. The authors also would like to acknowledge the remarks of

three anonymous reviewers who helped to greatly improve the manuscript.

REFERENCES

- Adam, O., T. Bischoff, and T. Schneider, 2016a: Seasonal and interannual variations of the energy flux equator and ITCZ. Part I: Zonally averaged ITCZ position. *J. Climate*, **29**, 3219–3230, <https://doi.org/10.1175/JCLI-D-15-0512.1>.
- , —, and —, 2016b: Seasonal and interannual variations of the energy flux equator and ITCZ. Part II: Zonally varying shifts of the ITCZ. *J. Climate*, **29**, 7281–7293, <https://doi.org/10.1175/JCLI-D-15-0710.1>.
- Banacos, P. C., and D. M. Schultz, 2005: The use of moisture flux convergence in forecasting convective initiation: Historical and operational perspectives. *Wea. Forecasting*, **20**, 351–366, <https://doi.org/10.1175/WAF858.1>.
- Biasutti, M., 2013: Forced Sahel rainfall trends in the CMIP5 archive. *J. Geophys. Res. Atmos.*, **118**, 1613–1623, <https://doi.org/10.1002/jgrd.50206>.
- Bijlsma, S. J., L. M. Hafkenscheid, and P. Lynch, 1986: Computation of the streamfunction and velocity potential and reconstruction of the wind field. *Mon. Wea. Rev.*, **114**, 1547–1551, [https://doi.org/10.1175/1520-0493\(1986\)114<1547:COTSAV>2.0.CO;2](https://doi.org/10.1175/1520-0493(1986)114<1547:COTSAV>2.0.CO;2).
- Bischoff, T., and T. Schneider, 2014: Energetic constraints on the position of the intertropical convergence zone. *J. Climate*, **27**, 4937–4951, <https://doi.org/10.1175/JCLI-D-13-00650.1>.
- , and —, 2016: The equatorial energy balance, ITCZ position, and double ITCZ bifurcations. *J. Climate*, **29**, 2997–3013, <https://doi.org/10.1175/JCLI-D-15-0328.1>.
- Bretherton, C. S., C. Smith, and J. M. Wallace, 1992: An intercomparison of methods for finding coupled patterns in climate data. *J. Climate*, **5**, 541–560, [https://doi.org/10.1175/1520-0442\(1992\)005<0541:AIOMFF>2.0.CO;2](https://doi.org/10.1175/1520-0442(1992)005<0541:AIOMFF>2.0.CO;2).
- Broccoli, A. J., K. A. Dahl, and R. J. Stouffer, 2006: Response of the ITCZ to Northern Hemisphere cooling. *Geophys. Res. Lett.*, **33**, L01702, <https://doi.org/10.1029/2005GL024546>.
- Cai, W., and P. H. Whetton, 2001: Modes of SST variability and the fluctuation of global mean temperature. *Climate Dyn.*, **17**, 889–901, <https://doi.org/10.1007/s003820100152>.
- Cherry, S., 1996: Singular value decomposition analysis and canonical correlation analysis. *J. Climate*, **9**, 2003–2009, [https://doi.org/10.1175/1520-0442\(1996\)009<2003:SVDAAC>2.0.CO;2](https://doi.org/10.1175/1520-0442(1996)009<2003:SVDAAC>2.0.CO;2).
- Chiang, J. C. H., Y. Kushnir, and S. E. Zebiak, 2000: Interdecadal changes in eastern Pacific ITCZ variability and its influence on the Atlantic ITCZ. *Geophys. Res. Lett.*, **27**, 3687–3690, <https://doi.org/10.1029/1999GL011268>.
- , —, and A. Giannini, 2002: Deconstructing Atlantic intertropical convergence zone variability: Influence of the local cross-equatorial sea surface temperature gradient and remote forcing from the eastern equatorial Pacific. *J. Geophys. Res.*, **107**, 4004, <https://doi.org/10.1029/2000JD000307>.
- Clarke, A. J., 2014: El Niño physics and El Niño predictability. *Annu. Rev. Mar. Sci.*, **6**, 79–99, <https://doi.org/10.1146/annurev-marine-010213-135026>.
- Dai, A., 2013: Increasing drought under global warming in observations and models. *Nat. Climate Change*, **3**, 52–58, <https://doi.org/10.1038/nclimate1633>.
- , P. J. Lamb, K. E. Trenberth, M. Hulme, P. D. Jones, and P. Xie, 2004: The recent Sahel drought is real. *Int. J. Climatol.*, **24**, 1323–1331, <https://doi.org/10.1002/joc.1083>.

- Dayan, H., J. Vialard, T. Izumo, and M. Lengaigne, 2014: Does sea surface temperature outside the tropical Pacific contribute to enhanced ENSO predictability? *Climate Dyn.*, **43**, 1311–1325, <https://doi.org/10.1007/s00382-013-1946-y>.
- Diallo, I., M. B. Sylla, M. Camara, and A. T. Gaye, 2013: Interannual variability of rainfall over the Sahel based on multiple regional climate models simulations. *Theor. Appl. Climatol.*, **113**, 351–362, <https://doi.org/10.1007/s00704-012-0791-y>.
- Diatta, S., and A. H. Fink, 2014: Statistical relationship between remote climate indices and West African monsoon variability. *Int. J. Climatol.*, **34**, 3348–3367, <https://doi.org/10.1002/joc.3912>.
- Donohoe, A., J. Marshall, D. Ferreira, and D. McGee, 2013: The relationship between ITCZ location and cross-equatorial atmospheric heat transport: From the seasonal cycle to the Last Glacial Maximum. *J. Climate*, **26**, 3597–3618, <https://doi.org/10.1175/JCLI-D-12-00467.1>.
- , —, —, K. Armour, and D. McGee, 2014: The interannual variability of tropical precipitation and interhemispheric energy transport. *J. Climate*, **27**, 3377–3392, <https://doi.org/10.1175/JCLI-D-13-00499.1>.
- Ebisuzaki, W., 1997: A method to estimate the statistical significance of a correlation when the data are serially correlated. *J. Climate*, **10**, 2147–2153, [https://doi.org/10.1175/1520-0442\(1997\)010<2147:AMTETS>2.0.CO;2](https://doi.org/10.1175/1520-0442(1997)010<2147:AMTETS>2.0.CO;2).
- Fink, A. H., and Coauthors, 2017: Mean climate and seasonal cycle. *Meteorology of Tropical West Africa: The Forecasters' Handbook*, D. J. Parker and M. Diop-Kane, Eds., John Wiley and Sons, 1–39.
- Folland, C. K., T. N. Palmer, and D. E. Parker, 1986: Sahel rainfall and worldwide sea temperatures, 1901–85. *Nature*, **320**, 602–607, <https://doi.org/10.1038/320602a0>.
- Fontaine, B., and S. Janicot, 1996: Sea surface temperature fields associated with West African rainfall anomaly types. *J. Climate*, **9**, 2935–2940, [https://doi.org/10.1175/1520-0442\(1996\)009<2935:SSTFAW>2.0.CO;2](https://doi.org/10.1175/1520-0442(1996)009<2935:SSTFAW>2.0.CO;2).
- , S. Trzaska, and S. Janicot, 1998: Evolution of the relationship between near global and Atlantic SST modes and the rainy season in West Africa: Statistical analyses and sensitivity experiments. *Climate Dyn.*, **14**, 353–368, <https://doi.org/10.1007/s003820050228>.
- , and Coauthors, 2010: Impacts of warm and cold situations in the Mediterranean basins on the West African monsoon: Observed connection patterns (1979–2006) and climate simulations. *Climate Dyn.*, **35**, 95–114, <https://doi.org/10.1007/s00382-009-0599-3>.
- , P.-A. Monerie, M. Gaetani, and P. Roucou, 2011: Climate adjustments over the African-Indian monsoon regions accompanying Mediterranean Sea thermal variability. *J. Geophys. Res.*, **116**, D23122, <https://doi.org/10.1029/2011JD016273>.
- Frierson, D. M. W., and Y.-T. Hwang, 2012: Extratropical influence on ITCZ shifts in slab ocean simulations of global warming. *J. Climate*, **25**, 720–733, <https://doi.org/10.1175/JCLI-D-11-00116.1>.
- Gaetani, M., B. Fontaine, P. Roucou, and M. Baldi, 2010: Influence of the Mediterranean Sea on the West African monsoon: Intraseasonal variability in numerical simulations. *J. Geophys. Res.*, **115**, D24115, <https://doi.org/10.1029/2010JD014436>.
- , C. Flamant, S. Bastin, S. Janicot, C. Lavaysse, F. Hourdin, P. Braconnot, and S. Bony, 2017: West African monsoon dynamics and precipitation: The competition between global SST warming and CO₂ increase in CMIP5 idealized simulations. *Climate Dyn.*, **48**, 1353–1373, <https://doi.org/10.1007/s00382-016-3146-z>.
- Giannini, A., R. Saravanan, and P. Chang, 2003: Oceanic forcing of Sahel rainfall on interannual to interdecadal time scales. *Science*, **302**, 1027–1030, <https://doi.org/10.1126/science.1089357>.
- , —, and —, 2005: Dynamics of the boreal summer African monsoon in the NSIPP1 atmospheric model. *Climate Dyn.*, **25**, 517–535, <https://doi.org/10.1007/s00382-005-0056-x>.
- Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**, 447–462, <https://doi.org/10.1002/qj.49710644905>.
- Grist, J. P., and S. E. Nicholson, 2001: A study of the dynamic factors influencing the rainfall variability in the West African Sahel. *J. Climate*, **14**, 1337–1359, [https://doi.org/10.1175/1520-0442\(2001\)014<1337:ASOTDF>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<1337:ASOTDF>2.0.CO;2).
- Hagos, S. M., and K. H. Cook, 2008: Ocean warming and late-twentieth-century Sahel drought and recovery. *J. Climate*, **21**, 3797–3814, <https://doi.org/10.1175/2008JCLI2055.1>.
- Harris, I., P. D. Jones, T. J. Osborn, and D. H. Lister, 2014: Updated high-resolution grids of monthly climatic observations—The CRU TS3.10 dataset. *Int. J. Climatol.*, **34**, 623–642, <https://doi.org/10.1002/joc.3711>.
- Janicot, S., V. Moron, and B. Fontaine, 1996: Sahel droughts and ENSO dynamics. *Geophys. Res. Lett.*, **23**, 515–518, <https://doi.org/10.1029/96GL00246>.
- , A. Harzallah, B. Fontaine, and V. Moron, 1998: West African monsoon dynamics and eastern equatorial Atlantic and Pacific SST anomalies (1970–88). *J. Climate*, **11**, 1874–1882, <https://doi.org/10.1175/1520-0442-11.8.1874>.
- , S. Trzaska, and I. Poccard, 2001: Summer Sahel-ENSO teleconnection and decadal time scale SST variations. *Climate Dyn.*, **18**, 303–320, <https://doi.org/10.1007/s003820100172>.
- Janowiak, J. E., 1988: An investigation of interannual rainfall variability in Africa. *J. Climate*, **1**, 240–255, [https://doi.org/10.1175/1520-0442\(1988\)001<0240:AIOIRV>2.0.CO;2](https://doi.org/10.1175/1520-0442(1988)001<0240:AIOIRV>2.0.CO;2).
- Jin, F., and B. J. Hoskins, 1995: The direct response to tropical heating in a baroclinic atmosphere. *J. Atmos. Sci.*, **52**, 307–319, [https://doi.org/10.1175/1520-0469\(1995\)052<0307:TDRTH>2.0.CO;2](https://doi.org/10.1175/1520-0469(1995)052<0307:TDRTH>2.0.CO;2).
- Joly, M., and A. Voldoire, 2009: Influence of ENSO on the West African monsoon: Temporal aspects and atmospheric processes. *J. Climate*, **22**, 3193–3210, <https://doi.org/10.1175/2008JCLI2450.1>.
- , and —, 2010: Role of the Gulf of Guinea in the inter-annual variability of the West African monsoon: What do we learn from CMIP3 coupled simulations? *Int. J. Climatol.*, **30**, 1843–1856, <https://doi.org/10.1002/joc.2026>.
- Jung, T., L. Ferranti, and A. M. Tompkins, 2006: Response to the summer of 2003 Mediterranean SST anomalies over Europe and Africa. *J. Climate*, **19**, 5439–5454, <https://doi.org/10.1175/JCLI3916.1>.
- Kang, S. M., I. M. Held, D. M. W. Frierson, and M. Zhao, 2008: The response of the ITCZ to extratropical thermal forcing: Idealized slab-ocean experiments with a GCM. *J. Climate*, **21**, 3521–3532, <https://doi.org/10.1175/2007JCLI2146.1>.
- , D. M. W. Frierson, and I. M. Held, 2009: The tropical response to extratropical thermal forcing in an idealized GCM: The importance of radiative feedbacks and convective parameterization. *J. Atmos. Sci.*, **66**, 2812–2827, <https://doi.org/10.1175/2009JAS2924.1>.
- Kraus, E. B., 1977: Subtropical droughts and cross-equatorial energy transports. *Mon. Wea. Rev.*, **105**, 1009–1018, [https://doi.org/10.1175/1520-0493\(1977\)105<1009:SDACEE>2.0.CO;2](https://doi.org/10.1175/1520-0493(1977)105<1009:SDACEE>2.0.CO;2).

- Lafore, J.-P., and Coauthors, 2011: Progress in understanding of weather systems in West Africa. *Atmos. Sci. Lett.*, **12**, 7–12, <https://doi.org/10.1002/asl.335>.
- Lau, N.-C., and M. J. Nath, 1994: A modeling study of the relative roles of tropical and extratropical SST anomalies in the variability of the global atmosphere–ocean system. *J. Climate*, **7**, 1184–1207, [https://doi.org/10.1175/1520-0442\(1994\)007<1184:AMSOTR>2.0.CO;2](https://doi.org/10.1175/1520-0442(1994)007<1184:AMSOTR>2.0.CO;2).
- Lavaysse, C., C. Flamant, S. Janicot, D. J. Parker, J.-P. Lafore, B. Sultan, and J. Pelon, 2009: Seasonal evolution of the West African heat low: A climatological perspective. *Climate Dyn.*, **33**, 313–330, <https://doi.org/10.1007/s00382-009-0553-4>.
- , —, and —, 2010: Regional-scale convection patterns during strong and weak phases of the Saharan heat low. *Atmos. Sci. Lett.*, **11**, 255–264, <https://doi.org/10.1002/asl.284>.
- Le Barbé, L., T. Lebel, and D. Tapsoba, 2002: Rainfall variability in West Africa during the years 1950–90. *J. Climate*, **15**, 187–202, [https://doi.org/10.1175/1520-0442\(2002\)015<0187:RVIWAD>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<0187:RVIWAD>2.0.CO;2).
- Losada, T., B. Rodríguez-Fonseca, S. Janicot, S. Gervois, F. Chauvin, and P. Ruti, 2010: A multi-model approach to the Atlantic equatorial mode: Impact on the West African monsoon. *Climate Dyn.*, **35**, 29–43, <https://doi.org/10.1007/s00382-009-0625-5>.
- , —, E. Mohino, J. Bader, S. Janicot, and C. R. Mechoso, 2012: Tropical SST and Sahel rainfall: A non-stationary relationship. *Geophys. Res. Lett.*, **39**, L12705, <https://doi.org/10.1029/2012GL052423>.
- Marshall, J., A. Donohoe, D. Ferreira, and D. McGee, 2014: The ocean's role in setting the mean position of the inter-tropical convergence zone. *Climate Dyn.*, **42**, 1967–1979, <https://doi.org/10.1007/s00382-013-1767-z>.
- Matsuno, T., 1966: Quasi-geostrophic motion in the equatorial area. *J. Meteor. Soc. Japan*, **44**, 25–43, https://doi.org/10.2151/jmsj1965.44.1_25.
- Mohino, E., B. Rodríguez-Fonseca, C. R. Mechoso, S. Gervois, P. Ruti, and F. Chauvin, 2011a: Impacts of the tropical Pacific/Indian Oceans on the seasonal cycle of the West African monsoon. *J. Climate*, **24**, 3878–3891, <https://doi.org/10.1175/2011JCLI3988.1>.
- , —, T. Losada, S. Gervois, S. Janicot, J. Bader, P. Ruti, and F. Chauvin, 2011b: Changes in the interannual SST-forced signals on West African rainfall. AGCM intercomparison. *Climate Dyn.*, **37**, 1707–1725, <https://doi.org/10.1007/s00382-011-1093-2>.
- , S. Janicot, and J. Bader, 2011c: Sahel rainfall and decadal to multi-decadal sea surface temperature variability. *Climate Dyn.*, **37**, 419–440, <https://doi.org/10.1007/s00382-010-0867-2>.
- Moron, V., R. Vautard, and M. Ghil, 1998: Trends, interdecadal and interannual oscillation in global sea-surface temperatures. *Climate Dyn.*, **14**, 545–569, <https://doi.org/10.1007/s003820050241>.
- Munemoto, M., and Y. Tachibana, 2012: The recent trend of increasing precipitation in Sahel and the associated inter-hemispheric dipole of global SST. *Int. J. Climatol.*, **32**, 1346–1353, <https://doi.org/10.1002/joc.2356>.
- Nicholson, S., 2000: Land surface processes and Sahel climate. *Rev. Geophys.*, **38**, 117–139, <https://doi.org/10.1029/1999RG900014>.
- , 2009a: A revised picture of the structure of the “monsoon” and land ITCZ over West Africa. *Climate Dyn.*, **32**, 1155–1171, <https://doi.org/10.1007/s00382-008-0514-3>.
- , 2009b: On the factors modulating the intensity of the tropical rainbelt over West Africa. *Int. J. Climatol.*, **29**, 673–689, <https://doi.org/10.1002/joc.1702>.
- , 2013: The West African Sahel: A review of recent studies on the rainfall regime and its interannual variability. *ISRN Meteor.*, 453521, <https://doi.org/10.1155/2013/453521>.
- , and P. J. Webster, 2007: A physical basis for the interannual variability of rainfall in the Sahel. *Quart. J. Roy. Meteor. Soc.*, **133**, 2065–2084, <https://doi.org/10.1002/qj.104>.
- , A. I. Barcion, and M. Challa, 2008: An analysis of West African dynamics using a linearized GCM. *J. Atmos. Sci.*, **65**, 1182–1203, <https://doi.org/10.1175/2007JAS2194.1>.
- Okumura, Y., and S.-P. Xie, 2004: Interaction of the Atlantic equatorial cold tongue and the African monsoon. *J. Climate*, **17**, 3589–3602, [https://doi.org/10.1175/1520-0442\(2004\)017<3589:IOTAE>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<3589:IOTAE>2.0.CO;2).
- Park, J.-Y., J. Bader, and D. Matei, 2015: Northern-hemispheric differential warming is the key to understanding the discrepancies in the projected Sahel rainfall. *Nat. Commun.*, **6**, 5985, <https://doi.org/10.1038/ncomms6985>.
- , —, and —, 2016: Anthropogenic Mediterranean warming essential driver for present and future Sahel rainfall. *Nat. Climate Change*, **6**, 941–945, <https://doi.org/10.1038/nclimate3065>.
- Poli, P., and Coauthors, 2016: ERA-20C: An atmospheric re-analysis of the twentieth century. *J. Climate*, **29**, 4083–4097, <https://doi.org/10.1175/JCLI-D-15-0556.1>.
- Polo, I., B. Rodríguez-Fonseca, T. Losada, and J. García-Serrano, 2008: Tropical Atlantic modes (1979–2002). Part I: Time-evolving SST modes related to West African rainfall. *J. Climate*, **21**, 6457–6475, <https://doi.org/10.1175/2008JCLI2607.1>.
- Pu, B., and K. H. Cook, 2012: Role of the West African westerly jet in Sahel rainfall variations. *J. Climate*, **25**, 2880–2896, <https://doi.org/10.1175/JCLI-D-11-00394.1>.
- Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**, 4407, <https://doi.org/10.1029/2002JD002670>.
- Redelsperger, J.-L., C. D. Thorncroft, A. Diedhiou, T. Lebel, D. J. Parker, and J. Polcher, 2006: African Monsoon Multi-disciplinary Analysis: An international research project and field campaign. *Bull. Amer. Meteor. Soc.*, **87**, 1739–1746, <https://doi.org/10.1175/BAMS-87-12-1739>.
- Rodríguez-Fonseca, B., I. Polo, J. García-Serrano, L. Losada, E. Mohino, C. R. Mechoso, and F. Kucharski, 2009: Are Atlantic Niños enhancing Pacific ENSO events in recent decades? *Geophys. Res. Lett.*, **36**, L20705, <https://doi.org/10.1029/2009GL040048>.
- , and Coauthors, 2011: Interannual and decadal SST-forced responses of the West African monsoon. *Atmos. Sci. Lett.*, **12**, 67–74, <https://doi.org/10.1002/asl.308>.
- , and Coauthors, 2015: Variability and predictability of West African droughts: A review on the role of sea surface temperature anomalies. *J. Climate*, **28**, 4034–4060, <https://doi.org/10.1175/JCLI-D-14-00130.1>.
- , and Coauthors, 2016: A review of ENSO influence on the North Atlantic. A non-stationary signal. *Atmosphere*, **7**, 87, <https://doi.org/10.3390/atmos7070087>.
- Rowell, D. P., 2001: Teleconnections between the tropical Pacific and the Sahel. *Quart. J. Roy. Meteor. Soc.*, **127**, 1683–1706, <https://doi.org/10.1002/qj.49712757512>.
- , 2003: The impact of Mediterranean SSTs on the Sahelian rainfall season. *J. Climate*, **16**, 849–862, [https://doi.org/10.1175/1520-0442\(2003\)016<0849:TIOMSO>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<0849:TIOMSO>2.0.CO;2).
- , 2013: Simulating SST teleconnections to Africa: What is the state of the art? *J. Climate*, **26**, 5397–5418, <https://doi.org/10.1175/JCLI-D-12-00761.1>.

- , C. K. Folland, K. Maskell, and M. N. Ward, 1995: Variability of summer rainfall over tropical North Africa (1906–92): Observations and modelling. *Quart. J. Roy. Meteor. Soc.*, **121**, 669–704, <https://doi.org/10.1002/qj.49712152311>.
- Sanogo, S., A. H. Fink, J. A. Omotosho, A. Ba, R. Redl, and V. Ermert, 2015: Spatio-temporal characteristics of the recent rainfall recovery in West Africa. *Int. J. Climatol.*, **35**, 4589–4605, <https://doi.org/10.1002/joc.4309>.
- Schneider, T., T. Bischoff, and G. H. Haug, 2014: Migrations and dynamics of the intertropical convergence zone. *Nature*, **513**, 45–53, <https://doi.org/10.1038/nature13636>.
- Stockdale, T. N., and Coauthors, 2010: Understanding and predicting seasonal-to-interannual climate variability—The producer perspective. *Procedia Environ. Sci.*, **1**, 55–80, <https://doi.org/10.1016/j.proenv.2010.09.006>.
- Suárez-Moreno, R., and B. Rodríguez-Fonseca, 2015: S⁴CAST v2.0: Sea surface temperature based statistical seasonal forecast model. *Geosci. Model Dev.*, **8**, 3639–3658, <https://doi.org/10.5194/gmd-8-3639-2015>.
- Sultan, B., and S. Janicot, 2000: Abrupt shift of the ITCZ over West Africa and intra-seasonal variability. *Geophys. Res. Lett.*, **27**, 3353–3356, <https://doi.org/10.1029/1999GL011285>.
- , and M. Gaetani, 2016: Agriculture in West Africa in the twenty-first century: Climate change and impacts scenarios, and potential for adaptation. *Front. Plant Sci.*, **7**, 1262, <https://doi.org/10.3389/fpls.2016.01262>.
- , C. Baron, M. Dingkuhn, B. Sarr, and S. Janicot, 2005: Agricultural impacts of large-scale variability of the West African monsoon. *Agric. For. Meteorol.*, **128**, 93–110, <https://doi.org/10.1016/j.agrformet.2004.08.005>.
- , B. Barbier, J. Fortilus, S. M. Mbaye, and G. Leclerc, 2010: Estimating the potential economic value of seasonal forecasts in West Africa: A long-term ex-ante assessment in Senegal. *Wea. Climate Soc.*, **2**, 69–87, <https://doi.org/10.1175/2009WCAS1022.1>.
- von Storch, H., and F. W. Zwiers, 2002: *Statistical Analysis in Climate Research*. Cambridge University Press, 496 pp.
- Wallace, J. M., C. Smith, and C. S. Bretherton, 1992: Singular value decomposition of wintertime sea surface temperature and 500-mb height anomalies. *J. Climate*, **5**, 561–576, [https://doi.org/10.1175/1520-0442\(1992\)005<0561:SVDOWS>2.0.CO;2](https://doi.org/10.1175/1520-0442(1992)005<0561:SVDOWS>2.0.CO;2).
- Wang, C., C. Deser, J.-Y. Yu, P. DiNezio, and A. Clement, 2012: El Niño–Southern Oscillation (ENSO): A review. *Coral Reefs of the Eastern Tropical Pacific: Persistence and Loss in a Dynamic Environment*, P. W. Glynn, D. Manzello, and I. C. Enochs, Eds., Springer Science, 3–19.
- Widmann, M., 2005: One-dimensional CCA and SVD, and their relationship to regression maps. *J. Climate*, **18**, 2785–2792, <https://doi.org/10.1175/JCLI3424.1>.