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**PAST, PRESENT, AND FUTURE IMPACTS OF OCEANIC
INTERNAL MODES OF VARIABILITY ON THE RAINFALL
OVER THE WEST AFRICAN GUINEA COAST**

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Acronyms

AEJ	African Easterly Jet
AEM	Atlantic Equatorial Mode
AEW	African Easterly Waves
AGCM	Atmospheric General Circulation Model
AMJ	April-May-June
AMM	Atlantic Meridional Mode
AMO	Atlantic Multi-decadal Oscillation
AMV	Atlantic Multi-decadal Variability
AOGCM	Atmosphere-Ocean general circulation model
AR6	Sixth Assessment Report
ATL3	Atlantic Niño 3 region
CDD	Consecutive dry days
CHIRPS	Climate Hazards Group Infrared Precipitation with Station Data
CINDO	Center of Indian Ocean
CMIP	Coupled Models Intercomparison Project
CMIP5	Coupled Models Intercomparison Project, phase 5
CMIP6	Coupled Models Intercomparison Project, phase 6
CORDEX	Coordinated Regional Climate Downscaling Experiment
CRU TS	Climate Research Unit Time Series
CWD	Consecutive wet days
DIV200/850	Wind divergence at 200 hPa minus wind divergence at 850 hPa
DJF	December-January-February
DMI	Dipole Mode (Indian Ocean)
EAB	Equatorial Atlantic box
ECMWF	European Centre for Medium-Range Weather Forecasts
EDOF	Effective degree of freedom
EMI	El-Niño Modoki
EnsMean	Ensemble Mean

EnsMedian	Ensemble Median
ENSO	El Niño-Southern Oscillation
EOF	Empirical Orthogonal Function
ERA5	fifth generation ECMWF atmospheric reanalysis
ESGF	Earth System Grid Federation
ETCCDI	Expert Team on Climate Change Detection and Indices
FRQW	Frequency of wet days
GC	Guinea Coast
GCB	Guinea Coast box
GCM	General Circulation Model
GCR	Guinea Coast rainfall
GMMIP	Global Monsoons Model Intercomparison
HadISST	Hadley Centre Global Sea Ice and Sea Surface Temperature
HGCR	high frequency Guinea Coast rainfall index
ICDC	Integrated Climate Data Center
IEMI	Improved El-Niño Modoki
IOD	Indian Ocean Dipole
IPCC	Intergovernmental Panel on Climate Change
IPO	Inter-decadal Pacific Oscillation
IPWP	Indo-Pacific Warm Pool
IQR	Interquartile range
ITCZ	Intertropical convergence zone
ITD	Intertropical discontinuity
ITF	Intertropical front
JAS	July-August-September
JJAS	June-July-August-September
MAM	March-April-May
MCS	mesoscale convective system
MED	Mediterranean Sea
MEDE	Eastern Mediterranean Sea
MEDW	Western Mediterranean Sea
NAO	North Atlantic Oscillation
NAT	North Atlantic Tropical SST
NATL	North Atlantic SST
NEP	North East Pole
NINO3	Niño3 index
NINO34	Niño3.4 index
NINO4	Niño4 index
NPAC	North Pacific SST index
ON	October-November
ORAS5	Ocean ReAnalysis System 5

PRCPTOT	Total wet-day precipitation
R10mm	Heavy precipitation days
R20mm	Very heavy precipitation days
R95p	Very wet days
R95pf	Precipitation percent due to very wet days
R99p	Extremely wet days
RCM	Regional Climate Model
RCP	Representative Concentration Pathway
RCP4.5	Representative Concentration Pathway 4.5
RCP8.5	Representative Concentration Pathway 8.5
RX1day	Maximum 1 day precipitation
RX5day	Maximum 5 days precipitation
SAM	Southern Annular Mode
SAO	South Atlantic Ocean
SAOD	South Atlantic Ocean Dipole
SAT	South Atlantic Tropical SST
SDA	Sign-Dependent Average
SDII	Simple daily intensity index
SDOF	Standard degree of freedom
SETIO	Southeastern Tropical Indian Ocean SST
SH	Sahel
SHL	Saharan Heat Low
SMC	Shallow Meridonal Circulation
SSH	Sea surface height
SSP5-8.5	Shared Socio-economic pathway 5-8.5
SST	Sea Surface Temperature
SSTA	Sea Surface Temperature Anomaly
STJ	Subtropical Jet
SWP	South West Pole
TAB1	Tropical Atlantic box 1
TAB2	Tropical Atlantic box 2
TASI	Tropical Atlantic SST meridional gradient
TEJ	Tropical Easterly Jet
TNA	Tropical Northern Atlantic
TSA	Tropical Southern Atlantic
VEF	variance explained fraction
WA	West Africa
WAB	West Africa box
WAHL	West African heat low
WAM	West Africa Monsoon
WAWJ	West African Westerly Jet

x

Acronyms

WTIO

West Tropical Indian Ocean SST

1

Introduction

For someone who loves physical laws applied to the climate system, the study of the West African hydroclimate could rapidly become addictive. The region is affected by so many processes that interact on various spatial and temporal scales. The topography complexity, the proximity to the Atlantic Ocean, the presence of the Saharan heat low, the influence of different jets from the low levels to the high levels as well as the waves generated in their core lead to a very rich dynamics associated with a wide range of phenomena, from extraordinary weather conditions to changes in the West African monsoon system. Intensive research fields have developed after the pronounced drought that affected the Sahelian region in the 70s and 80s. Part of the variability of the Sahelian rainfall has been attributed to different phases of some oceanic internal modes of variability. However, less attention has been paid to the rainfall variability in the Guinea Coast, a region located in the south of the Sahel. In order to fill this gap, this thesis aims to understand the past, present and future role of the leading oceanic internal modes of variability on the year-to-year rainfall variability over the Guinea Coast.

1.1 Key features of the West African Monsoon system

1.1.1 The West African region, relief, bioclimatic regions, population and economy

The geographic region of West Africa (WA) is made up of 17 countries: Benin, Burkina Faso, Cabo Verde, Côte d'Ivoire, Gambia, Ghana, Guinea, Guinea-Bissau, Liberia, Mali, Mauritania, Niger, Nigeria, Saint Helena, Senegal, Sierra Leone and Togo [United Nations, Department of Economic and Social Affairs, Population Di-

vision, 2022]. From a climatic point of view, this region can be extended to 30°N , to highlight its key physical aspects that will be discussed later on (Fig. 1.1).



Figure 1.1: Countries and topography of West Africa (Source : [Parker and Diop-Kane, 2017, Chapter 1]).

The topography of West Africa is diverse, with a mix of sand dunes in the Sahara Desert and plains in the coastal regions, passing through mountain ranges in the easternmost and westernmost regions. In the west, plateau mountains form the Guinean Highlands which link Guinea to the western Côte D'Ivoire through Sierra Leone and Liberia. In Guinea, these high elevated areas are located in the central part of the region, where they are known as the mountainous massif of Fouta Djallon, peaking at 1515 m (the mont Loura). In the southeast, the Simandou Range extends and peaks at 1658 m (Pic de Fon), as well as the Kourando massif which reaches 1236 m. The highest point of the Guinean Highlands is the mount Bintumani, 1945 m, in Sierra Leone. Hamilton et al. [2017] showed that Guinean Highlands induce an upslope of winds in that region and induce the development of African easterly waves. Those waves spread through the region, generating low-level cyclonic vorticity and convergence, which results in heavy convective rain. To the east of the Guinean Highlands, the Atakora range crosses Benin and Togo. Its

highest elevation is reached in southwestern Togo, and is known as the Pic d'Agou (986 m). In the eastern part of WA, there are two important relief structures: the Jos plateau in Nigeria, culminating at 1829 m, and the Cameroon mountains that reach the highest point of 4095 m at Mount Cameroon. That mount is a part of the Cameroon Volcanic Line.

From a bioclimatic point of view, West Africa can be divided into five different regions, based on the spatial distribution of the rainfall and vegetation (Fig. 1.2) (more details can be found in CILSS [2016]). From the north to the south, the first region encountered is the Sahara, which is characterized by the presence of a desert that covers an important part of the northern Africa. The annual rainfall in this region ranges from 0 to 150 mm over the 1901-2020 period. As we will show below, the development of the heat low in this region induces a shallow meridional circulation over West Africa (Sect. 1.1.3.5), a key aspect of the West African Monsoon (WAM) dynamics. Bare soil and very sparse vegetation dominate the Saharan area (Fig. 1.3).

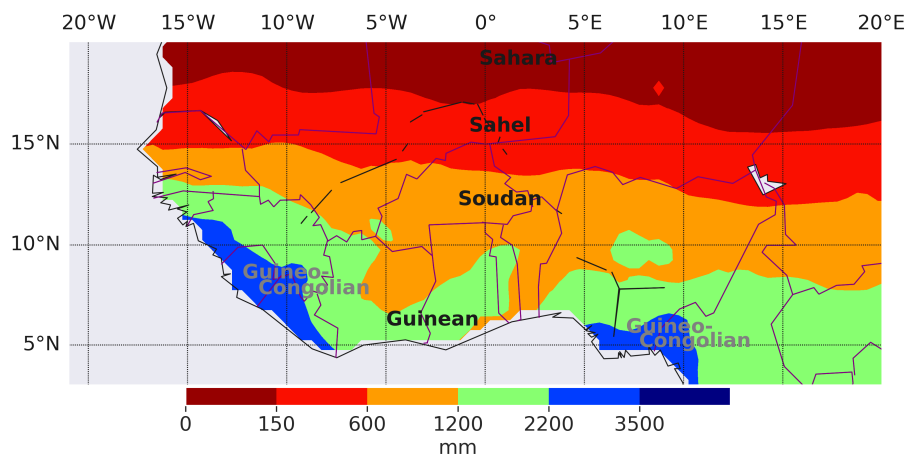


Figure 1.2: Bioclimatic zones of West Africa based on annual rainfall mean over the 1901-2020 period. Data source: Climatic Research Unit gridded Time Series [Harris et al., 2020].

The Saharan region is followed by the Sahelian region, which is characterized by an annual rainfall that ranges from 150 to 600 mm. This region is covered by an herbaceous vegetation type such as small grass savanna and steppe (Fig. 1.3). The third region, the Soudan, displays an annual rainfall that ranges between 600 and 1200 mm. The landscape is characterized by savannas, shrubland, and croplands. South to the Soudan lies the Guinean region, where the annual rainfall ranges between 1200 and 2200 mm. This region is covered by very dense forests and woodlands. The last region is termed "Guineo-Congolian". It receives an an-

nual rainfall between 2200 and 3500 mm and is divided into a western part (the Upper Guinean) and an eastern part (the Lower Guinean). Very dense forests dominate the vegetation in this region. Note that the limits of these regions are not fixed, and can vary depending on the applications. For example, focusing on the boreal summer rainfall variability over West Africa, two main regions can be identified: the Sahel (between 10 and 20°N) and the Guinea Coast (between 4 and 10°N), respectively [Giannini, 2003]. These regions are the center of action of the leading modes of West African rainfall variability over the last century.

With a population of 430 millions in 2022 (5 % of the world population), West Africa is among the world regions with the fastest growth rate since 1950 and its population is expected to exceed one billion people by 2100 [United Nations, Department of Economic and Social Affairs, Population Division, 2022]. Farming activities, fishing, and breeding are key sectors of the West African economy, mainly in the rural areas. This fast population growing rate will increase the water demand of the region, with a strong impact of any West African Monsoon variation as the monsoon is the main water supply source. Changes in the hydrological cycle also affect hydroelectric power generation in the region, where solar and wind farms are less developed. In the next section, we present an overview of the WAM and highlight the detailed atmospheric circulation characteristics over the region in four standard seasons of the year: December-February (DJF), March-May (MAM), June-August (JJA) and September-November (SON).

1.1.2 Classical picture of the West African Monsoon (WAM)

Much of the rainfall over West Africa occurs during the monsoon season. The monsoon onset can be defined in terms of rainfall characteristics, using for example a threshold in the daily rainfall to quantify the magnitude and the duration of rainfall episodes in a season. A standard mean onset date of monsoon over the region has been defined as the 24th of June, although delays can occur in different years [Sultan and Janicot, 2003]. This date can also vary across different datasets, and different definitions at local and regional scales [Fitzpatrick et al., 2015].

During the past decades, many studies were focused on the West African climate [Dettwiller, 1965; Germain, 1968; Hamilton et al., 1945; Sansom, 1965; Walker, 1957]. It was thought that the rainfall associated with the WAM is a result of local thermal instability facilitated by the convergence of low-level trade winds during the WA rainy season. The region of the convergence is termed the Intertropical Convergence Zone (ITCZ, Fig. 1.4(a)), where hot and dry northeasterly "harmattan" winds blow over the Sahara and meet cold and moist southwesterly winds that emanate from the Atlantic Ocean (Fig. 1.4(b)). The ITCZ is assumed

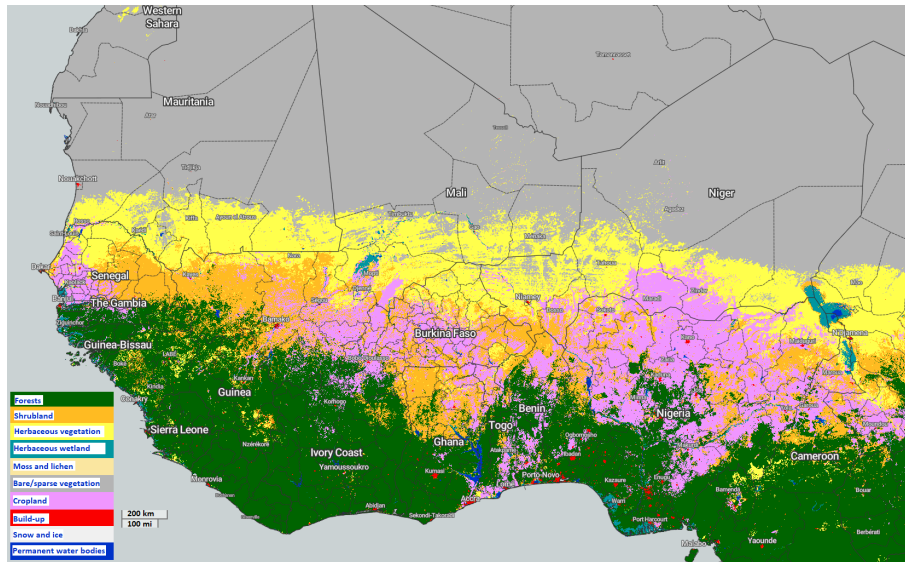


Figure 1.3: 2019 Land cover classification over West Africa. Source: Copernicus Service Information 2019 [Buchhorn et al., 2019].

to "follow the Sun," by moving northward in the boreal summer and southward in the austral summer, traversing twice the equatorial regions [Nicholson, 2009].

This classic picture of the WAM has several shortcomings. The first one raised by Nicholson [2013] is that the ITCZ concept has been developed in association with the global Hadley circulation (Fig. 1.4(a)), which describes the global mean state of inter-tropical regions, and not individual regions. The scheme describes well the state of the ocean (marine ITCZ, although this is strictly valid over all ocean basins) where trade winds are strong but the latter are too weak over lands to induce an important rising motion due to their convergence. Consequently, they cannot be associated with important rainfall at the location where they meet. The second issue is related to the ambiguity of the term ITCZ, which is defined in three different ways in literature, based respectively on wind convergence, surface air pressure and rainfall or outgoing long-wave radiation. As a result, the tracking of the ITCZ may be based on a pressure minimum, a surface wind convergence, a maximum rainfall, a maximum cloudiness or a minimum outgoing long-wave radiation. Unfortunately, all those conditions are not obtained at a single location. For example, defining the ITCZ as the location of the wind surface convergence, Nicholson [2009] found that it is shifted to the north by 1000 km compared to the zone of maximum rainfall and the zone of maximum ascent which are termed the

(a)

WARM DRY

WARM DRY

COOL DRY

COOL DRY

TRADE WIND BOUNDARY LAYER

TRADE WIND INVERSION

SUBTROPICAL HIGH

EQUATORIAL TROUGH

SUBTROPICAL HIGH

HADLEY CELL

ITCZ

HADLEY CELL

(b)

N

20

10

0

1000 ft.

HIGH CIRRUS

DRY NE HARMATTAN

MOIST EASTERLY AIR

MOIST SW MONSOON

Cu

Cu

Cu

Cu

STABLE

A

ITCZ

B

C

D

18°N

12°N

6°N

S

Weather zones of West Africa.

1.1.3 Revised view of the West African Monsoon picture

- a reduction of the importance of the ITCZ;
- an addition of the role played by several jetstreams at different levels and different locations: among them, there are the African Easterly Jet (AEJ), the Tropical Easterly Jet (TEJ) and the Low-level Westerly Jet (LLWJ);
- an importance of the African Easterly Waves (AEWs);

- a clear role of the low surface pressure system over North Africa, due to high surface temperature, named Saharan Heat Low (SHL) and its associated Shallow Meridional Circulation (SMC) [Shekhar and Boos, 2017];
- a teleconnection relationship with several world oceanic basins (discussed in details in Sect. 1.2);

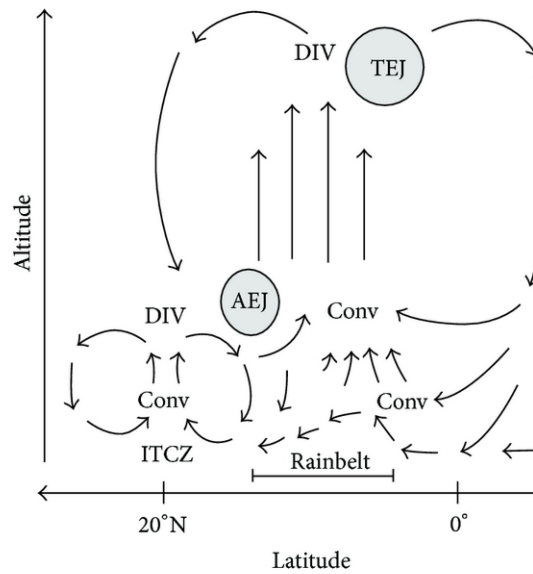


Figure 1.5: Revised WAM. The terms AEJ and TEJ show the positions of the Tropical Easterly Jet and the African Easterly Jet. Conv and Div represent areas of convection and divergence, respectively. Figure from Nicholson [2009].

The monsoon system is basically regulated by the surface temperature gradient between the WA continent and the Atlantic Ocean. During the boreal winter, which is termed the harmattan season (DJF), strong northeasterlies prevail over West Africa, while the southwesterlies penetration into the continent (the monsoon layer) is limited to the coastal regions (Fig. 1.6 (a)). The zone of confluence of the two air masses over the continent is named the Intertropical Discontinuity (ITD), or the Intertropical Front (ITF), whose position is monitored by forecasters to follow the progression or retreat of the monsoon air layer. The harmattan season is the driest season of the year, with rainfall intensity lower than $3 \text{ mm} \cdot \text{day}^{-1}$. Knippertz and Fink [2009] found that an average of 1.87 wet events occurs each boreal winter. They result from subtropical disturbances at upper levels and a reduction of the mean sea level pressure over the Sahara which induce a moisture advection

from southerlies into West Africa. In their work, the dry season is between the 2nd of November and the 1st of March, and an event is considered as wet if the rainfall amounts to 3 times the climatology over the 1979-2002 period.

Progressively, the rainfall band migrates northward in March-May (MAM), which corresponds to the first rainy season over the Guinea Coast (Fig. 1.6(b)). Northerlies and northeasterlies are limited to 15°N, whereas the southerlies and southwesterlies from the equatorial Atlantic progress into the continent. The heat low development is limited to the central to eastern Sahel in this season (see also Sec. 1.1.3.5, Fig. 1.10). In June-August (JJA), the monsoon layer is further north over West Africa (Fig. 1.6 (c)). A cold tongue develops over the Atlantic Ocean [Kouassi et al., 2017], while the West African Heat Low (WAHL) strengthens and is positioned over the northern Mali (Fig. 1.6 (c)). This allows the intensification of northerlies in the western flank of the heat low and a cyclonic circulation around the heat low. The Guinean Highlands and Cameroon mountains show the highest rainfall amounts over this season. At the same time, coastal areas experience the "little dry season" [Adejuwon and Odekunle, 2006], due to the development of the Guinea Coast upwelling [Tanguy et al., 2022]. In September-November, the monsoon layer is shifted southward, with a reduction of the rainfall over the Sahelian region (Fig. 1.6 (d)).

Thorncroft et al. [2011] described the WAM in four phases: an oceanic phase (from November to mid-April), a coastal phase (from mid-April to the end of June), a transitional phase (the first half of July) and a sahelian phase (mid-July to September). The first three phases are called the oceanic regime by Lebel [2003], and the last one, the continental regime. Next, we will see in detail the above-mentioned additional aspects of the revised picture of the WAM.

1.1.3.1 The reduction of the importance of the ITCZ as source of rainfall

In a typical month of August, Nicholson [2009] showed that the "land ITCZ" coincides with the position of the Saharan heat low (around 20°N). The cyclonic circulation in the "land ITCZ" is associated with a shallow overturning cell in the region, with a surface ascent bringing much less rainfall than the deep cell of the classical view. Instead, the rainfall maximum over West Africa in a typical month of August lies south of the surface convergence in the ITD (the term ITD must be used instead of "land ITCZ"). This maximum rainfall position coincides with the maximum ascent over the Sahel, between the axes of the AEJ and TEJ. Over the Guinea Coast, surface convergence and ascent are due to friction between the low-level southwest monsoon flow and the coast. Moreover, the depth of the humid air layer over West Africa seems constant between 5°N and 20°N, which is contrary to the arguments in the classic view [Nicholson, 2009]. In other words,

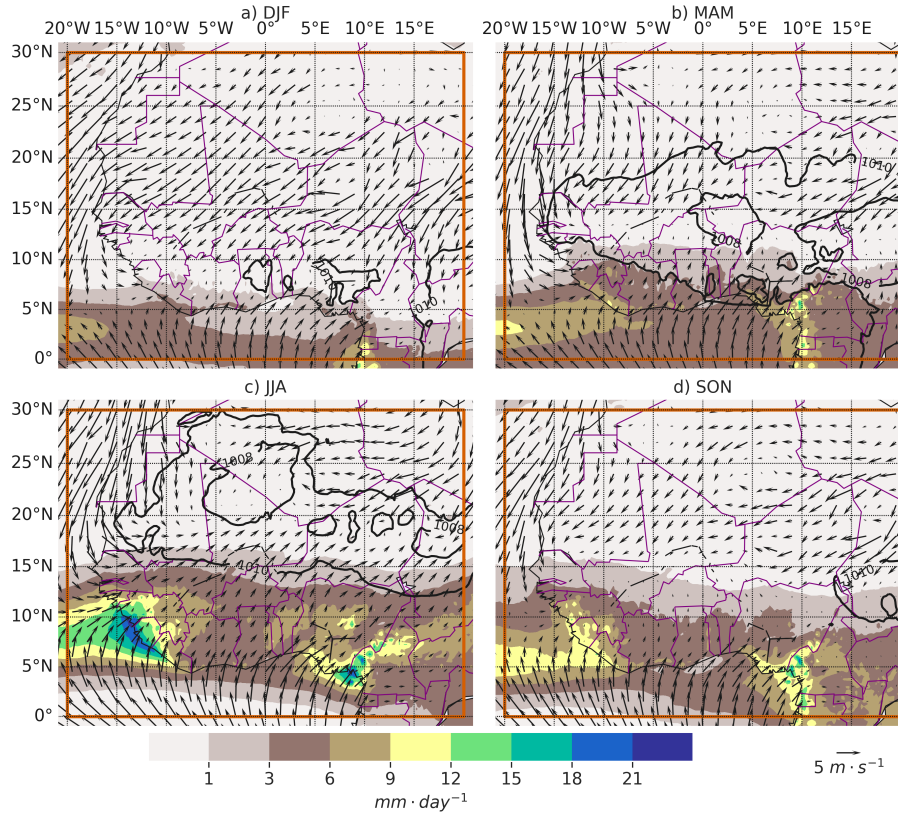


Figure 1.6: Seasonal average of the mean daily rainfall (in colours), mean sea level pressure (in contours) and 10 m wind vectors over the 1992-2021 period, for the (a) DJF, (b) MAM, (c) JJA and (d) SON seasons. Data source: ERA5 [Hersbach et al., 2020].

whatever the season, changes in the rainfall over West Africa depend on additional key elements, other than the north-south migration of the ITCZ and the ITD.

1.1.3.2 The African Easterly Jet (AEJ) and African Easterly Waves (AEWs)

In the boreal summer, the AEJ takes its origin in the meridional surface temperature gradient and soil moisture gradient between the Sahara desert and the Atlantic Ocean to the south [Cook, 1999; Thorncroft and Blackburn, 1999]. This mid-tropospheric jetstream is located at around 650 hPa and its core lies on average over the western Sahel (Fig. 1.7 (c)). At that vertical level, the sign of the meridional temperature gradient is reversed compared to the sign at the surface

[Kebe et al., 2016]. In the 1992-2021 climatology, the JJA position of the AEJ is centered around 15°N , and its speed ranges between 11 and $13 \text{ m} \cdot \text{s}^{-1}$ (Fig. 1.7 (c)). The southern flank of the cyclonic circulation around the WAHL might reinforce the westward direction of the AEJ. During the harmattan season, the AEJ is displaced equatorward, and its speed ranges between 5 and $7 \text{ m} \cdot \text{s}^{-1}$ over the Guinea Coast (Fig. 1.7 (a)). This speed increases up to $11 \text{ m} \cdot \text{s}^{-1}$ in MAM over the Guinea Coast (Fig. 1.7 (b)). The core of the jet then jumps to the center of the Sahel in JJA, before it retreats southward in SON (Figs. 1.7 (c) and (d)).

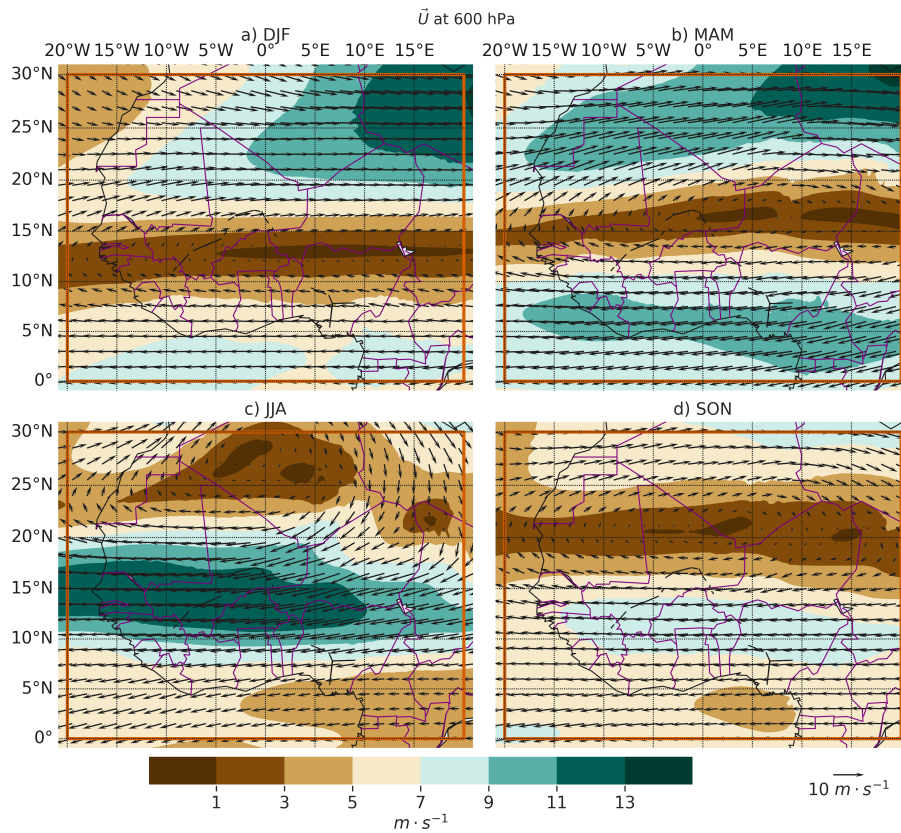


Figure 1.7: Seasonal mean of the 600 hPa wind over the 1992-2021 period for the (a) DJF, (b) MAM, (c) JJA and (d) SON seasons. Data source: ERA5 [Hersbach et al., 2020].

Moreover, vertical and horizontal shears in this westward flow lead to baroclinic-barotropic instabilities, which are the sources of the African Easterly Waves (AEWs) [Thorncroft and Hoskins, 1994a,b]. These latter modulate the frequency and magnitude of cloud cluster and the amount of rainfall [Nicholson and Grist, 2003]. Other

processes such as cumulus convection within the ITD could also generate the AEWs [Hsieh and Cook, 2005]. AEWs and the associated cloud system remain in a narrow band of latitude near and southward of the AEJ core, in the axis of the AEJ and tropical easterly jet (TEJ).

1.1.3.3 The Tropical Easterly Jet (TEJ) and Subtropical Jet (STJ)

The TEJ is an important feature of the WAM circulation whose origin is not well known. It could result from the intense north-south temperature gradient between the Himalayan Plateau and the Indian Ocean and then being associated with the summer Indian monsoon processes [Koteswaram, 1958; Gu and Adler, 2004; Nicholson and Webster, 2007; Nicholson et al., 2007; Lemburg et al., 2019]. This jet is present only during the boreal summer in West Africa and is also thought to result from the high-level gradient of geopotential height between the Sahara and the Sahel/central equatorial Africa [Fink et al., 2017]. Once formed, the east-west Walker circulation and the north-south Hadley circulation contribute to its maintenance. It is an upper-level jetstream located at 200 hPa at around 8°N in boreal summer (Fig. 1.8 (c)). This latitudinal position seems to be maintained from year to year, and the wind speed in the JAS 1992-2021 climatology ranges from 10 to 15 m · s⁻¹. Stronger TEJ are statistically associated with wetter conditions over the Sahel, and inversely [Hulme and Tosdevin, 1989; Grist and Nicholson, 2001; Nicholson, 2008; Lemburg et al., 2019; Nicholson and Klotter, 2020]. Note that the TEJ can move southward in the Southern Hemisphere (5 to 10°S) during the austral summer, where it can be very strong.

The link between the TEJ and the Sahelian rainfall in the monsoon season could be mainly explained by the upper-level divergence that occurs in the jet core over WA, which could induce convection and rising motion over the Sahel. Other mechanisms such as wave activities within the jet and their impact on the formation of mesoscale convective systems (MCSs) have been advanced [Nicholson et al., 2007, 2008; Besson and Lemaître, 2014]. Interestingly, Lemburg et al. [2019] found no link between the TEJ and MCSs initiation over West Africa. The existence of a common driver (e.g. sea surface temperature (SST) modes of variability) that could force both the rainfall and the TEJ has also been speculated [Nicholson and Klotter, 2020].

Furthermore, in DJF, MAM and SON, the high-level circulation field over West Africa is dominated by a westerly flow: the Subtropical Jet (STJ) (Figs. 1.8 (a), (b) and (d)). The STJ is very intense over the Sahel and the northern Africa, mainly in DJF and MAM.

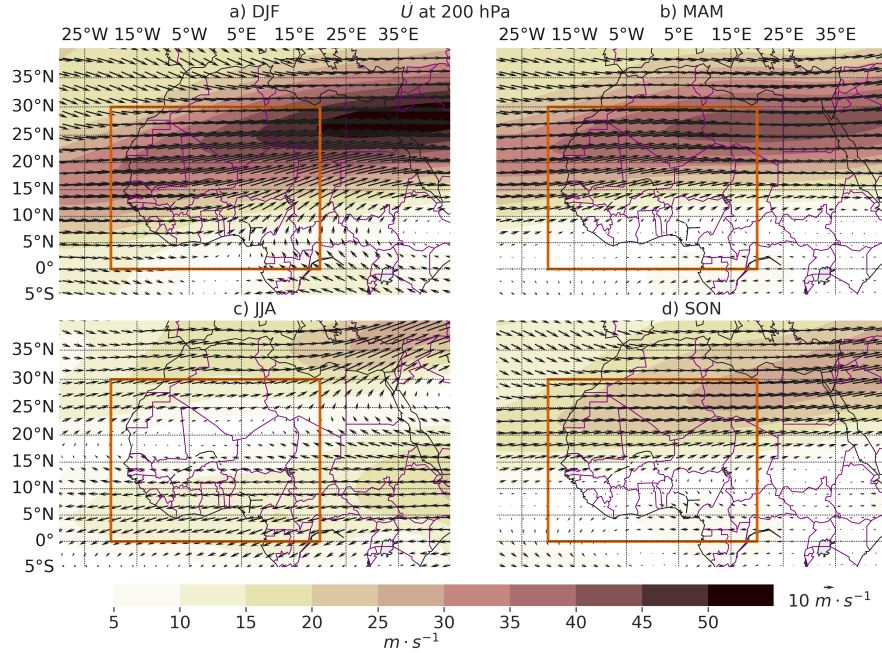


Figure 1.8: Seasonal mean of the 200 hPa wind over the 1992-2021 period for the (a) DJF, (b) MAM, (c) JJA and (d) SON seasons. The brown box indicates the West African region. Data source: ERA5 [Hersbach et al., 2020].

1.1.3.4 The West African Westerly Jet

The West African Westerly Jet (WAWJ) is a low-level westerly wind jet that is different from the low-level monsoon flow. It develops over an area where the oceanic ITCZ coincides with the oceanic extension of the continental thermal heat low (8 - 11°N), between the eastern Atlantic and West African coast [Grodsky, 2003; Pu and Cook, 2010, 2012; Liu et al., 2019]. The WAWJ is characterized by a vertical extent from the surface to 700 hPa level. The speed of the WAWJ is the highest from July to September, with a maximum at around 925 hPa. In the JJA climatology at 925 hPa over the 1992-2021 period, this maximum speed ranges between 2 and 7 $\text{m} \cdot \text{s}^{-1}$ (Fig. 1.9 (c)). In the other seasons, the WAWJ is weaker or non-existent (Figs. 1.9 (a),(b) and (d)). The main importance of the WAWJ is to collect moist air above the Atlantic Ocean and carry it to the WA region. This role is crucial during years when the monsoon southerly flows are weak.

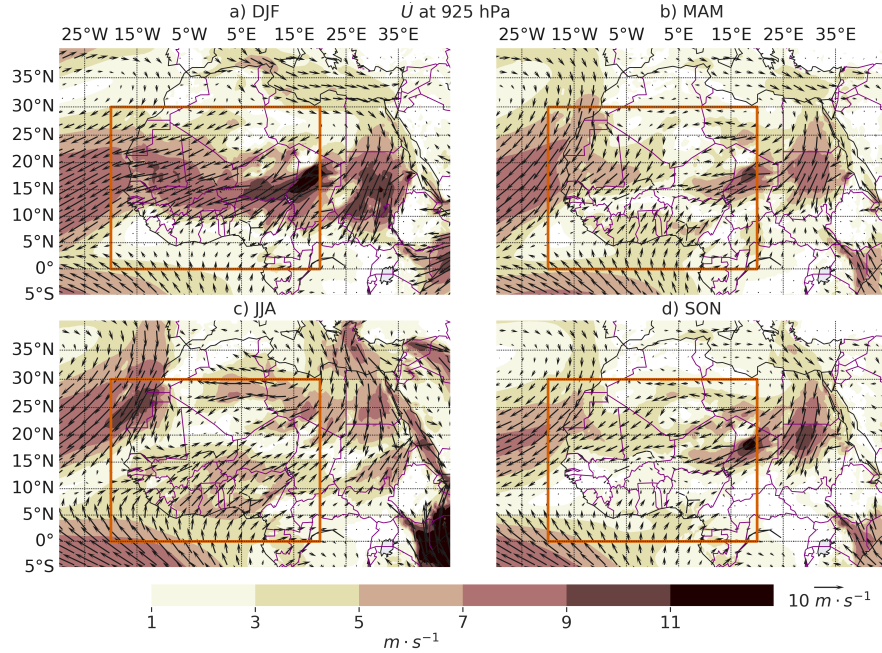


Figure 1.9: Seasonal mean of the 925 hPa wind over the 1992-2021 period for the (a) DJF, (b) MAM, (c) JJA and (d) SON seasons. Data source: ERA5 [Hersbach et al., 2020].

1.1.3.5 The West African heat low and the Saharan heat low

The West African heat low (WAHL) is formed over the continental area of West Africa where high incoming solar radiation and low evaporation lead to a very high surface temperature and a low surface pressure [Ramel et al., 2006; Lavaysse et al., 2015]. During the SON, DJF and MAM seasons, the WAHL is located in the south of the Darfour mountains (Fig. 1.10 (d), (a), (b)). Then, in JJA, the WAHL position migrates northwestward and is centered over the Sahara desert, in the north of Mali and Mauritania (Fig. 1.10 (c)). At this period of the year, it is named "the Saharan Heat Low (SHL)" [Lavaysse et al., 2009]. Usually, the WAHL and the SHL terms are used without any difference between them.

The SHL is associated with a shallow meridional circulation which is characterized by a northward near-surface moist flow that comes from the coastal region of West Africa, crosses the Sahel and reaches 20°N. The zonal extension of this flow covers the Sahara desert to 40°E, and is associated with a cyclonic circulation over the low. From the surface to 700 hPa, the vertical velocity is upward. Above this level, the winds become anticyclonic, with a divergent southward flow [Shekhar and Boos, 2017]. This mid-level anticyclonic motion also reinforces the

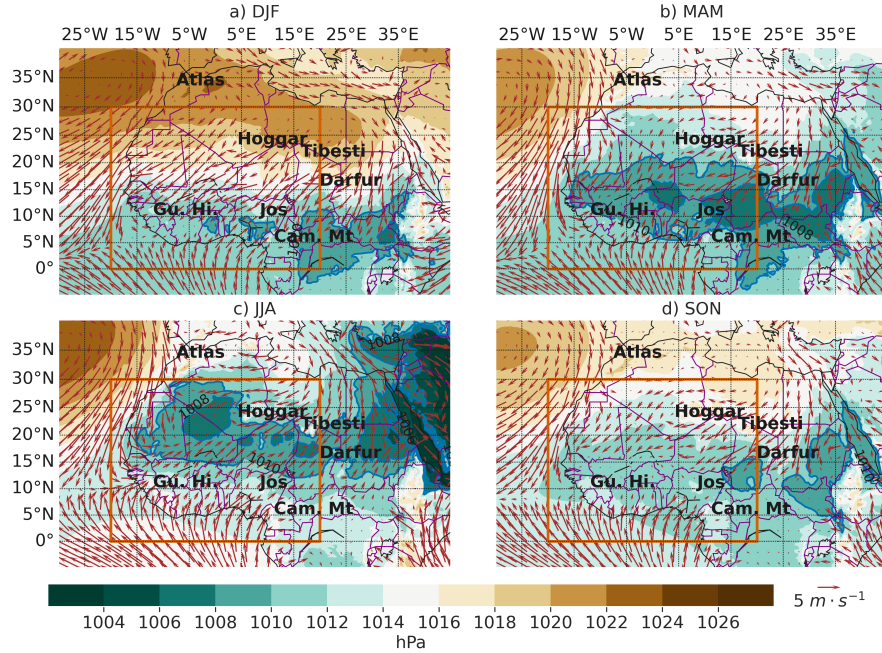
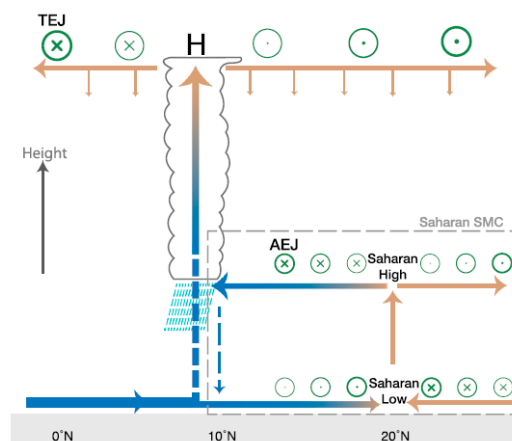


Figure 1.10: Seasonal mean of the mean sea level pressure (in colors) and 10 m wind vectors over the 1992-2021 period, for (a) DJF, (b) MAM, (c) JJA and (d) SON seasons. Names of some important mountains are indicated. Data source: ERA5 [Hersbach et al., 2020]

AEJ [Lavaysse et al., 2009]. The SHL and its associated SMC are important components of the WAM (Figure 1.11).

The SHL exhibits intraseasonal, interannual and decadal variations, with an important impact on the Sahel rainfall. During warm phases of the Atlantic multi-decadal variability [AMV, a North Atlantic basin SST fluctuation on multi-decadal timescale, Zhang et al., 2019; Deser and Phillips, 2021; Lai et al., 2022], the Sahel SMC strengthens and enhances rainfall over the Sahel [Martin and Thorncroft, 2014]. Lavaysse et al. [2015] demonstrated that the SHL has different impact on the West African circulation on different timescales over the 1979-2012 period. The high-frequency variability (less than 25 days) of the SHL is related to a low-level cyclonic circulation around the center of the heat low and an increase of southerlies in the eastern Sahel, which reinforces the rainfall in this area, while the rainfall decreases in the eastern Guinea Coast. In the western part of West Africa, the rainfall decreases over the Sahel and increases over the Guinea Coast. The variability of the SHL in the 15-100 days band is associated with the Madden-Julian



Oscillation, and shows an overall decrease of the rainfall over West Africa. For the low-frequency variability of the SHL (above 300 days), Lavaysse et al. [2015] found an enhanced westerly flow over the entire Sahel. This induces an increase (a decrease) of the rainfall over the central and eastern Sahel (the western Sahel and Guinea Coast). Interestingly, Shekhar and Boos [2017] recently suggested that the strengthening of the Saharan SMC advects warm and dry air in the mid-troposphere, which reduces rainfall over the Sahel. This result is thus contrary to the idea that strong SHL and its associated SMC lead to wet conditions over the Sahel [Haarsma et al., 2005; Biasutti et al., 2009; Lavaysse et al., 2015]. This can be considered as an open debate.

Figure 1.12 summarizes in a 3D view the different features of the WAM that have been discussed. Next, we focus on the oceanic forcings of the rainfall variability in West Africa, through changes in the circulation described above. In the literature, much of these kinds of analysis are focused on the July-September season, when the rainfall over the Sahel is maximum [Rowell, 2003; Fontaine et al., 2009; Park et al., 2016; Mohino et al., 2011a, 2016]. To take into account the variability of the rainfall over the Guinea Coast, we will consider the June-September (JJAS) season, following Polo et al. [2008].

1.1.3.6 The Rodwell-Hoskins mechanism and the ventilation

Other mechanisms such as soil moisture feedback, ocean heat transport, ventilation and interactive Rodwell-Hoskins mechanism [Chou et al., 2001; Chou and

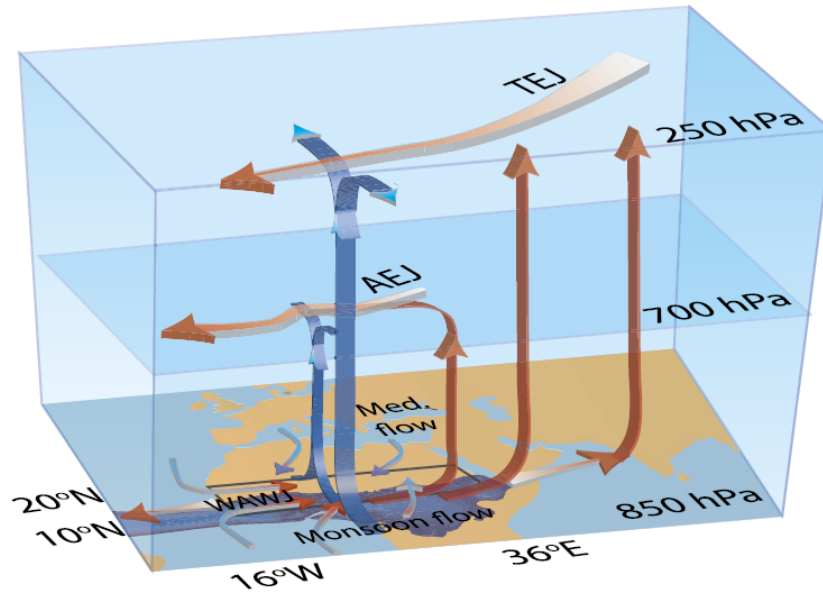


Figure 1.12: 3D view of the different atmospheric circulation characteristics of the WAM, from Sheen et al. [2017].

Neelin, 2003] could also play a major role in the monsoon regions. A positive net heat flux into the atmosphere in the summer is thermodynamically favorable to convection where the criteria are met, and to a poleward extent of the convection zone. However, some dynamical factors such as ventilation and interactive Rodwell-Hoskins mechanisms could limit this poleward extent of the monsoon convergence zone. The Interactive Rodwell-Hoskins [Rowell et al., 1995; Rodwell and Hoskins, 1996] mechanism refers to the interaction between the baroclinic Rossby wave dynamic and the convective heating (the Rossby waves induce downward motion to the west of the diabatic heating zone), whereas the ventilation is defined by temperature and moisture advection (advection of moist static energy). These dynamical mechanisms are however less important for the African monsoon, where the strong albedo of the Saharan region leads to a weaker positive net heat flux into the atmosphere, compared to the other monsoon regions of the Earth.

1.2 Variability of rainfall and of sea surface temperature

An empirical orthogonal function (EOF) technique can be used to identify the main rainfall modes of variability over West Africa [Giannini, 2003; Monahan et al., 2009]. The application of this technique reveals two leading modes of rainfall variability which explains 53% of the total rainfall variability over the region, and during the 1901-2020 period (Fig. 1.13).

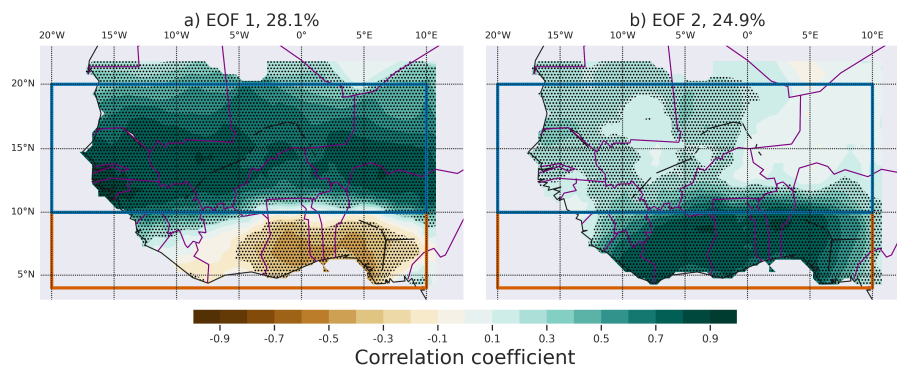


Figure 1.13: EOF1 (a) and EOF2 (b) of the JJAS rainfall anomalies over West Africa and during the 1901-2020 period. They represent the correlation between their respective standardized principal component time series and the JJAS rainfall anomalies over West Africa. The total rainfall variance explained fraction by each mode is indicated in percentage. Stippling indicates significant correlations at 95% confidence level according to a Student test. Rainfall data source: Climate Research Unit [Harris et al., 2020].

1.2.1 The Sahelian rainfall mode

The JJAS leading rainfall mode of variability over the West Africa in the 1901-2020 period corresponds to the Sahelian mode (EOF1). It explains 28% of the total rainfall variability over the region (Fig. 1.13 (a)). The spatial pattern of the EOF1 exhibits a rainfall dipole with opposite signs over West Africa. One pole is centered over the Sahelian region, between 10°N and 20°N. The second pole is centered over the Guinea Coast region (south of 10°N). This pattern shows that, over the 1901-2020 period, an increase in rainfall over the Sahelian region is associated with a decrease in rainfall over the Guinea Coast. However, after the 1970s, the coupling between the rainfall in the Sahel and the Guinea Coast weakened, and the rainfall dipole disappeared [Losada et al., 2012].

1.2.1.1 SST impact on the Sahelian rainfall on the decadal to multi-decadal timescale

Figure 1.14 (a) illustrates that the principal component time series corresponding to the EOF1 (PC1) are very close to the time series of the rainfall anomalies averaged over the Sahel region. The correlation between the two time series is 0.93. The rainfall variability over the Sahel can be separated into a low-frequency component and a high-frequency component. Figure 1.14 (a) shows a strong multi-decadal variability in the Sahelian rainfall, with dominant wetter conditions in the first half of the 20th century, and drier conditions afterwards. Particularly, there is a negative trend between the 1950s and 1980s, while the trend is positive later on. Giannini [2003] demonstrated that the oceanic SST variability is the main driver of the multi-decadal variability of the rainfall over the Sahel, which is amplified by land-atmosphere processes.

Mohino et al. [2011a] attributed the Sahelian drought in the 1970s and 1980s to the global warming of the oceanic basins (mainly the tropical Atlantic), the positive phase of the inter-decadal Pacific oscillation [IPO, a basin wide SST fluctuation in the Pacific Ocean on multi-decadal timescale, Power et al., 1999; Deser et al., 2004; Dai, 2012; Henley et al., 2015] and the negative phase of the AMV. Moreover, Mohino et al. [2016] showed that the AMV is the main predictor for the hindcast of the Sahelian rainfall on the multi-decadal timescale. Giannini [2003] and Giannini et al. [2005] also suggested that the warming trend of the Indian Ocean (also found by Bader and Latif [2003]) and the southern Atlantic basin have contributed to the negative trend of the Sahelian rainfall between the 1950s and the 1980s. The recent recovery of the rainfall over the Sahel has been attributed, on the one hand, to the anthropogenic warming of the Mediterranean basin [Park et al., 2016]. On the other hand, Dong and Sutton [2015] argued that this recovery of the Sahelian rainfall over the last decades is mainly due to the increase of the greenhouse-gas concentration in the atmosphere, which is amplified by changes in aerosol forcing. They noted, however, that different SST background mean states can modify the magnitude of the Sahelian rainfall changes due to greenhouse gas and anthropogenic aerosol forcings.

Despite these contrasting results, monitoring the ocean-atmosphere-land interactions is a key element for understanding the changes in the rainfall variability over West Africa. SST anomalies (SSTAs) regressed onto the low-frequency component of the PC1 over the 1901-2020 period (Fig. 1.15 (e)) indicate warmer than normal surface conditions over the North Atlantic, the Mediterranean Sea and the extratropical North Pacific associated with wet conditions over the Sahel, and inversely. Furthermore, colder than normal sea surface conditions in the Indian Ocean are related to wetter conditions over the Sahel, and inversely.

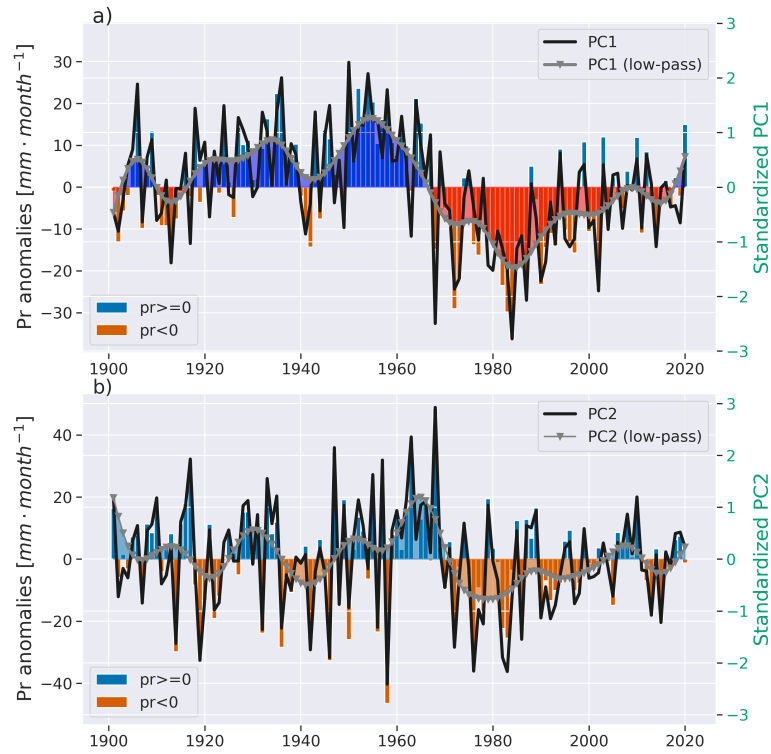


Figure 1.14: (a) Time series of the JJAS rainfall anomalies averaged over the Sahel region (blue and red bars, left scale), the first principal component (PC1, black curve, right scale), and the low-pass PC1 (gray curve, right scale) during the 1901-2020 period. (b) Time series of the JJAS rainfall anomalies averaged over the Guinea Coast region (blue and red bars, left scale), the second principal component (PC2, black curve, right y-axis), and the low-pass PC2 (gray curve, right scale) during the 1901-2020 period. The low-frequency timeseries are obtained by applying a fourth order Butterworth filter with a cutoff of $\frac{2}{13\text{years}}$.

1.2.1.2 Links with the tropical Pacific and Indian Oceans on the interannual timescale

On the interannual timescale, an increase (a decrease) in the rainfall anomalies over the Sahel is associated with below (above) normal SST conditions over the eastern tropical Pacific (Fig. 1.15 (c)). The relation between the Sahelian rainfall and the tropical Pacific appears as the leading mode of covariability between the West African rainfall and the global tropical basins [Rodríguez-Fonseca et al., 2011]. Rowell [2001] demonstrated that an El-Niño phase is characterized by a low-level convergence over the eastern equatorial Pacific and by an upper-level divergence which excites an eastward propagation of a Kelvin stationary wave. This wave crosses the Atlantic Ocean and leads to a large-scale convergence over the Sahel in the upper levels. It results in a subsidence over the Sahel, which warms the troposphere and reduces the convective instability. These conditions disfavor the rainfall occurrence over the Sahel. Moreover, Rowell [2001] argued that the subsidence over the Sahel can be reinforced by stationary Rossby waves originating from an enhanced convection over the Indian Ocean.

Throughout the seasons, changes in the positions of the strongest SST anomalies over the tropical Pacific can lead to diverse impacts of El-Niño and La-Niña events on the Sahelian rainfall. Mohino et al. [2011b] showed that, during the boreal spring, above normal SST conditions in the eastern Pacific and the western Indian Ocean induce a strong low (high) -level convergence (divergence) over the eastern Pacific and a weak low (high) -level convergence (divergence) over the Indian Ocean. These two rising branches generate a large-scale subsidence over the Maritime Continent (an oceanic region between the eastern Indian Ocean and the Western Pacific Ocean) and the equatorial Atlantic (and Guinea Coast). This latter is compensated by an enhanced ascent over the Sahel, which increases rainfall there. During the boreal summer, the highest SST anomalies extends westward over the central Pacific, while the anomalies over the western Indian Ocean weaken. These SST conditions lead to a large-scale upper-level convergence and subsidence over the West Africa. Subsequently, the rainfall is reduced over the entire West Africa. Furthermore, some studies pointed out an opposite phase relationship between the Indian basin surface temperature and the Sahelian rainfall [Biasutti et al., 2008]. More specifically, Bader and Latif [2009] argued that the Sahelian drought in 1983 was mainly forced by warm sea surface conditions over the Indian basin.

1.2.1.3 Links with the Mediterranean Sea on the interannual timescale

During the 1901-2020 period, positive SST anomalies in the Mediterranean Sea and over isolated areas in the extratropical North Atlantic are associated with increased Sahelian rainfall (Fig. 1.15 (c)). Focusing on the influence of the Mediter-

anean Sea, Rowell [2003] argued that years when this basin is warmer (colder) than normal are linked with wetter (drier) than normal conditions over the Sahel. Abnormal warming of the Mediterranean sea surface increases the low-level convergence and the rising motion over the basin. The moist air is then advected southward by the low-level flow and penetrates into the Sahel. This enhances the moisture convergence over the Sahel and tends to increase its rainfall. Moreover, Fontaine et al. [2009] demonstrated that there are differences between the eastern and the western Mediterranean basin forcings on the rainfall over the Sahel. By conducting sensitivity experiments with atmospheric general circulation models, they showed that above normal surface temperatures in the Eastern Mediterranean Sea are associated with an enhancement of the low-level monsoon flow, a decrease in the mid-level AEJ, an enhancement of the Tropical Easterly Jet, a northward migration of the ITD and a weakening of the northeasterlies.

Furthermore, the increase of the rainfall in Sahel associated with warm conditions in the Mediterranean Sea appears as the leading mode of covariability between SST anomalies in the Mediterranean Sea and the rainfall anomalies over the entire West Africa [Polo et al., 2008]. The large-scale SST pattern related to this leading mode of covariability is shifted from a more localized SST anomalies in the 1979-1998 period to more global SST anomalies in the 1957-1978 period, which involve the tropical Atlantic and Pacific [Rodríguez-Fonseca et al., 2011]. Suárez-Moreno et al. [2018] showed that the relation between the Mediterranean SST and the Sahelian rainfall is not stationary.

1.2.1.4 Links with the Atlantic Ocean on the interannual timescale

Figure 1.15 (c) reveals that negative (positive) JJAS SST anomalies in the equatorial and southeastern Atlantic regions are related to an increase (a decrease) in the JJAS rainfall anomalies over the Sahel during the 1901-2020 period. The pattern of SST anomalies resembles the negative phase of the Atlantic Equatorial Mode (AEM), which is the dominant oceanic internal mode of variability in the tropical Atlantic [Zebiak, 1993; Polo et al., 2008; Losada et al., 2010a; Lübbecke et al., 2018]. Alike the AMV, which acts on multi-decadal timescale, the AEM phenomenon occurs on interannual timescales. Warm phases of the AEM are associated with a decrease in the sea level pressure gradient between the West African continent and the Atlantic Ocean. This limits the northward penetration of the monsoon flow into lands, and leads to a reduction of the rainfall over the Sahel. Coastal areas, however, benefit from wetter conditions.

The relation between the Sahelian rainfall and the SST changes in the eastern equatorial Atlantic is non-stationary [Losada et al., 2012; Suárez-Moreno et al., 2018]. The dipolar rainfall anomalies response over West Africa in relation with the AEM is valid before the 1970s, when there were no connection between SST

in the eastern equatorial Atlantic and the eastern equatorial Pacific. However, after the 1970s, the coupling between the two basins was enhanced, with warmer than normal surface conditions in the AEM region occurring simultaneously with colder than normal surface conditions in the eastern Pacific. As positive SST anomalies over the eastern equatorial region induce a decreased Sahelian rainfall, and negative SST anomalies over the eastern equatorial Pacific lead to an increased Sahelian rainfall, the superposition of these counteracting impacts from each basin weakens the Sahelian rainfall response to the AEM phases. The dipolar structure of the West African rainfall responses to AEM events thus disappears, toward a uniform response which is more localized over the Guinea Coast [Worou et al., 2020].

1.2.2 Atlantic equatorial mode: origins and mechanisms

Internal climate variability refers to the variability of the climate system that develops with the action of any external forcings such as solar radiation, volcanic activities, anthropogenic aerosol, and greenhouse gas emissions. It results from processes intrinsic to the different components of the climate system (atmosphere, ocean, cryosphere, land surface, and terrestrial biosphere) as well as interactions between them [Goosse, 2015]. The Atlantic equatorial mode of variability is one of the many internal climate modes of variability of the climate system such as the ENSO (El Niño/Southern Oscillation) phenomenon in the Pacific Ocean, the Indian Ocean Dipole (IOD), the North Atlantic Oscillation (NAO) and the Southern Annular Mode (SAM).

1.2.2.1 Origins and mechanisms

The AEM has a dominant variability on timescales from two to four years [Zebiak, 1993; Servain et al., 1999; Deser et al., 2010]. There are different origins of the AEM events. Canonical AEM events are related to the positive Bjerknes feedback over the equatorial Atlantic, which is considered as the main driver of the AEM variability [Bjerknes, 1969; Chang et al., 2006; Keenlyside and Latif, 2007; Deppenmeier et al., 2016; Silva et al., 2021]. The Bjerknes feedback involves three components. Before addressing the first component, it is worth remembering that during the boreal summer, easterlies dominate the surface wind circulation over the equatorial Atlantic, and a cold tongue is well developed over the eastern equatorial Atlantic (the Atlantic cold tongue). Sea waters are warmer in the western basin, compared to the eastern basin of the tropical Atlantic. The zonal SST gradient favors a zonal surface pressure gradient over the equatorial Atlantic which favors the easterlies.

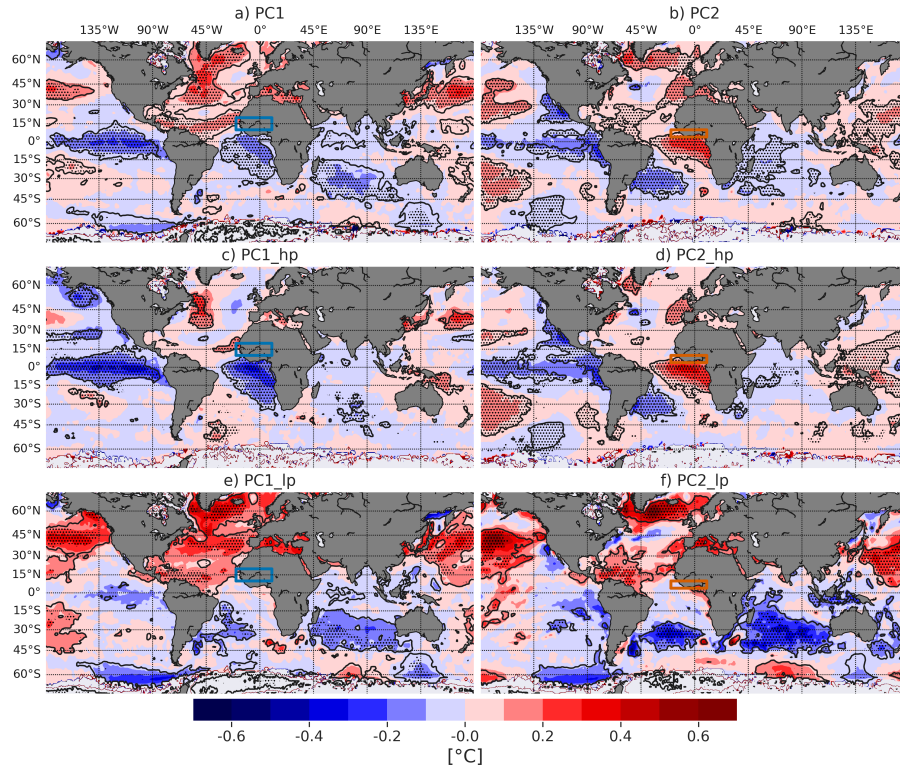


Figure 1.15: JJAS SST anomalies regressed onto the JJAS standardized principal components (a) and (b), the high-frequency principal components (c) and (d), and the low-frequency principal components (e)-(f) during the 1901-2020 period. The low-frequency timeseries are obtained by applying a fourth order Butterworth filter with a cutoff of $\frac{2}{13\text{years}}$. The high-frequency timeseries are obtained by removing the low-frequency timeseries from the raw principal component timeseries. Contour lines and stippling represent statistically significant regression coefficients at 95% confidence level according to a F-test with standard and effective degree of freedom, respectively. The left (right) panels correspond to the first (second) principal component patterns. The blue (orange) boxes indicate the region of Sahel (Guinea Coast).

A positive temperature anomaly in the eastern equatorial Atlantic can start from anomalous vertical upwelling processes [Polo et al., 2015]. This warming then reduces the SST gradient which in turn leads to a relaxation of easterlies west of the perturbation. This is the first component of the Bjerknes feedback, which is the forcing of the surface winds in the western equatorial Atlantic by positive SST anomalies in the eastern equatorial Atlantic. It results in westerly anomalies over the western basin.

The second component of Bjerknes feedback is the forcing of the eastern equatorial Atlantic heat content by the surface westerly anomalies in the western equatorial Atlantic. This forcing operates via eastward propagation of oceanic Kelvin waves and leads to an anomalous increase in the sea surface height (SSH), oceanic heat content, and thermocline depth in the eastern equatorial Atlantic. It can be considered in a first order that a positive heat content anomaly would lead to a deeper thermocline in the eastern equatorial Atlantic [Rebert et al., 1985; Williams et al., 2015].

The last component of the Bjerknes feedback involves the impact of the eastern equatorial Atlantic heat content on the sea surface temperature. The anomalous increase in the eastern equatorial Atlantic thermocline depth reduces the upwelling of cold water. In turn, this reinforces the initial positive SST anomaly in the eastern equatorial Atlantic. The three elements of the Bjerknes feedback are reversed for a cold AEM event: negative SSTAs in the eastern equatorial Atlantic reinforce the west-east zonal surface pressure gradient and strengthen the surface trade winds. This reduces the thermocline depth in the eastern equatorial Atlantic and increases the upwelling of cold water. In turn, this reinforces the initial negative SST anomalies. Keenlyside and Latif [2007] quantified the strength of the Bjerknes feedback over the equatorial Atlantic and found that an initial perturbation of 1°C over the eastern equatorial Atlantic can lead to an anomalous westerly wind stress of $0.75 \times 10^{-2}\text{Pa}$. This in turn would produce one month later a positive SSH of 1.65 cm. The resulting SSH anomaly would lead one month later to an increase of 0.3°C in the eastern equatorial Atlantic SST.

A positive phase of the AEM can be terminated by damping processes such as the surface turbulent latent and sensible heat exchanges between the atmosphere and the ocean [Polo et al., 2015; Hu and Huang, 2006]. These damping processes represent a delayed negative feedback which confer to the AEM an oscillatory system character [Ding et al., 2015]. Interactions with the ITCZ could also play a role in the decay of AEM events (details in Sect. 1.2.2.2).

Additionally, among other processes, a positive phase of the AEM could result from warm water advection from the tropical North Atlantic Richter et al. [2012]. Recently, Martín-Rey and Lazar [2019] argued that the Atlantic Meridional Mode (AMM) events in boreal spring could trigger AEM events in the following summer. AMM is characterized by an interhemispheric gradient of SST anomalies in the

tropical Atlantic, with a positive (negative) core north (south) of the equator during its positive (negative) phase [Nobre and Srukla, 1996; Chiang and Vimont, 2004; Xie and Carton, 2013].

1.2.2.2 Comparison of the AEM with ENSO and IOD : specificities and commonalities

Keenlyside and Latif [2007] compared the Bjerknes feedback in both the tropical Atlantic and Pacific oceans. However, it is weaker in the former compared to the latter. AEM events peak in the boreal summer and last 3 months, with an SST amplitude of about 1 °C [Richter et al., 2016]. The highest SST anomalies in positive AEM phases occur in June, one month after the maximum in the zonal wind anomalies in the western basin. The AEM phenomenon is also known as the Atlantic Niño, because it resembles to the El-Niño phenomenon in the eastern equatorial Pacific [Lübbecke et al., 2018]. However, ENSO phenomenon in the Pacific, peaks in the boreal winter and last 9 months [Richter et al., 2016]. In both cases, the phase locking of warm or cold events (boreal summer for AEM and boreal winter for ENSO) is related to the seasonal migration of the ITCZ (also for IOD events, in boreal fall). More specifically, over the equatorial Atlantic, the northward displacement of the ITCZ from the end of spring to the boreal summer entrains a northward shift of the surface wind anomalies at the equator. This in turn triggers the decline of the AEM event. In the tropical Pacific and Indian Ocean, the migration of the ITCZ is southward from the equator, and a positive event last for a much longer time [Richter et al., 2016].

1.2.2.3 Impact on the tropical and extratropical regions

Finally, the AEM variability can impact the climate of the other tropical and extratropical regions through modifications in the Walker and Hadley circulations, and wave dynamics. A Gill-Matsuno [Gill, 1980] type response to SSTAs in the eastern equatorial Atlantic could modulate or trigger ENSO events in the Pacific [Rodríguez-Fonseca et al., 2009]: a warm AEM phase leads to surface convergence, ascent, and upper-level divergence over the Atlantic Ocean, with two anticyclones on both sides of equator; over the central Pacific, a large-scale upper-level convergence leads to a downward motion and a surface divergence. This shoals the thermocline. The reduction in the thermocline depth propagates eastward throughout the season, together with the surface wind divergence. Easterlies are then accelerated and favor the development of cold water in the East. A La-Niña event can develop, through the positive Bjerknes feedback loop. Keenlyside et al. [2013], however, found that the equatorial Atlantic variability could modulate ENSO events rather than trigger them. Furthermore, through mechanisms such as extratropical Rossby wave response to upper-level divergence, a positive AEM

event could have an impact on the North Atlantic, Mediterranean Sea, and Europe climates [Haarsma and Hazeleger, 2007; García-Serrano et al., 2008, 2011; Losada et al., 2011; Losada and Rodríguez-Fonseca, 2015; Mohino and Losada, 2015]. The equatorial Atlantic SST mode of variability also impacts the Asian monsoon through changes in the circulation, following a Gill-type mechanism: the upper level divergence over the tropical Atlantic leads to Kelvin waves propagation which induce an upper level convergence and subsidence over India [Kucharski et al., 2007, 2008, 2009; Kucharski and Joshi, 2017].

1.2.3 The Guinean mode

The second mode of the 1901-2020 rainfall variability over the West Africa (EOF2) explains 24.1% of the total rainfall variance over the region (Fig. 1.13 (b)). Its center of action lies mainly over the Guinea Coast region where it is highly correlated with the rainfall anomalies. This mode is termed the Guinean mode. Much of the variability of the Guinea Coast rainfall (GCR) occurs on the interannual timescale, although there is a little multi-decadal modulation of the rainfall over the region (Fig. 1.14 (b)). In particular, the persistent dry conditions from the 1970s to 2000s is easily identifiable. The principal component time series (PC2) associated with the EOF2 is significantly correlated with the rainfall averaged over the Guinea Coast ($r = 0.97$).

1.2.3.1 SST decadal to multi-decadal forcing on the Guinea Coast rainfall

The SST anomalies pattern related to the low-frequency component of the PC2 displays some similitude with the SST anomalous pattern related to the low frequency component of the Sahelian rainfall mode principal component (Figs. 1.15 (e)-(f)). This suggests that similar oceanic forcings are at play on the multi-decadal timescale, in driving the low-frequency fluctuations of the rainfall over all the West Africa. Although there is no correlation between PC1 and PC2 raw values, their low-frequency values are significantly correlated ($r = 0.52$). Thus, during the entire 1901-2020 period, the colder (warmer) than normal sea surface conditions over the Indian Ocean and the tropical South Atlantic induce above (below) normal rainfall over the Guinea Coast. In the Northern Hemisphere, positive (negative) SST anomalies over the tropical North Atlantic (except a hole of non-significant anomalies in the center of the basin), the Mediterranean Sea and the Pacific result in an increase (decrease) in the rainfall over the Guinea Coast. Next, we focus the discussion on the oceanic forcing of the Guinea Coast rainfall variability on the interannual timescale.

1.2.3.2 Role of the Atlantic equatorial mode

The JJAS regression of the SSTAs onto the high-frequency component of the PC2 highlights an in-phase relationship between the eastern equatorial and South Atlantic SST (Fig. 1.15 (d)) and the rainfall over the Guinea Coast. Polo et al. [2008] found that the coupling between the two regions appears as the leading mode of covariability between the tropical SSTAs and the West African rainfall anomalies on the interannual timescale. The anomalous warmings and coolings in the eastern equatorial Atlantic are related to different phases of the AEM. During the 20th century, the interrelation between the AEM and the rainfall over the Guinea Coast appears more stationary than some of the links discussed above [Nnamchi and Li, 2011; Losada et al., 2012; Diatta and Fink, 2014]. Moreover, Fig. 1.16 indicate that over the 1901-2020 period, the rainfall anomalies occur more often when the eastern Atlantic is colder than normal than when it is warmer. This fact raises the question of asymmetry in the influence of the AEM on tropical climate, the GCR in particular.

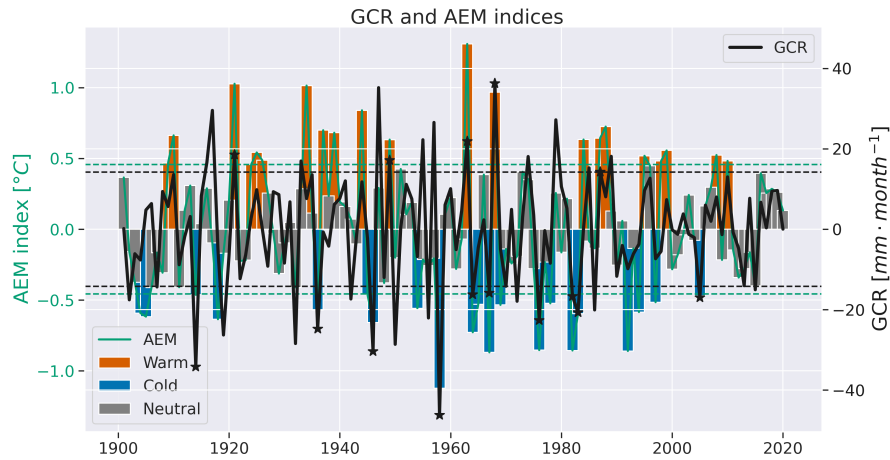


Figure 1.16: Timeseries of the AEM index (green curve) and the high frequency GCR index (HGCR) over the 1901-2020 period. The dashed green (black) lines denote $\pm$ one standard deviation of AEM (HGCR) index. Red (blue) bars indicate years when the AEM was in warm (cold) phase. Gray bars indicate neutral AEM phases. Years with above (below) normal HGCR coinciding with above (below) normal AEM index are marked with a star.

1.2.3.3 Role of the South Atlantic Ocean Dipole

Figure 1.15 (d) also indicates the development of negative SSTAs in the southwestern tropical Atlantic in relation with an increase in the rainfall over the Guinea Coast on the interannual timescale. These negative SSTAs together with the northeastern Atlantic positive SSTAs, form a pattern which resembles a positive phase of the South Atlantic Ocean Dipole (SAOD) [Nnamchi and Li, 2011; Nnamchi et al., 2011]. During positive phases of the SAOD, strong surface convergence and upward motions develop over the warmer than normal eastern equatorial Atlantic (termed the northeast pole, NEP). Over the colder than normal southwestern Atlantic (the southwest pole, SWP), off Argentina, Uruguay and Brazil, there is a surface divergence. Part of the surface divergent winds circulation above the SWP leads to northerlies which are deflected into westerlies at the equator and reinforces the convergence over the NEP. Moreover, there is an important moisture advection from the equator toward the Guinea Coast, which increases the rainfall over the region.

1.2.3.4 Role of the tropical Pacific

Over the tropical Pacific, a La-Niña like SST pattern, with temperatures lower than the climatological mean in the eastern equatorial Pacific, is concurrently related to positive rainfall anomalies over the Guinea Coast during the 1901-2020 period (Fig. 1.15 (d)). Interestingly, the interannual rainfall variability over the Guinea Coast during the whole 1901-2020 period involves an opposite phase relation between the eastern equatorial Atlantic and the eastern equatorial Pacific, whereas the variability over the Sahelian rainfall is associated with an in-phase relationship between both basins. Experiments conducted with atmospheric general circulation models (AGCMs) confirmed the reduction in the May-September rainfall anomalies over the Guinea Coast subsequent to abnormal warming in the eastern equatorial Pacific [Mohino et al., 2011b].

1.3 Three scientific questions

The previous sections showed that the mean climate of West Africa has been investigated in many studies. Changes in the atmospheric circulation, in the position of the heat low, and in the different jets are monitored every day by forecasters to provide meaningful information to the population, although there is still a lot of effort to be made to densify the region with observational stations. The role of SST in modifying the mean circulation and the stability of the atmosphere, or in influencing the transport of moist air from the adjacent or remote Indian and Pacific Oceans, has also been highlighted during the last decades. The majority of

these works were focused on the variability of the boreal summer rainfall in the Sahel [e.g. Rowell, 2001, 2003; Fontaine et al., 2011; Martin et al., 2014; Mohino et al., 2011a; Suárez-Moreno et al., 2018; Pomposi et al., 2020]. This season accounts for 85% of the Sahelian annual rainfall, and experiences large variability from interannual to multi-decadal timescales. The non-stationarity of the relationships between the Sahelian rainfall and the different oceanic basins has also been deeply explored (e.g. Suárez-Moreno et al. [2018]), as well as the impact of the oceanic forcings on the moderate and heavy rainfall events [Diakhaté et al., 2019, 2020]. However, few of these works highlighted the differences between the Sahelian and the Guinea Coast SST forcings [e.g. Giannini, 2003]. Moreover, the studies focused on the SST influence on the rainfall variability in Guinea Coast considered generally the boreal summer, which account for 60% of the annual rainfall over this region [e.g. Polo et al., 2008; Fontaine et al., 2009; Losada et al., 2010a, 2012; Rodríguez-Fonseca et al., 2011; Mohino et al., 2011b; Rodríguez-Fonseca et al., 2015; O'Reilly et al., 2017]. This means that there are still a lot of questions about the climate variability in the Guinea Coast in the other seasons, and on the potential role of oceanic internal modes of variability.

Climate models are mathematical representations of physical laws that govern interactions between different components of the climate system. These tools are useful to understand the past of the climate of the Earth as well as its future. An important question arises as to the performance of the most advanced General Circulation Models (GCMs) in representing the internal modes of climate variability, the impact of these modes on the climate of different regions (teleconnections) as well as their potential future evolution. From the third phase to the fifth phase of the Coupled Model Intercomparison Project (CMIP), Rowell [2013] found no improvement in the representation of the July-September (JAS) teleconnection from SSTAs to rainfall over different regions in Africa by the GCMs. Particularly, they showed that GCMs in the fifth CMIP phase (CMIP5) fail to simulate the link between the eastern equatorial Atlantic SST and the GCR. In parallel, the variability of SSTAs in the eastern equatorial Atlantic is well simulated by GCMs participating in CMIP5 and the following intercomparison project (CMIP6), despite the presence of warm (cold) biases in the eastern (western) tropical Atlantic Ocean [Richter et al., 2014; Richter and Tokinaga, 2020]. With these new versions of the GCMs, it will be interesting to evaluate the present and future connections between SST modes of variability and the GCR.

Another aspect of the rainfall characteristics in West Africa is the occurrence of extreme rainfall events. There is a variety of rainfall types in this region, according to their origins [Maranan et al., 2018]. At a synoptic scale, sea-breeze and land-breeze systems are known to impact the wind circulation, the cloud system formation and the rainfall over the Guinea Coast whatever the season [Cornforth et al., 2017]. Interactions between tropical and extra-tropical regions during the

post-monsoon and the harmattan, or even the pre-monsoon seasons, can also lead to extreme rainfall events over coastal areas of West Africa [Knippertz et al., 2003; Knippertz and Martin, 2005; Knippertz, 2007; Knippertz and Fink, 2008, 2009]. Moreover, slow propagating MCSs over the Guinea Coast are the source of heavy rainfalls over the region [Acheampong, 1982; Omotosho, 1985]. MCSs are rare during the harmattan season, and their number peaks during the first rainy season in the Guinea Coast (MAM) [Maranan et al., 2018, 2019]. The onset of the Guinea Coast rainy season in boreal spring generally coincides with the development of the cold tongue in the equatorial Atlantic, where the equatorial upwelling is enhanced. The surface atmospheric divergence over the cold sea favors southerlies, which transport moist air into the Guinea Coast area and provide good conditions for rainfall occurrences [Leduc-Leballeur et al., 2012; Meynadier et al., 2015; Tanguy et al., 2022]. Besides, statistical analyses indicate an increase in extreme rainfall events in relation with abnormal warming in the eastern equatorial Atlantic [Diatta et al., 2020]. Colder than normal conditions in the eastern equatorial Atlantic and warmer than normal conditions in the Mediterranean Sea have been in part associated with an extraordinary heavy rainfall that occurred on the 1st of September 2009 in Ouagadougou (located just North of the Guinea Coast), by providing good conditions for waves activities and MCSs propagation [Engel et al., 2017; Lafore et al., 2017; Beucher et al., 2019].

All these studies demonstrate an increase in the knowledge of the dynamics of the Guinea Coast system. There is a clear and well-known remote impact of SST anomalies in the equatorial Atlantic on the GCR in spring and summer, as well as on the occurrence of heavy rainfall events over the Guinea Coast. However, the influence of the different oceanic basins on the GCR as well and the stationarity of the linkages are not well known for the seasons beyond the monsoon season. Moreover, global warming affects the energy balance of the Earth system and modify its water cycle [Trenberth et al., 2003; Seager et al., 2010; Giorgi et al., 2019]. In a warmer future, changes in the precipitation mean and variability are expected in different regions of the world [van der Wiel and Bintanja, 2021]. Over the Guinea Coast, the rainfall variability in a warmer future climate is projected to increase from synoptic to interannual timescales, compared to the present-day situation [Akinsanola et al., 2020]. The projections also suggest an increase in the mean summer rainfall, in the rainfall intensity and in the maximum consecutive dry days as well as a decrease in the rainfall frequency [Klutse et al., 2018; Akinsanola and Zhou, 2019]. As oceanic conditions are important for the Guinea Coast rainfall variability, there is a need to quantify the impact of the oceanic forcings on the total rainfall and extreme rainfall indices over the Guinea Coast in the present-day climate and in a future warmer climate. All these observations bring us to shape the overarching research question of our thesis:

What is the role of oceanic internal modes of variability in driving the past, present and future Guinea Coast rainfall variability on the interannual timescale?

This question will be developed into three different questions:

- (i) On the interannual timescale, what are the main SST drivers of the Guinea Coast rainfall in different seasons of the year over the last century and the last decades?
- (ii) How do the current GCMs represent the boreal summer relationship between the SST in the eastern equatorial Atlantic and the Guinea Coast rainfall and what are the projected changes in this teleconnection?
- (iii) What are the present-day and future contributions of the AEM to the variability of the extreme rainfall events over the Guinea Coast during the Northern Hemisphere summer?

1.4 Three contributions

The responses to the general question of our thesis through the three sub-questions provide three different contributions to the understanding of the interannual rainfall variability over the Guinea Coast.

- (i) We have highlighted, on the interannual timescale, the strongest and most stationary links between oceanic internal modes of variability and the rainfall variability in the Guinea Coast, for four different seasons, during the 1901-2016 period. In addition, we explored the physical mechanisms at play (Chapter 2, published in *Climate Dynamics*). Assessing the non-stationarity in the SST-GCR relationships in the seasons other than the monsoon season is the novelty of this work, compared to the existing literature. Moreover, for the first time, we document in the same paper the statistical links between different oceanic areas and the Guinea Coast hydroclimate for different seasons, as well as the dynamics for the most striking relations found.
- (ii) The simulation of the AEM impact on the Guinea Coast rainfall by the GCMs participating in CMIP6 has been assessed, and the changes in this connection under future global warming have been thoroughly examined. We found a weakening of the AEM variability and its impact on the Guinea Coast rainfall in the Shared Socioeconomic pathway 5-8.5 (SSP5-8.5) climate scenario (Chapter 3, published in *Earth System Dynamics*). We have also found the processes by which some models which better simulate the AEM-GCR link differ from those who do not have the correct sign of the connection in the present-day climate. While in the literature the annual cycle of the rainfall in

the Sahel and Guinea Coast, as well as SST over the eastern equatorial Atlantic have been studied with the outputs from CMIP6 GCM simulations, here for the first time, we use these model outputs to study the teleconnection between the eastern equatorial Atlantic and the Guinea Coast.

- (iii) The present and future impacts of the AEM on the extreme rainfall events over the Guinea Coast have been evaluated for the first time within CMIP6 models simulations (Chapter 4, under revisions in *Weather and Climate Dynamics*). Studies in the literature have evaluated the changes in the mean and variability of extreme indices, while our analysis is oriented toward the influence of the AEM on the extreme rainfall occurrence over the Guinea Coast. Future changes in the contribution of the AEM to the extreme rainfall variability over the Guinea Coast have been assessed by using the CMIP6 historical and SSP5-8.5 simulations.

1.4.1 Interannual variability of rainfall in the Guinea Coast region and links with SST changes over the twentieth century for the different seasons (Chapter 2)

The novelty of the work in Chapter 2 consists in drawing a complete picture of teleconnection processes involving the observed rainfall over the Guinea Coast and different oceanic basins in seasons beyond the monsoon season. Our analysis is focused on the detection of the strongest interrelations between 21 different SST indices representative of the SST variability in different sectors of the world Ocean and the Guinea Coast rainfall index during the 1901-2016 period. We used observed rainfall and SST datasets for the diagnostics. We consider the harmattan season (December-February), the pre-monsoon season (March-May), the monsoon season (June-September) and the post-monsoon season (October-November). Different statistical tools involving regressions with different time-window length are applied to test the sensitivity of the connections to the starting year of the indices, as well as the length of the samples that are correlated with each other. We retain for each season the strongest connection found, and we use a reanalysis data to understand the physical processes at play.

Link with the next chapter The relation between the Atlantic equatorial mode of variability and the rainfall fluctuations over the Guinea Coast in the boreal summer appears as the strongest and most stationary over the 20th century, compared with the other SST-GCR relationships evaluated for the different seasons. In the next chapter, we will explore how current coupled Atmosphere-Ocean General Circulation Models (AOGCMs) represent the AEM influence on the rainfall over the equatorial Atlantic and the Guinea Coast under the present-day conditions, as well as the projected changes in this connection.

1.4.2 Weakened impact of the Atlantic equatorial mode on the future equatorial Atlantic and Guinea Coast rainfall (Chapter 3)

In Chapter 3, we evaluate the ability of the most advanced state-of-the-art AOGCMs participating in the CMIP6 to represent the boreal summer influence of the equatorial Atlantic mode on the Guinea Coast rainfall in the present-day climate and analyse the projected changes in this teleconnection in a future warmer climate. The present-day climate evaluation is performed with the historical simulations, whereas the future climate is based on the SSP5-8.5 simulations. We were able to:

- Understand the key physical processes that differentiate models which are able to better represent the AEM-GCR link from the others.
- Evaluate the changes in the Guinea Coast rainfall response to the AEM events.
- Determine the role of the changes in the mean state of the tropical Atlantic on the changes in the AEM, and subsequently the implication for the teleconnection to the Guinea Coast.

Link with the next chapter In this chapter, we found a reduction of the AEM variability in a future warmer climate due to a weakening of the Bjerknes feedback over the Atlantic Ocean. This leads to a reduction of the July-September rainfall variability associated with the AEM. Next, by using outputs from CMIP6 simulations, we will evaluate how the AEM impacts extreme rainfall events over the Guinea Coast in two different climates.

1.4.3 Future changes in the mean and variability of extreme rainfall indices over the Guinea Coast and role of the Atlantic equatorial mode (Chapter 4)

Chapter 4 evaluates the performance of the GCMs in simulating different extreme rainfall indices based on the definitions provided by the Expert Team on Climate Change Detection and Indices (ETCCDI). This analysis is not new, as many studies already tackled the subject [Klutse et al., 2021; Faye and Akinsanola, 2021]. This is also the case for the mean changes and the changes in the variability of the extreme indices over West Africa [Klutse et al., 2018; Akinsanola and Zhou, 2019; Akinsanola et al., 2020; Almazroui et al., 2020; Dosio et al., 2021]. The novelty of our study relies on the role of the oceanic internal modes of variability, mainly the AEM, in driving the changes in the extreme indices. We also show how the influence of the AEM on these extreme indices is modified in a warmer climate,

compared to their interrelation under the present-day climate conditions.

Finally, in Chapter 5, we summarize our main contributions to the understanding of the Guinea Coast hydroclimate variability by providing some answers to the general question raised above. We also discuss some perspectives to improve the results of our study and the understanding of some aspects of the teleconnections we have highlighted.

2

Interannual variability of rainfall in the Guinean Coast region and its links with sea surface temperature changes over the twentieth century for the different seasons

Abstract

The summer Guinean Coast (GC) rainfall (GCR) displays a strong variability on different timescales that are driven by Sea Surface Temperature (SST) variations and amplified by land-atmosphere processes. However, the relationships between the GCR and SST modes of variability in the pre-monsoon (March to May, MAM), post-monsoon (October to November, ON) and Harmattan (December to February, DJF) seasons are not well known nor understood. Using observational dataset covering the 20th century, we extend the conclusion obtained in previous studies that mainly analyzed the summer period (June – September, JJAS) by considering changes in SST-rainfall linkages throughout the year. We show that, in boreal winter, SST interannual variability in the tropical basins are anticorrelated with the GCR. The South Atlantic Ocean Dipole (SAOD) and the Atlantic Niño (ATL3) appear, however, as major drivers of the pre-monsoon and monsoon GCR. In MAM, both modes are in opposite phases with the GCR. Below normal SST in the tropical South Atlantic in MAM leads to a surface divergence south of the equator, and the resulting southerlies bring moist air into coastal Guinea, increasing rainfall. During JJAS, ATL3 and SAOD are in phase with the GCR. During ON, the eastern Mediterranean Sea anomalous warming strengthens the Saharan Heat Low, whose extension in the tropical North Atlantic enhances the low-level westerly Jet.

This chapter is based on: Worou, K., Goosse, H., Fichet, T., Guichard, F., and Diakhaté, M. (2020). Interannual variability of rainfall in the Guinean Coast region and its links with sea surface temperature changes over the twentieth century for the different seasons. *Climate Dynamics*, **55**, 449–470, <https://doi.org/10.1007/s00382-020-05276-5>. Supplementary materials are in Appendix A.

This jet transports moisture into GC. The stationarity of the correlations between the GCR and SST indices has also been assessed, and the strongest and most stationary links are obtained during the monsoon season.

2.1 Introduction

In mid-2019, the estimated western African population was about 391 440 000 [United Nations Department of Economic and Social Affairs/Population Division, 2019], which is more than five times the population recorded in the mid-1950s. This rapid growth of population causes an important pressure on water availability in the region. Agriculture, livestock and fishing are the lungs of the West African economy, which makes the West African population vulnerable to any change in the rainfall regime. Over the 20th century, the boreal summer SST variability in the eastern equatorial Atlantic has strongly impacted the GCR which is the main focus of this study. The covariability between ATL3 and GCR has been shown to be stationary [Diatta and Fink, 2014]. On the other hand, it has been shown that the relation between the rainfall in the coastal regions and the Sahel is not stationary, due to changes in the connection between the tropical Atlantic and Pacific [Losada et al., 2012]. The vulnerability of the Sahelian area due to the strong multi-decadal variability of its rainfall and the associated social-economic consequences prompted studies that aim to identify the drivers of such variability. Indeed, this area suffered from a long drought between the 1970s and 1980s, after wet conditions in the 1950s and 1960s [Hulme, 1992]. As briefly summarized below, our current knowledge about rainfall variability over West Africa comes from studies which focused on the Sahel. Part of the changes in the rainfall regime in this region has been attributed to different forcings, in particular the remote influence of regional and world oceans known as teleconnections, through their influence on atmospheric circulation and moisture content [Losada et al., 2009; Fontaine et al., 2009; Janicot et al., 2011; Mohino et al., 2011a; Rodríguez-Fonseca et al., 2011; Rowell, 2013; Rodríguez-Fonseca et al., 2015; Villamayor et al., 2018].

On interannual timescales, the observed summer rainfall fluctuations in the Sahel are driven by SST changes at the global scale [Folland et al., 1986; Giannini, 2003], with a competition between tropical and extratropical oceanic basins. The remote effects of the different oceanic areas have been drawn out in many studies. It has been demonstrated that the Mediterranean Sea plays an important role in modulating the rainfall fluctuation in the Sahel on interannual timescales. Years of positive summer Mediterranean SST anomalies are associated with positive summer rainfall anomalies in the Sahel and inversely [Rowell, 2003]. In observational climate data, there is a symmetric interrelation between warm and cold situations of the Mediterranean Sea and their respectively associated higher and lower than average rainfall in Sahel. However, sensitivity experiments conducted with Atmo-

spheric Ocean General Circulation Models (AOGCMs) show only an increased Sahelian rainfall associated with positive SST anomalies of the eastern Mediterranean basin [Fontaine et al., 2009]. In addition, it has been demonstrated that the Mediterranean SST changes affect the daily moderate, heavy and extreme rainfall in the Sahel [Diakhaté et al., 2019]. Other recent sensitivity experiments have shown that, on the one hand, among the worldwide oceans, the Mediterranean basin is the main driver of the recent wet conditions over the Sahel in the boreal summer. On the other hand, this basin should be determinant for the future Sahelian rainfall [Park et al., 2016]. All those studies agree that above normal SST conditions in the Mediterranean Sea give rise to moist air over the basin which is then advected southward into the Sahel. This escalates the low-level convergence in Sahel and increases rainfall over the area, together with various positive feedback processes [Jung et al., 2006]. It has also been shown that the western Mediterranean SST changes drive rainfall variations in the Guinean Coast via an enhancement of deep convection in the eastern equatorial Atlantic [Fontaine et al., 2009].

In the North Atlantic, numerous studies pointed out the major influence of the Atlantic Multidecadal Oscillation (AMO) on the seasonal, interannual and decadal to multidecadal rainfall variability in West Africa, especially in the Sahel. The AMO-Sahelian rainfall link has been shown to be very strong on decadal-multidecadal timescales [Berntell et al., 2018]. During AMO positive phases, the strengthening of the shallow meridional circulation (SMC) associated with the Saharan Heat Low (SHL) allows a southerly moisture flux which increases spring rainfall in the Sahel [Martin and Thorncroft, 2014]. In summer, in addition to a strengthening of the SMC, positive AMO phases induce above normal SST conditions in the Mediterranean Sea which in turn are favorable to more moisture convergence in the Sahel leading to more rainfall. Thus, AMO positive (negative) phases are associated with positive (negative) rainfall anomalies in the Sahel [Knight et al., 2006; Rodríguez-Fonseca et al., 2011; Martin and Thorncroft, 2014; Rodríguez-Fonseca et al., 2015; Schubert et al., 2016].

The leading mode of covariability between the West African rainfall and the tropical Atlantic SST links the summer rainfall in the Gulf of Guinea with SST changes in the eastern equatorial Atlantic [Polo et al., 2008]. These SST changes associated with this mode are referred to as the Atlantic equatorial mode (AEM) or Atlantic Niño (ATL3). The Atlantic Niño also appears as the leading SST mode of variability in the tropical Atlantic [Dommenget and Latif, 2000, 2002; Keenlyside and Latif, 2007; García-Serrano et al., 2008; Richter et al., 2012; Mohino and Losada, 2015]. Basically, warm phases of the sea surface temperature anomalies (SSTAs) in the ATL3 area lead to a dipole pattern of rainfall in West Africa, with positive rainfall anomalies along the Coast of Guinea and negative rainfall anomalies in Sahel. The linkages between the Atlantic Niño and the West African rainfall

are not linear, as underlined by model sensitivity experiments. The positive and negative phases of the Atlantic equatorial mode induce, indeed, changes in rainfall of different magnitudes [Losada et al., 2010a]. Over the 20th century, the summer ATL3 - GCR relationship appears stationary [Diatta and Fink, 2014].

In the South Atlantic Ocean (SAO), a dipole defined as the difference between SSTAs in the North East Pole (NEP, an area in the ATL3 domain) and the South West Pole (SWP, an area along the Argentina-Uruguay-Brazil) is termed the South Atlantic Ocean Dipole (SAOD). On interannual timescales positive phases of the SAOD are strongly related to positive summer rainfall anomalies in the Guinean Coast and inversely [Nnamchi and Li, 2011]. The connection of this coupled ocean-atmosphere phenomenon with the Guinean Coast operates via the Lindzen-Nigam process [Lindzen and Nigam, 1987]: during positive phases of the SAOD, large subsidence and divergence occur in the SWP, and are related to strong convergence and upward motion of humid air over the NEP. This leads to a flux of moisture from the Gulf of Guinea and equatorial Atlantic into areas in the Coast of Guinea. Note that the SAOD positive phases are associated with a reduction of the Indian monsoon rainfall [Kucharski and Joshi, 2017].

Remote effects of the Pacific Ocean have been linked with changes of the Sahelian rainfall in boreal summer: in general, above normal SST anomalies in the central and eastern Pacific or SST anomaly gradient from the western Pacific to the eastern Indian Ocean are associated with negative rainfall anomalies in the Sahel [Rowell, 2001]. The connection between the Pacific Ocean and West Africa under those particular SST conditions operate via the propagation of Kelvin (eastward) and Rossby (westward) waves. This enhances subsidence over the Sahel region. In addition, since the 1970s, from May to September, the dry belt induced by the positive SSTAs in the Pacific Ocean migrates progressively from the Guinean Coast to the Sahel [Mohino et al., 2011b]. In the northern Pacific, an opposite phase relationship has been established between the Interdecadal Pacific Oscillation and Sahelian rainfall in boreal summer on decadal to multidecadal timescales [Mohino et al., 2011a].

The Indian Ocean exhibits an opposite phase relationship with the Sahelian rainfall on decadal timescales [Rowell, 2001; Bader and Latif, 2003]. Indeed, the Sahelian wet conditions in the 1950s - 60s are linked with negative SST anomalies in the Indian Ocean, while the warming trend of this latter in the 1970s could be associated with the severe 1970s Sahelian long drought.

On smaller, regional scale, some features of the atmospheric circulation also play important roles in the rainfall genesis over West Africa. One key feature is the presence of different veins of wind which combine with the low-level moisture flow, such as the mid-level African Easterly Jet (AEJ) at 650 hPa [Cook, 1999; Hsieh and Cook, 2005]) and the high-level Tropical Easterly Jet (TEJ, 200 hPa). Their positions, intensities as well as the activity of waves induced by e.g. instabil-

ities in their core modulate the frequency and intensity of rainfall [Thorncroft and Hoskins, 1994a,b; Thorncroft and Blackburn, 1999; Nicholson and Grist, 2003; Gu and Adler, 2004; Nicholson, 2009, 2013; Sylla et al., 2011; Kebe et al., 2016]. The African Easterly Jet takes its origin in the meridional temperature gradient between the Saharan desert and the Atlantic Ocean, while the Tropical Easterly Jet results from an intense north-south temperature gradient between the Himalayan Plateau and the Indian Ocean. Additionally, the presence of the Saharan Heat Low (SHL) and its associated Shallow Meridional Circulation (SMC) have been identified as a major driver of the Sahelian rainfall [Hagos and Zhang, 2009; Lavaysse et al., 2015; Shekhar and Boos, 2017]. The Saharan Heat Low forms as a result of high temperature over the Saharan desert. Its presence induces a shallow meridional circulation: northward winds from the Sahel converge at the core of the SHL, leading to an ascent followed by a divergence at 700 hPa and a southward displacement of the wind at that level. Those studies update the former view of the West African rainfall variability which has been thought to be controlled by the seasonal cycle of the Intertropical Convergence Zone (ITCZ), where at the surface, hot and dry northeasterly Harmattan winds blow over the Sahara and meet cold and moist southwesterly winds that originate from the Atlantic Ocean. Ascending motion of air then leads to cloud systems and the associated rainfall [Germain, 1968; Detwiler, 1965; Walker, 1957; Hamilton et al., 1945; Sansom, 1965]. In addition, over the West African continent, the term ITCZ is confusing and sometimes, the Intertropical Discontinuity and Intertropical Front are used instead, and correspond to the place where dry northeasterlies meet southwesterlies (monsoon flow).

The different studies presented above attest of the great progress accomplished in the identification and understanding of the role of the oceanic forcing and other synoptic systems on rainfall variability in West Africa. Most of those works are, however, much focused on the Sahelian area and the boreal summer, which account for about 85% of the Sahelian annual rainfall. The number of studies devoted to the Guinean Coast is much smaller, and the dynamics over this region is quite distinct. Indeed, spring rainfall in the Guinean Coast is controlled by the development of a cold tongue in the equatorial Atlantic [Nguyen et al., 2011; Meynadier et al., 2015]; this cold tongue can also influence the rainfall onset of the Sahel without being the main driver [Caniaux et al., 2011; Steinig et al., 2018]. Moreover, those studies investigated the SST-rainfall relationship in the Guinean Coast during the summer period. However, 40% of the annual rainfall falls in this area from October to May (against 15% in the Sahel). This motivates the present study which analyzes the oceanic influence on the interannual Guinean Coast rainfall for all the seasons over the 20th century and the last decades. The stationarity of the relationships are then analyzed, and the major SST indices are pointed out.

2.2 Data and methods

The present study focuses on the 1901 - 2016 period. The monthly rainfall product comes from CRU TS v4.01 (Climate Research Unit Time Series). It is a gridded rainfall data which covers only lands at a resolution of 0.5° and which spans from 1901 to 2016 [Harris et al., 2013; University of East Anglia Climatic Research Unit (CRU) et al., 2017]. The monthly SST data selected for the analyses are derived from the version 1 of the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST1) [Rayner et al., 2003], which extends from 1870 to 2017 at a resolution of 1° latitude by 1° longitude.

Four different seasons have been considered: the Harmattan season or DJF (December, January, February), the pre-monsoon season or MAM (March, April, May), the monsoon season or JJAS (June, July, August, September) and the post-monsoon season or ON (October, November). The definition of DJF and MAM are consistent with Zhang et al. [2006]; Abiodun et al. [2012]; Senghor et al. [2017], while JJAS is consistent with Rodríguez-Fonseca et al. [2011]; Diatta and Fink [2014].

Monthly SST and rainfall anomalies are first computed from the linearly detrended SST and rainfall data. Then, monthly SST and rainfall indices are computed as area-averaged anomalies over different regions, based on the literature and defined in Table 2.1. Particularly, for West Africa, two rainfall indices are computed for areas lying between $5^\circ - 10^\circ\text{N}$ and $20^\circ\text{W} - 10^\circ\text{E}$. This classic delimitation is based, for example, on the center of action of the second mode of summer rainfall variability in West Africa. The leading mode has its main center of action on the Sahel. Those modes are obtained by applying an Empirical Orthogonal Functions (EOFs) analysis [Monahan et al., 2009; Dommenges and Latif, 2002; Wilks, 2011] on rainfall anomalies over West Africa. For the North Pacific SST index (NPAC), the monthly SST indices are defined as the standardized expansion coefficients of the leading EOF mode of variability of the northern Pacific Ocean. Besides, seasonal SST and rainfall indices are derived from the monthly indices by their averages over months defining each season.

In addition, as much of the rainfall variability of Guinea Coast occurs on interannual timescales [Diatta and Fink, 2014], the present study focuses on that time period. This is achieved by first applying a low pass Lanczos filter [Duchon, 1979] to all the time series of each seasonal index, with a cutoff of 8 years. The main results do not change if a cutoff of 10 years is chosen. This provides low frequency changes which are secondly removed from the initial seasonal index. The remaining time series are here considered as the high-frequency interannual timescale index. For the Guinean Coast rainfall indices, filtered and unfiltered time series are also compared. Finally, the stationarity of the covariability between the rainfall and oceanic modes of variability has been assessed by performing a run-

ning correlation, with different lengths of the sliding window and different starting years. The significance level of the Pearson correlation coefficients calculated in this work has been set to 95% confidence level according to a Fisher-test. Following Taubenheim [1969], the significance of a linear correlation r between two time series $x(t)$ and $y(t)$ of size N can be assessed by computing a critical value of the F parameter defined as:

$$F_{cri} = \frac{r^2}{1 - r^2} (N - 2) \quad (2.1)$$

where $N - 2$ is the standard degree of freedom ($SDOF$). To take into account autocorrelation in the time series, an effective degree of freedom ($EDOF$) is computed [Taubenheim, 1974; Fink and Speth, 1997] :

$$EDOF = 1 + \frac{N}{1 + 2 \sum_{l=1}^{N-1} \frac{N-l}{N} K_x(l) K_y(l)} \quad (2.2)$$

where $K_x(l)$ and $K_y(l)$ are the autocorrelation functions at lag l . Then, from the F-distribution with $d_1 = 1$ and $d_2 = SDOF$ or $d_2 = EDOF$ degree of freedom (d_1 and d_2 are the degrees of freedom of the Fisher's law), the probability (p_{value}) to exceed the F_{cri} is computed. Therefore, r values are significant if $p_{value} \leq 0.05$. The results of the two tests are shown to highlight differences if our series are autocorrelated.

Finally, we used ERA-20C [Poli et al., 2016], a reanalysis product from the European Centre for Medium-Range Weather Forecasts to highlight the physical key mechanisms associated to the strongest connections that are obtained. It is available for the 1900 - 2010 period. For each season, SST index of the main oceanic mode of variability is calculated from the reanalysis data. Then, this index is standardized and regressed onto various atmospheric fields to underline the mechanisms by which the oceanic mode influences the rainfall in the coastal Guinea. Similarly, circulation fields associated with each seasonal rainfall index is available in the Appendix A. Although we have selected standard observational datasets used in many applications (e.g. Rowell [2003]; Gómara et al. [2017] for HADISST data, and Diatta and Fink [2014]; Bamba et al. [2015]; Salih et al. [2018]; Okoro et al. [2019] for CRU data), including in West Africa, their validation is beyond the scope of this work. One must keep in mind that some limitations of our results may come from uncertainties in these datasets (as well as biases in the reanalysis products) which are the object of several ongoing studies (e.g. Hirahara et al. [2014]; Huang et al. [2016]).

Table 2.1: Summary of the various indices computed for the analyses. Results for indices which names are tilted are not shown in these analyses. Latitudes and longitudes are in °. Some indices are defined by the following link: <https://stateoftheocean.osmc.noaa.gov/sur/>. Additionally, indices defining El-Niño Modoki follow Li et al. [2010].

	Index definition	Lat min	Lat max	Lon min	Lon max	Method
GC	Guinean Coast	5	10	-20	10	Area average
SH	Sahel	10	20	-20	10	" "
NATL	North Atlantic SST	0	70	-75	-10	" "
MED	Mediterranean Sea	30	40	-5	35	" "
MEDE	Eastern Mediterranean Sea	30	40	20	35	" "
MEDW	Western Mediterranean Sea	30	40	-5	20	" "
NAT	North Atlantic Tropical SST	5	20	-40	-20	" "
SAT	South Atlantic Tropical SST	-20	-5	-15	5	" "
TASI	Tropical Atlantic SST meridional gradient	-	-	-	-	NAT - SAT
TNA	Tropical Northern Atlantic	5	25	-55	-15	Area average
TSA	Tropical Southern Atlantic	-20	0	-30	10	" "
ATL3	Atlantic Niño	-3	3	-20	0	" "
NEP	North East Pole	-15	0	-20	10	" "
SWP	South West Pole	-40	-25	-40	-10	" "
SAOD	South Atlantic Ocean Dipole	-	-	-	-	NEP - SWP
NINO3	Niño3	-5	5	-150	-90	Area average
NINO34	Niño3.4	-5	5	-170	-120	" "
NINO4	Niño4	-5	5	160	-150	" "
TA	Region A	-10	10	165	-140	" "
TB	Region B	-15	5	-110	-70	" "
TC	Region C	-10	20	125	145	" "
EMI	El-Niño Modoki	-	-	-	-	1.0.TA - 0.5.TB - 0.5.TC
IEMI	Improved El-Niño Modoki	-	-	-	-	3.0.TA - 2.0.TB - 1.0.TC
NPAC	North Pacific	20	70	120	-105	EOF 1
WTIO	West Tropical Indian Ocean SST	-10	10	50	70	" "
SETIO	Southeastern Tropical Indian Ocean SST	-10	0	90	110	" "
DMI	Dipole Mode (Indian Ocean)	-	-	-	-	WETIO - SETIO
CINDO	Center of Indian Ocean	-25	10	55	95	Area average
IPWP	Indo-Pacific Warm Pool	-25	25	40	-140	" "

2.3 West Africa: mean climate, annual cycle and variability

The West African rainfall is dominated by a monsoon system which is responsible for more than 58% of the annual rainfall in the Guinean Coast and 84% in the Sahel (the contribution of each season for the total annual rainfall of the Sahel and Guinean Coast can be found in Table 2.2). During the Harmattan season (DJF), when northeasterlies are strong over the region, weak rain falls over West Africa (Fig. 2.1 (a)). More intense rainfall occurs in March in coastal areas. During the MAM season, the rainfall maximum lies between 4 and 9°N. In June, the rainfall reaches a maximum mean of about $8 \text{ mm} \cdot \text{day}^{-1}$ over the Guinean Coast (Fig. 2.1,(b)). Then, the rainbelt moves northward in the Sahel, where the mean rainfall rate peaks to about $6 \text{ mm} \cdot \text{day}^{-1}$ in August. At the same time, rainfall slightly decreases over the Guinean Coast. This corresponds to the "little dry season" [Adejuwon and Odekunle, 2006; Froidurot and Diedhiou, 2017; Fink et al., 2017]. In September, the rainfall maximum moves back equatorward, which implies a second maximum over the Coast of Guinea, with a mean intensity of more than $8 \text{ mm} \cdot \text{day}^{-1}$. This corresponds to the highest monthly fraction of the annual rainfall in the Guinean Coast (16%, Fig. 2.1,(b)). Consequently, the Guinean Coast and the Sahel rainfall annual cycles exhibit two different patterns, with respectively a bimodal and a unimodal structure (Fig. 2.1).

A change of the rainfall annual cycle in the Guinean Coast has been observed during the last century. Regarding the 1901-2016 period, the annual cycle is characterized by a bimodal structure, with two maxima, in June and September respectively. When focusing on the four consecutive 20-year periods prior to 1980, this bimodal annual cycle is well defined. However, after 1980, the bimodality becomes less pronounced, it even vanishes over 1981-2000, with a single maximum in August. Then it shifts to September for the 2001 – 2016 period (Fig. 2.2 (a)). The analysis of the monthly Guinean Coast rainfall variance shows no substantial change in each month, except July, August and September (Fig. 2.2 (b)). However, the rainfall variance has changed for the other months. For example, low variance is observed over the 1921 – 1940 period, for July, August and September. Then, it increases over 1941 – 1960. For the following period, the variance decreases for August and September while it continues to grow in July. In the 1981 – 2000 period, the variance decreases dramatically for July, while it slightly increases for August and September. Then, an increasing trend is observed for June, July and September in the 2001-2016 period relative to the 1981 – 2000. Otherwise, for August, the trend is decreasing for those last two periods. These changes are reflected in the seasonal rainfall indices of the coast of Guinea (Fig. 2.3 (i-l)) which will be analyzed rather than monthly rainfall indices.

Figure 2.3 (a-d) shows that each Guinean Coast rainfall index exhibits a negative trend for the 1901-2016 period. This trend is more pronounced for the pre-

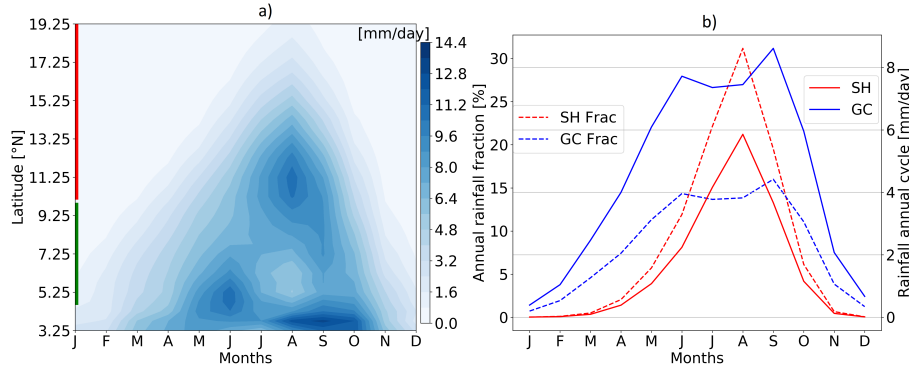


Figure 2.1: (a) Time-latitude evolution of rainfall mean over 20°W and 10°E. The green and red lines delineate latitudes corresponding to the Guinean Coast and Sahel regions respectively. (b) Annual cycle of rainfall mean (full lines) and annual rainfall fraction (dotted lines) for the Guinean Coast (in blue) and the Sahel (in red). Analyses are performed for the 1901-2016 period.

monsoon and post-monsoon seasons, with a slope of -0.005 and $-0.004 \text{ mm d}^{-1} \text{ yr}^{-1}$ respectively. For the DJF and JJAS seasons, the value of the slope is reduced to $-0.002 \text{ mm d}^{-1} \text{ yr}^{-1}$. The maximum rainfall variance of the Guinean Coast rainfall has been obtained in the monsoon season (JJAS), with a standard deviation of $0.85 \text{ mm} \cdot \text{day}^{-1}$, followed by the ON, MAM and DJF seasons. The values obtained for these three seasons are 0.75 , 0.5 and $0.22 \text{ mm} \cdot \text{day}^{-1}$, respectively. Similar results are found for the variance of the high pass filtered indices (Fig. 2.3 (e-f)). This shows that much of the GC rainfall variability happens on interannual timescales. In addition, the 20-year sliding window of the standard deviation of the rainfall is illustrated in Fig. 2.3 (i-l). The variances obtained are normalized to the 1901-2016 period. Results show for DJF an increase of the rainfall variance from the 1950s to the 1970s. For MAM, the rainfall variance decreases in the first two decades, then it remains constant until the 1960s before increasing. From the 1970s, a decreasing variance is observed. For JJAS, a decreasing trend of the variance is observed over the 1940 - 2000 period. Then, it increases afterwards. ON rainfall also shows the same decreasing trend until 1975s, from which the variance remains constant. For JJAS and ON, the rainfall variance increases over the 1925 - 1950 period. The pronounced decrease of the rainfall variance in the Guinean Coast (particularly in the boreal summer) has coincided with a reduction of the variability of the Atlantic cold tongue [Tokinaga and Xie, 2011]. This is due to an enhanced warming of SST in the Atlantic cold tongue area, which has decreased the trade winds intensity and lead to a weakening of upwelling in the eastern equatorial Atlantic. Consequently, the thermocline has deepened, reduc-

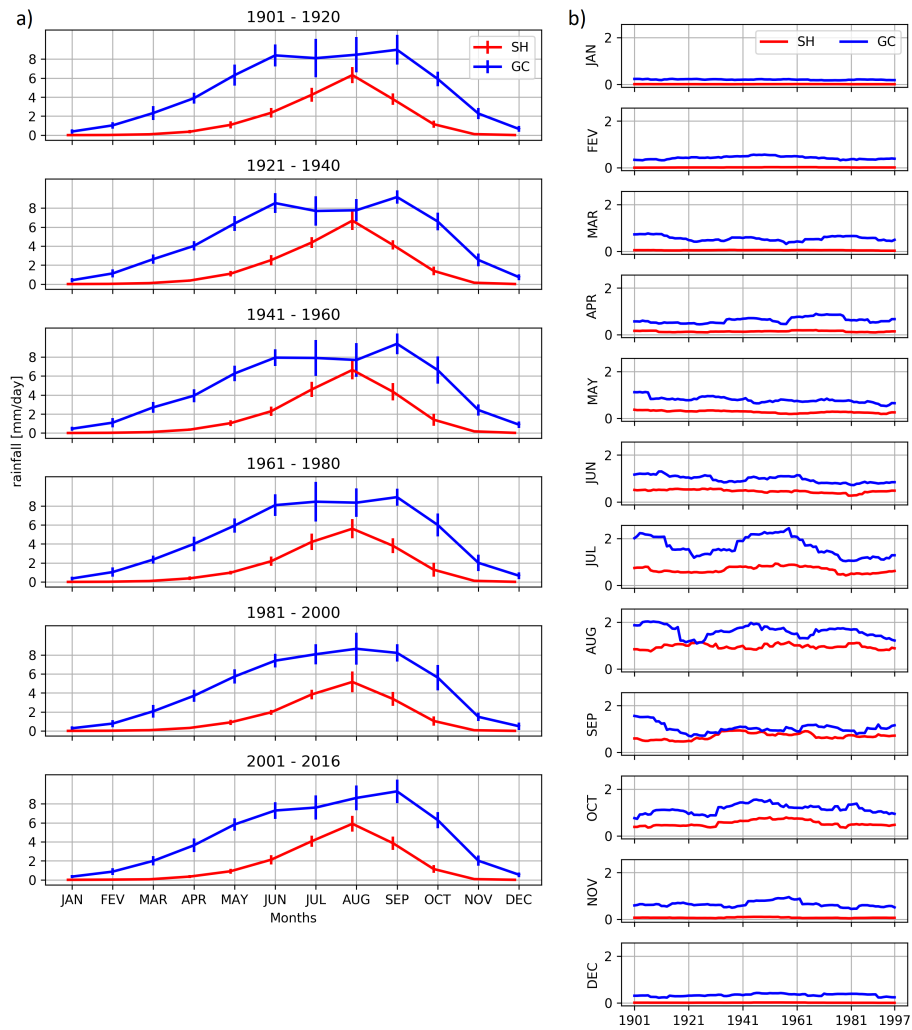


Figure 2.2: (a) Annual cycle of rainfall in the coast of Guinea (in blue) and Sahel (in red) over different periods of 20 years (the last period counts 16 years). (b) Standard deviations of the mean Guinean Coast (in blue) and Sahelian (in red) rainfall over sliding 20-year periods.

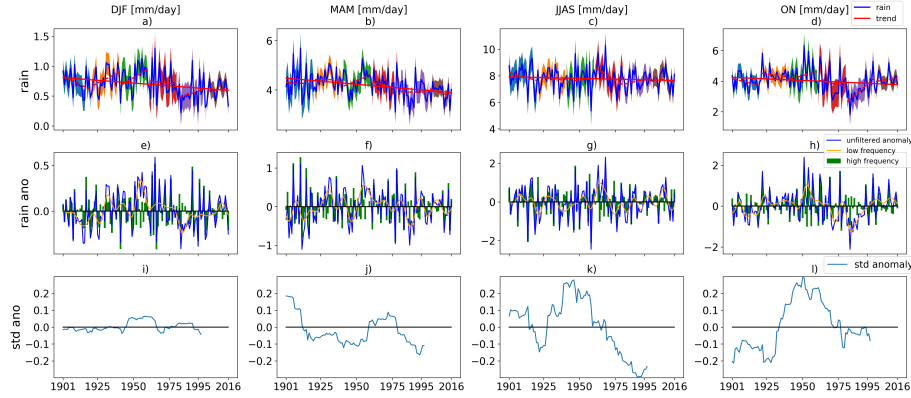


Figure 2.3: (a,b,c,d) Seasonal unfiltered rainfall indices and trends for the Guinean Coast . The trend over the 1901 - 2016 period is indicated by the red lines. Red dashes (shades) show the trend (standard deviation) over periods of at least 20 years from 1901. (e,f,g,h) Low (orange curves) and high (green bars) frequency components of the unfiltered seasonal rainfall anomalies of the coast of Guinea (blue bars). (i,j,k,l) Anomalies of the sliding 20-years standard deviation of the seasonal high-pass Guinean Coast rainfall index relative to its full value for 1901 - 2016. Indices are computed for the period 1901 - 2016.

ing the cold tongue and the variability of the Atlantic equatorial mode as well as the rainfall in the coastal Guinea.

Hereafter the influence of SST modes of variability on the Guinean Coast rainfall regime will be evaluated separately for the four seasons on interannual timescales. Figure 2.4 (a-d) presents the spatial distribution of the mean rainfall over West Africa and its evolution through the seasons. It shows an increase of the rainfall mean from the Harmattan season (DJF) to the pre-monsoon season (MAM), mainly in the Guinean Coast. Then the rainbelt moves northward in JJAS, with the highest values of rainfall in the Guinean Highlands and Cameroon mountains. In the post-monsoon season, the maximum values are located in the Guinean Coast. Figure 2.4 (e-h) shows that the core of the Guinean Coast rainfall correlation pattern (correlation between the seasonal Guinean Coast rainfall index and the West African rainfall anomalies) is located in the Guinean Coast latitudinal band in DJF and JJAS. As shown in Fig. 2.4 (a), weak rain falls in the Sahel in DJF, explaining the southward location of the correlation pattern. For JJAS, differences in the oceanic drivers of the West African rainfall could explain the main coastal core of the Guinean Coast rainfall correlation pattern. In particular, the disappearance of the West African Westerly Jet (WAWJ) in warm phases of the eastern equatorial Atlantic could reduce the zonal moisture transport onto the Sahel [Lélé and Leslie, 2016]. In MAM and ON, the Guinean Coast rainfall correlation pattern

extends over whole West Africa. The corresponding SST patterns are displayed in Fig. 2.4 (i-l). This figure shows high correlation values over large areas during the Harmattan and monsoon seasons, while in the pre-monsoon and post-monsoon seasons, the correlation values are low, and confined in few oceanic areas. In addition, a reversal of the sign of the correlation is observed between the DJF and JJAS seasons in the tropical South Atlantic and areas in the eastern tropical Indian Ocean and Maritime Continent (oceanic area between the eastern Indian Ocean and the western tropical Pacific Ocean). Those observations show that, throughout the seasons, the interrelation between the Guinean Coast rainfall and SST changes, and needs to be evaluated separately.

Table 2.2: Annual rainfall fraction (in %) for each season and each area (Sahel and Guinean Coast). Fractions are computed for the 1901-2016 period, as a seasonal cumulative mean rainfall over the cumulative annual rainfall. Those values are computed from CRU TS v4.01 rainfall data [University of East Anglia Climatic Research Unit (CRU) et al., 2017].

	DJF	MAM	JJAS	ON
Sahel (20° W - 10° E, 10° - 20° N)	0.2	8.3	84.7	6.8
Guinean Coast (20° W - 10° E, 5° - 10° N)	3.9	23.3	57.9	14.9

2.4 Oceanic forcing of the Guinean Coast rainfall variability: dominant SST modes of variability and robustness of the linkages

2.4.1 Oceanic forcing of the Harmattan (DJF) Guinean Coast rainfall

The mean DJF rainfall in West Africa over the 1901-2016 period ranges between 0 and 3 mm · day⁻¹, depending on the location (Fig. 2.4 (a)). Rainfall during this season corresponds to about 4% of the annual rainfall over the Guinean Coast (Table 2.2). Although this value is low compared to the other seasons, the DJF Guinean Coast rainfall plays an important role, as meteorological conditions of that season are crucial for the growing of agriculture products such as coffee and cocoa [Ehounou et al., 2019a,b; Lahive et al., 2018]. Studies show that this growth requires an annual rainfall between 1200 and 1500 mm [Boni, 1978; Lachenaud, 1987, 1991; Brou et al., 2003; Kassin et al., 2008; Köhler et al., 2010; Saj et al., 2017]. Moreover, three consecutive months of dry conditions could be dramatic for the growth of cocoa, which is why it is capital to monitor and understand the drivers of DJF rainfall variability, as this season is the driest one of the year. The rainfall correlation pattern associated with the interannual Guinean Coast index

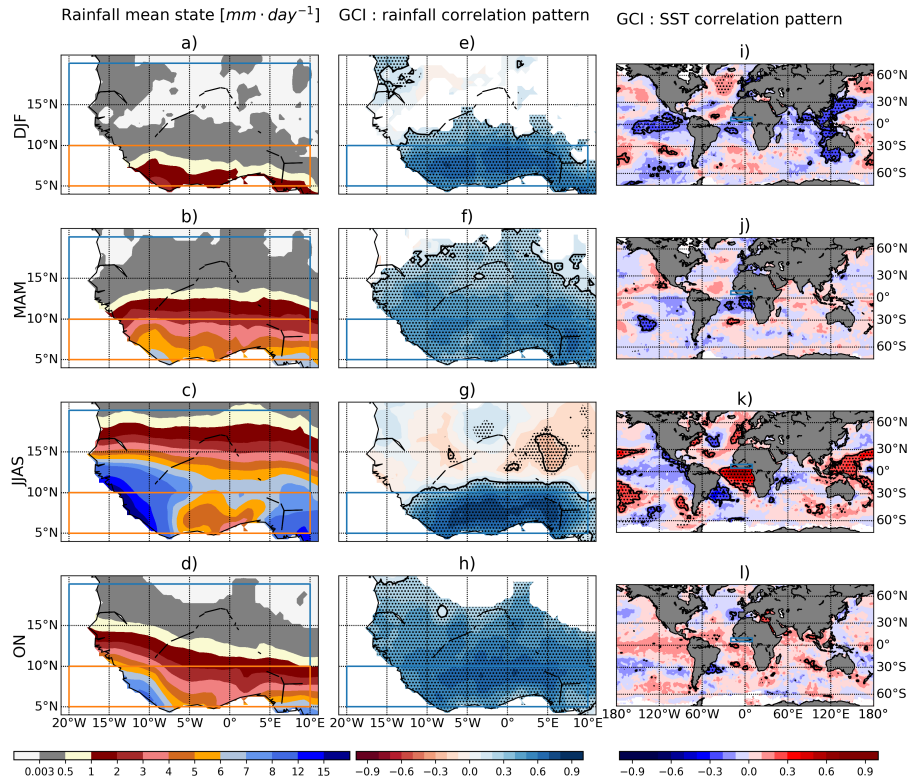


Figure 2.4: (a,b,c,d) Seasonal mean of rainfall rates over West Africa. (e,f,g,h) Correlation between the high-frequency Guinean Coast rainfall index and the seasonal rainfall anomaly at each grid point. (i,j,k,l) Correlation between the high-frequency Guinean Coast rainfall index and SST anomalies for each season. Contour lines and dots in the correlation patterns represent significant values at 95% confidence level according to a *F*-test with respectively a standard degree of freedom and the effective degree of freedom. Analysis is performed for the 1901 - 2016 period.

has a core of positive values which principally lies between 5 and 10°N (Fig. 2.4 (e)). To determine the oceanic areas that are related to this rainfall mode, the DJF Guinean Coast rainfall index has been correlated with global SSTAs of the same season (Fig. 2.4 (i)). The results show that above normal rainfall anomalies in the coast of Guinea are associated with strong negative SSTAs in the eastern tropical Pacific, eastern (western) boundaries of the tropical Indian (Pacific) Ocean. Few oceanic areas in the tropical South Atlantic and western Mediterranean Sea present negative correlations, while positive correlations are obtained in the Red Sea, eastern Mediterranean Sea, extratropical North Atlantic and tropical North Atlantic.

In a second step, the role of different SST modes of variability on the West African rainfall in DJF has been evaluated. The correlation between each SST index and the Guinean Coast rainfall for all seasons is displayed in Fig. 2.5. However, the discussion will be focused on the main SST modes whose results are summarized in Fig. 2.7. Those oceanic modes have the strongest impact (highest correlation and best stationarity feature) on the Guinean Coast rainfall. In agreement with the DJF large-scale SST pattern in Fig 2.4 (i), the El-Niño (Niño3), Indo-Pacific warm pool (IPWP, defined as the oceanic area within 40°E – 140°W, 25°S – 25°N, where the SST is above 28°C all the year, Weller et al. [2016]; Deckker [2016]) and Atlantic Niño (ATL3) indices are negatively correlated with rainfall anomalies in the Coast of Guinea, while the eastern Mediterranean SST index shows overall positive correlations (Fig. 2.7 (a-e)). The influence of the Atlantic Niño is located in the western area of the Guinean Coast, while that of IPWP is more pronounced in the northeastern boundary of Guinea Coast, as well as in the southwestern coast. Relative to the IPWP SST index, indices of El-Niño have few significant correlations with the rainfall anomalies in West Africa. As for the eastern Mediterranean SST index, the significant correlations lay between 5°W - 5°E and 7°N - 10°N. The spatial pattern of the West African rainfall correlation with each oceanic mode of variability is instructive. It shows the spatial rainfall differences in response to the influence of a given oceanic internal mode of variability. Although Figs. 2.4 (i-l) and 2.5 show the most robust linkages over the full 1901-2016 period, they could not identify the stationarity of those links nor provide any information about the periods over which the connections were the strongest. There is then a need to test the changes in the correlation according to the length of the samples as well as their starting year. The robustness of the DJF SST-rainfall relationships has been evaluated and displayed in Fig. 2.8 (a-e). For the 1901 - 2016 period, the IPWP index has the strongest correlation of -0.23 with the Guinean Coast rainfall index. The correlation remains constant (the sign) and significant for samples wider than 65 years in that period, with values between -0.4 and -0.2. For samples whose sizes are between 20 and 65 years and starting in the 1920s - 1970s period, this anticorrelation strengthens, with predominant val-

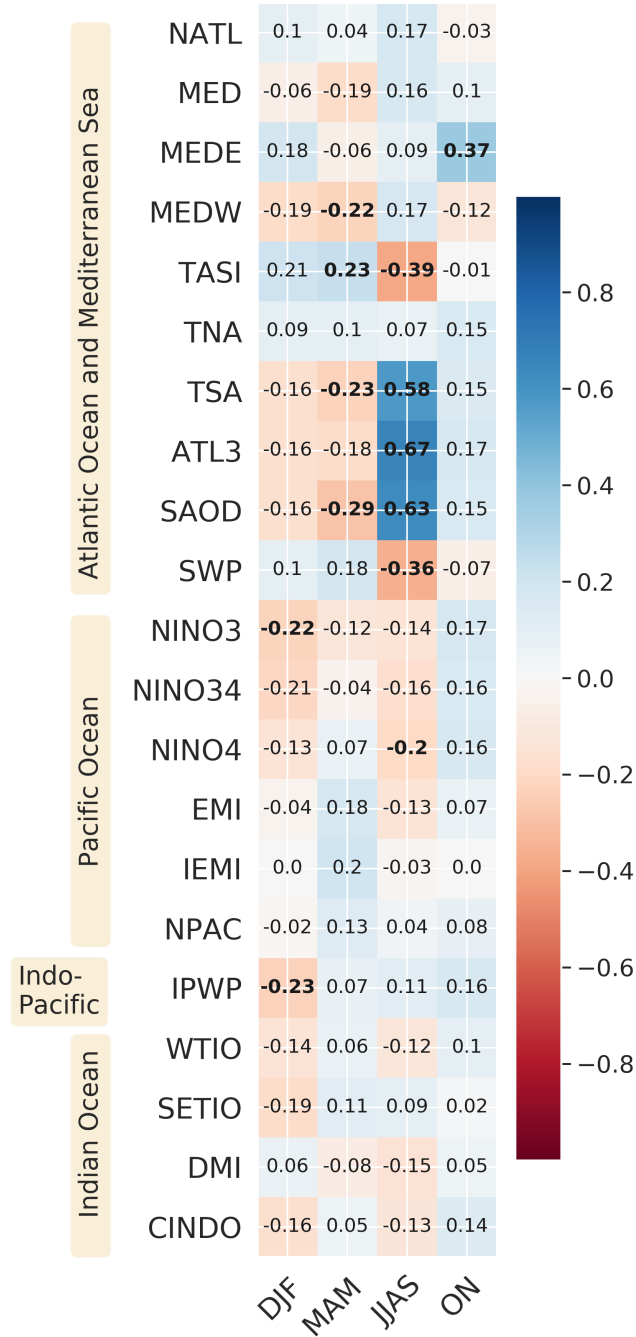


Figure 2.5: Correlation between different SST indices and the Guinean Coast rainfall index for different seasons (1901 - 2016 period). Significant values at 95% confidence level based on a F-test taking into account serial correlations are shown in bold.

ues ranging between -0.75 and -0.4. This highlights that the influence of the IPWP happens mainly on interannual timescales. Likewise, IPWP, ATL3 and Niño3 indices present overall similar changes of their correlation with the GCR index. In addition, the interannual indices of NINO3 and IPWP are strongly coupled all the time, whereas the IPWP interannual linkage with the coastal Guinean rainfall takes place when the IPWP co-varies with the tropical South Atlantic. Consequently, the weakening of the coupling between IPWP and ATL3 before 1920s and after 1970s could explain the non-stationary correlations between the IPWP SST and GCR on interannual timescales (Fig. A.14). This is illustrated in the Figures A.6 and A.7 (d): over the 1925 - 1970 period, the ATL3 is positively correlated with SSTAs in the tropical oceans, while over 1980 - 2016, the core of the ATL3 correlation pattern with SSTAs is mainly restricted to the Atlantic equatorial and southern basin. Consistently, the 1925 - 1970 correlation between the GCR index and SSTAs (Fig. A.2 (i)) shows negative values in the whole tropics, while over 1980 - 2016, weak coupling is found in tropical South Atlantic (Fig. A.3 (i)). During the 1925 - 1970 period, a well pronounced negative rainfall correlation with IPWP and ATL3 extends over the Guinea Coast and north of 10°N (Figs. A.4 (b) and (d)). For 1980-2016, except in very localized small areas, weak correlations are found between rainfall anomalies in Guinea Coast and the SST indices (Figs. A.5 (a)-(e)).

Results with one single 31-year sliding window are available in Fig. 2.9, which illustrates the methodology of the stationarity evaluation for a particular window length. It can be seen that for the DJF season (Fig. 2.9 (a-e)), the eastern Mediterranean SST and the Guinean Coast rainfall correlation is positive but hardly significant between 1930s and 1950s. For the IPWP SST index, the correlation with the Guinean Coast index is stationary from 1935s to 1985s, with values ranging between -0.6 and -0.4. In addition, the Niño3 mode is in opposite phases with the Guinean Coast rainfall in the 1950 - 1985 period (correlation between -0.6 and -0.2). Finally, the ATL3 mode shows negative correlations (between -0.6 and -0.3) in the 1940 - 1975 period. Note that the rainfall index relative to the Sahelian area exhibits a similar stationarity relation with those selected SST indices. The correlations obtained are, however, less significant in general. Lagged correlations between the DJF GCR and the indices of successive 3-month SSTAs averaged over the IPWP and prior to DJF are computed. Results show a robust negative correlation 2 months in advance (Fig. A.1 (a)).

This analysis has demonstrated that over the 1925 - 1970 period, the GCR variability has been mainly driven by tropical SST. Next, the mechanisms associated with IPWP are diagnosed. From the reanalysis data, SSTAs are regressed onto the standardized high-pass SST index of the IPWP over the 1925 - 1970 period (a time period chosen in the period of stationary linkage between IPWP and GC). Results indicate warmer than normal tropical basins related to an anomalous warming of the IPWP. Moreover, this warming is more pronounced in the central

and eastern Pacific Ocean, and appears as an El-Niño-like SST pattern (Fig. 2.6 (a)), which would connect West Africa by the propagation of a Kelvin (Rossby) wave to the East (West) [Rowell, 2001; Gill, 1980; Matsuno, 1966]. A warming in the eastern Pacific leads as a result to a rising over the Pacific and a divergence at upper levels. This is illustrated in Fig. 2.6 (b), where at 300 hPa, a pair of anticyclones is present over the eastern Pacific, as well as minimum negative potential velocity (Fig. 2.6 (c)). Over West Africa, the upper-level circulation is characterized by a convergence at 300 hPa, and a subsidence which dries the atmosphere (Fig. 2.10 (a)). These conditions stabilize the Guinean Coast atmosphere, thus reducing its rainfall. A similar circulation pattern is obtained when regressing the atmospheric fields onto the Guinean Coast rainfall index (Fig. A.10), as it is the case for Niño3 and ATL3 (Fig. A.11 (c)-(d)). In the ATL3 positive phases, otherlies converge over the above normal warm sea in the tropical south Atlantic, southward the equator (at around 10°S). Subsequently, the surface winds diverge in Guinea Coast and imply a downward flow which stabilizes the atmosphere.

This analysis has demonstrated that in general, except in the eastern Mediterranean area, the SST changes display a negative correlation with the Guinean Coast rainfall. For some periods, and for long samples, robust linkages are drawn out between several important SST modes of variability and the Guinean Coast rainfall. The results suggest a leading role of the tropical basins in driving the DJF Guinean Coast rainfall, mainly during 1925s - 1970s.

2.4.2 Oceanic forcing of the pre-monsoon (MAM) Guinean Coast rainfall

The pre-monsoon season accounts for 23% of the Guinean Coast annual rainfall (Table 2.2). Over the 1901 - 2016 period, the MAM rainfall average ranges between 0 and 3 mm d^{-1} in most of the areas north of 10°N . South of this limit, the mean rainfall rate increases, and fluctuates between 3 and 7 mm d^{-1} (Fig. 2.4 (b)). The MAM Guinean Coast rainfall mode has a correlation pattern which extends from the coastal areas to the northernmost area of the Sahel (Fig. 2.4 (f)). The correlation values are overall positives. However, the highest values (between 0.3 and 0.7) are mainly located in the Guinean Coast region. Broadly, compared with the SST pattern obtained in the Harmattan season, the MAM SST anomalies are weakly correlated with the Guinean Coast rainfall index (Fig. 2.4 (j)). This SST pattern shows significant negative correlations in the tropical South Atlantic, central subtropical South Pacific, and western Mediterranean Sea. In the subtropical South Atlantic and eastern tropical North Pacific, the correlations are positive in very small areas, with similar absolute values.

From Fig. 2.5, it can be seen that the South Atlantic Ocean Dipole index has the highest correlation with the Guinean Coast rainfall index for the 1901 - 2016

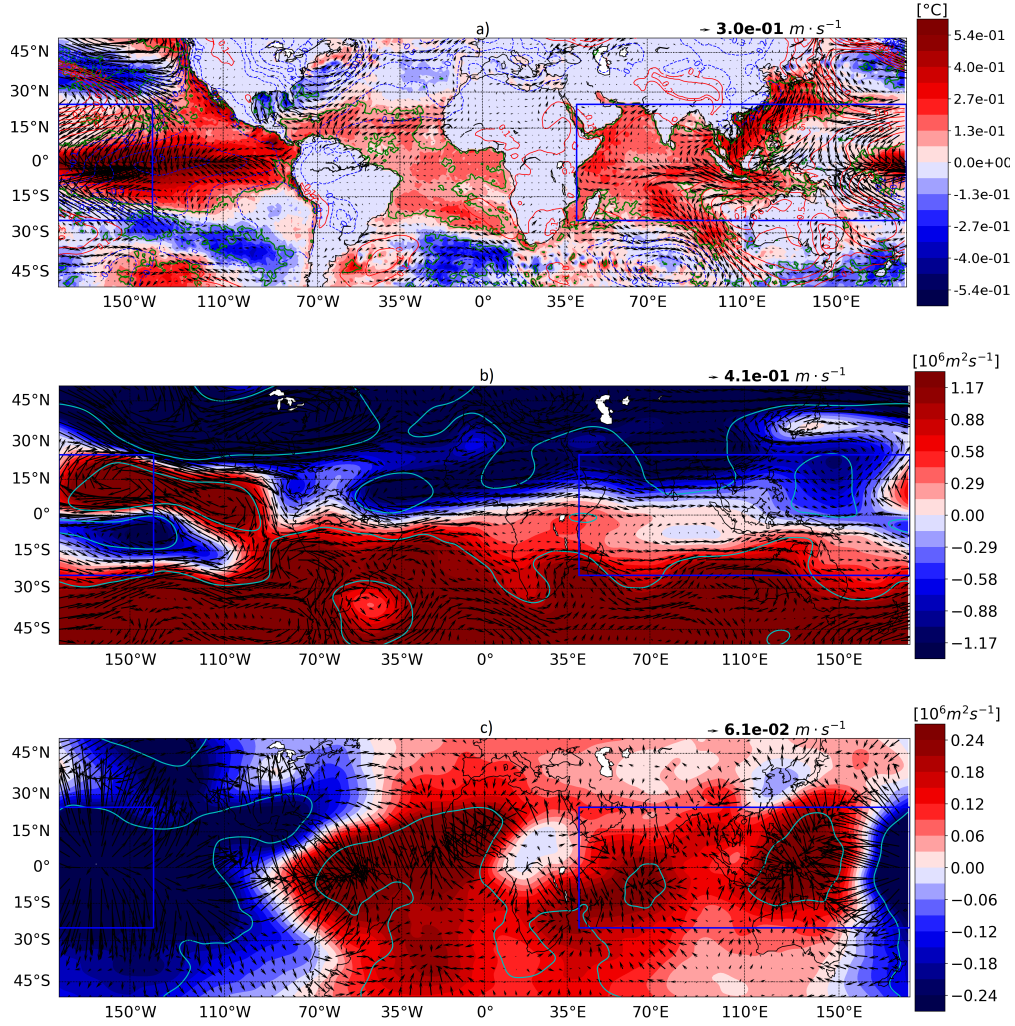


Figure 2.6: DJF: regression of (a) sea surface temperature anomalies (colors, in °C), sea level pressure anomalies (blue and red contours, in hPa), surface winds anomalies (arrows); (b) 300 hPa anomalies of streamfunction (colors) and non-divergent component of the wind (arrows); (c) 300 hPa anomalies of potential velocity (colors) and divergent component of the wind (arrows) onto the standardized high-pass IPWP SST index. Green contours in (a), and blue contours in (b) and (c) indicate significant shaded values at 95% confidence level, according to a Student test. The regressions are computed for the 1925 - 1970 period. The blue box indicate the area used to compute the Indo-Pacific warm pool SST index.

period (-0.29), followed by the tropical South Atlantic and the tropical Atlantic SST meridional gradient indices (0.23 and -0.23 respectively). The results of the correlation between the main SST indices with the Guinean Coast rainfall in MAM are represented in Fig. 2.7 (f-j). They suggest that years of warm (cold) phases of the tropical South Atlantic or the South Atlantic Ocean Dipole are associated with a reduction (an intensification) of the West African rainfall. In the coast of Guinea, significant values of the spatial correlation between the TSA and SAOD indices and anomalies of rainfall are mostly restricted to the central southernmost and eastern areas. Finally, the Mediterranean Sea (mainly the western area) also exhibits significant negative correlations in a localized area over to the central coastal regions. The analysis of the robustness of the MAM SST-rainfall interrelation con-

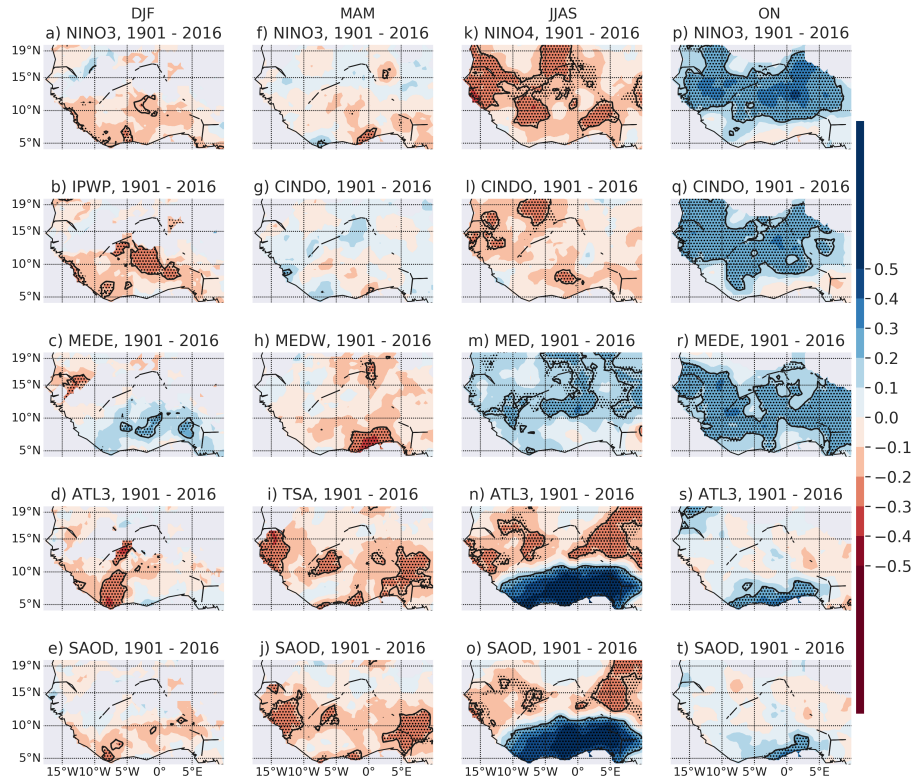


Figure 2.7: Rainfall patterns associated with the main SST modes of variability for each season: (a-e) DJF, (f-j) MAM, (k-o) JJAS, (p-t) ON. Those patterns are obtained by correlating the interannual SST indices with rainfall anomalies in West Africa. Contour lines and dots indicate significant correlations at 95% confidence level according to a F-test with a standard and effective degree of freedom, respectively.

firms that the South Atlantic Ocean Dipole has the strongest and most stationary link with the Guinean Coast rainfall (Figs. 2.8 (j) and 2.9 (j)). For samples starting between 1901 and 1940, and for periods between 20 and 65 years, the correlations calculated range between -0.4 and -0.6. In the first two decades of the 20th century, they exceed -0.6 for shorter samples. Interestingly, this negative correlation is reversed in the 1980s, where a strong coupling has been found between the Niño3 and the Guinean Coast rainfall indices. As a result, after the 1980s, El-Niño positive phases are associated with anomalous cooling in the eastern tropical South Atlantic (Fig. A.7 (f,i,j)), causing a reduction of rainfall over coastal Guinea (Fig. A.5 (f)). Thus, the reduction effect of El-Niño should weaken or dominate the rainfall increase related to the cooling of the tropical South Atlantic during this season (Fig. A.5 (i,j)). Before the 1980s, the eastern tropical Pacific and the tropical South Atlantic are weakly connected (Fig. A.6 (f,i)). Furthermore, the rainfall and SST correlation patterns displayed by indices of the tropical South Atlantic and the South Atlantic Ocean Dipole are similar, as well as their stationarity features.

Incidentally, remote effects of SST changes in the western Mediterranean Sea on the MAM coastal Guinean rainfall are globally weak and not significant for shorter timescales. Lagged correlations between the GCR in MAM and the successive 3-month means of SST indices indicate that the influence of the tropical South Atlantic and the meridional SST gradient in the tropical Atlantic (TASI, also called the interhemispheric SST mode) starts two months in advance, while the delay is one month for the South Atlantic Ocean Dipole (Fig. A.1 (b)). Further analyses (not shown) demonstrate that the northern and southern core of TASI are not simultaneously coupled between them (which is in accordance with Gu [2010]) nor with the Guinean Coast rainfall. As for the SAOD, much of the strong correlations in the stationarity analysis between TASI and GCR coincide with that of the tropical South Atlantic and GCR. This reveals the importance of the tropical South Atlantic in driving the MAM Guinean Coast rainfall, which could be strengthened by the development of the southern (northern) core of the SAOD (TASI).

From the reanalysis data, the regression map of SSTAs and surface winds associated with the standardized SAOD index (regression over the 1901 - 1946 period, Fig. A.8 (b)) shows that positive phases of the SAOD are associated with warm SST anomalies in the most part of the tropical South Atlantic and cold SSTAs in its south-western area. According to Lindzen-Nigam process [Lindzen and Nigam, 1987], these surface gradients of SST anomalies induce anomalous surface pressure gradients which accelerate South-North surface winds from the South West Pole to the North East Pole. In addition, at the equator, winds are northwesterlies. This winds combination produces a surface convergence between 10°S and the equator, followed by an ascent centered at 4°S in the eastern tropical South Atlantic. At upper-levels, the divergent northward wind favors a convergence over West Africa and a subsidence which decreases its humidity (Fig.

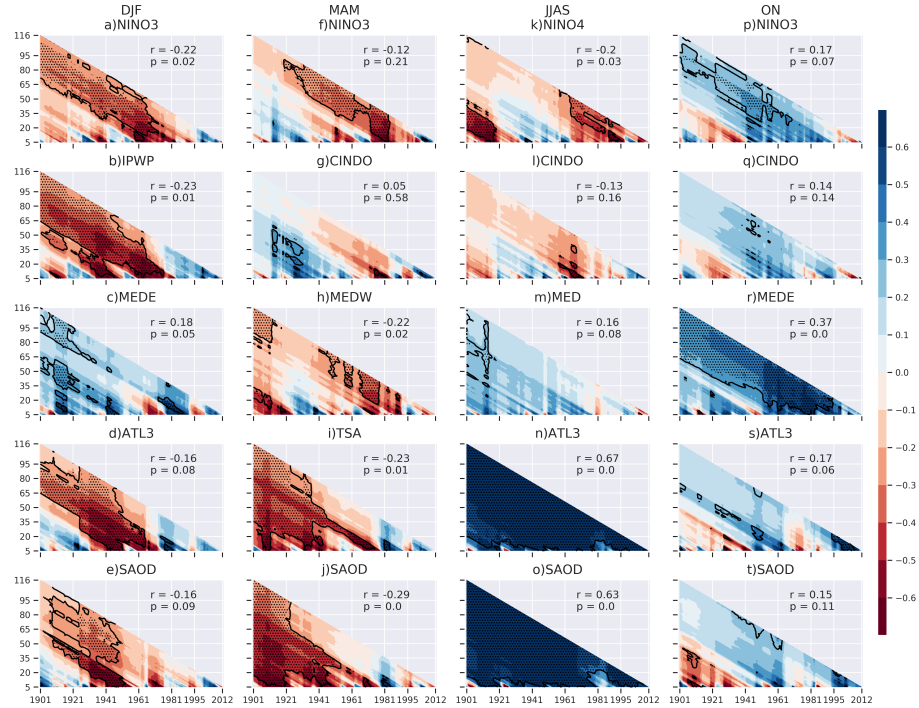


Figure 2.8: Running correlation between the interannual component of the Guinean Coast rainfall index and indices of the main SST modes of variability for different starting years (horizontal axis), different window lengths (vertical axis) and different seasons: (a-e) DJF, (f-j) MAM, (k-o) JJAS, (p-t) ON. Contour lines and dots represent significant correlation values at 95% confidence level based on a F-test with a standard and effective degree of freedom respectively. Displayed r and p values correspond to the correlation value of the full series and the associated Student-test p_{value} respectively.

2.10 (b)). In the tropical South Atlantic, positive values of the vertically integrated moisture flux convergence are observed over 0 - 10° S, while they are negative north of the equator, including West Africa (Fig. A.9 (b)). The moisture flux divergence in the coast of Guinea is consistent with the subsequent rainfall reduction. Consistently, the circulation fields related to the coastal Guinean rainfall point out a dipole-like pattern of SSTAS in the tropical South Atlantic, and a similar circulation feature (Fig. A.12 (a)-(b)). It also shows in the tropical Atlantic a pattern reminiscent of the interhemispheric mode which is in opposite phase with the SAOD, and increases the cross-equatorial northerlies and their convergence in the north east pole of the SAOD.

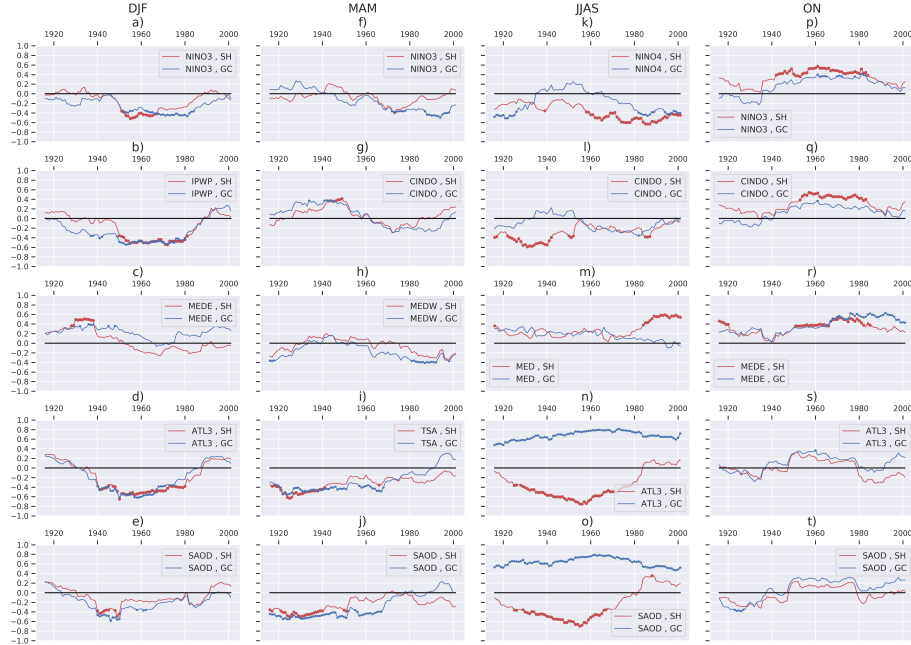


Figure 2.9: Running correlation of 31-year sliding window between the Guinean Coast rainfall and SST modes of variability for different seasons (blue curves): (a-e) DJF, (f-j) MAM, (k-o) JJAS, (p-t) ON. Results for the Sahel (red curves) are also shown (but not discussed). Stars and dots mark significant values at 95% confidence level based on a F-test with a standard degree of freedom. Years in horizontal axes represent the middle of the 31-year sliding window.

The opposite phase relationship between the MAM GCR and the SST anomalies in the tropical South Atlantic is in agreement with the finding of Meynadier et al. [2015], who linked the appearance of a cold tongue in the eastern equatorial Atlantic to the beginning of the early raining season in the coastal regions of West Africa (also Nguyen et al. [2011]; Leduc-Leballeur et al. [2012]). These cold SST conditions increase the northern front of the cold tongue which accelerates northerlies from the equator to the coast of Guinea. Accordingly, moist air is brought to the Gulf of Guinea and marks the onset of the raining season. Moreover, the appearance of the Atlantic cold tongue has been shown to be related to a strengthening of southeasterlies associated with the northward march of the St Helena anticyclone. It results in an intense upwelling around 4°S [Caniaux et al., 2011].

In summary, except in years around 1980 and afterward, SST changes in the tropical South Atlantic are the main oceanic drivers of the Guinean Coast rainfall

variability during the pre-monsoon season. This influence is reinforced by the development of the southern (northern) lobes of the South Atlantic Ocean Dipole (Interhemispheric SST mode). Note that for the selected SST modes, the Sahelian and Guinean Coast rainfall indices have similar covariability stationarity features (Fig. 2.9 (f-j)).

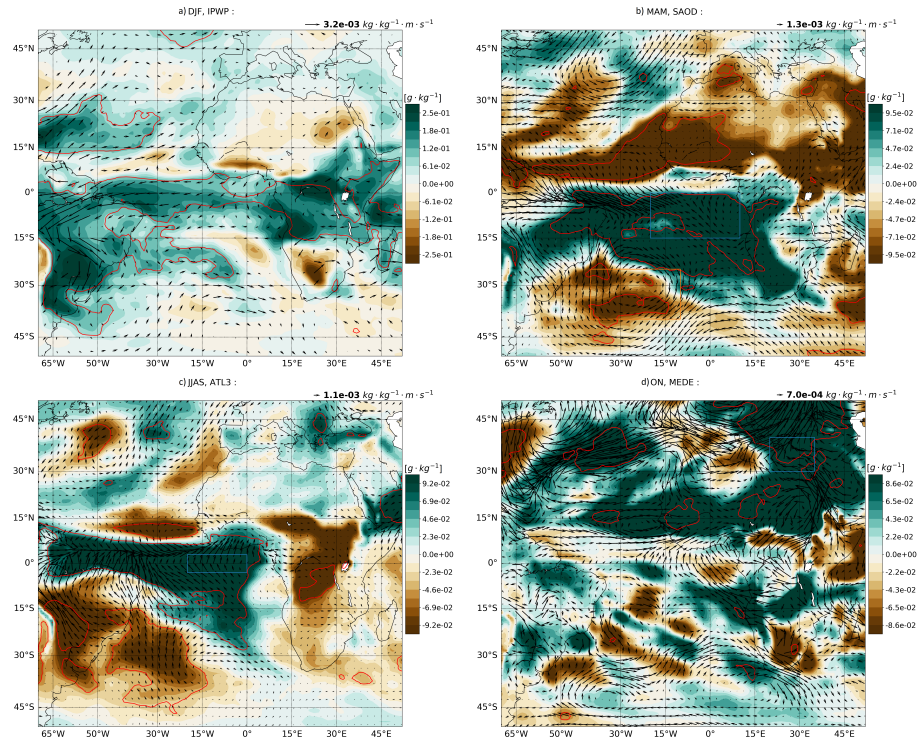


Figure 2.10: Regression of anomalies of specific humidity (colors) and moisture flux (arrows) at 850 hPa onto standardized high-pass SST indices of (a) IPWP, (b) SAOD, (c) ATL3, (d) MEDE. Red contours indicate significant specific humidity at 95% confidence level. The blue rectangle on each map delineates the domain used to compute the SST index. The regressions are evaluated for the 1925 - 1970, 1901 - 1946, 1901 - 2010, 1950 - 1995 periods respectively.

2.4.3 Oceanic forcing of the monsoon (JJAS) Guinean Coast rainfall

The JJAS rainfall accounts for 58% of the Guinean Coast annual rainfall (Table 2.2). Note that the monsoon season in this analysis contains 4 months, unlike the

other seasons which have 2 or 3 months. The spatial distribution of the average seasonal rainfall rate reveals highest values (between 7 and 19 mm · day⁻¹) in the westernmost (18°W - 7°W) and easternmost (4 - 10°E) areas of the Guinean Coast (Fig. 2.4 (c)). In the rest of that zone, the rainfall mean values lie between 3 and 7 mm · day⁻¹. The rainfall correlation pattern that characterizes the Guinean Coast rainfall variability during this season shows positive values that are positioned in the 5 to 10°N latitudinal band, stretching from West to East (Fig. 2.4 (g)). North of 10°N, significant negative weak correlations are obtained in the eastern Sahel. This reveals that the monsoon Guinean Coast rainfall mode is well separated from the Sahelian mode, and this could be explained by differences in external SST forcing of the West African rainfall. For instance, Polo et al. [2008] have analysed the covariability between the boreal summer West African rainfall and the tropical oceanic basins. They showed that the leading mode links the Guinean rainfall mode with the Atlantic Equatorial mode, while the Sahelian rainfall mode is more related to the Mediterranean Sea surface temperature changes. In parallel, the West African Westerly Jet (WAWJ) makes an important contribution to the zonal moisture transport onto the Sahel, as several studies showed a reduction of the rainfall in the Sahel associated with a negative zonal moisture transport, and a weakening of the WAWJ [Lélé et al., 2015]. The global-scale SST influence on this jet in summer, could impact its moisture supply to the Sahel, whereas Guinea Coast remains under a more regional SST control.

The Guinean Coast rainfall index correlation with SSTAs in boreal summer highlights two main subregions in the Atlantic Ocean (Fig. 2.4 (k)). On the one hand, a core of positive correlations appears in the region across the equatorial and southeastern tropical Atlantic. Above (below) normal SST in that region produces positive (negative) Guinean Coast rainfall anomalies. On the other hand, in the southwestern tropical Atlantic, negative correlations are obtained, meaning that below (above) normal SST conditions in that area are associated with an increase (a decrease) in rainfall in the Guinean Coast. Additionally, significant positive correlation values are found in the subtropical North Atlantic, offshore areas in the extratropical North Atlantic and the western tropical Pacific. Weak correlations are seen in the eastern tropical Pacific and Indian Ocean, while negative values are obtained in the Maritime Continent oceanic area and eastward.

The Atlantic Niño and the South Atlantic Ocean Dipole appear as the fundamental oceanic internal modes of variability associated with the boreal summer Guinean Coast rainfall variability. The sign of their correlation is opposite compared with the MAM season. For the 1901 - 2016 period, both modes have a positive correlation with the Guinean Coast rainfall index, with values of 0.67 and 0.63 for ATL3 and SAOD, respectively (Fig. 2.5). Figures 2.7 (n) and (o) indicate that they tend to create a dipole rainfall pattern in West Africa, with a band of positive correlations in Guinea Coast and negative values in Sahel [Adler et al., 2003;

Nnamchi and Li, 2011]. Then, warm (cold) phases of the Atlantic Niño or South Atlantic Ocean Dipole are associated with rainfall anomalies of positive (negative) sign in the Guinean Coast and negative (positive) sign in the Sahel. This dipole is present in our analysis because of the long period over which the analysis is performed. It was strong over the 1901 -1970 period (Fig. A.4 (n-o)), while it weakens afterward (Fig. A.5 (n-o)). This is explained by the fact that the global SST in the tropical basins have started to covary with the rainfall in West Africa. After 1970s, positive SST anomalies in the eastern Atlantic equatorial area are indeed associated with negative (positive) SST anomalies in the equatorial eastern Pacific (Maritime Continent oceanic area) , while before 1970s the linkage with the eastern equatorial Pacific is non-existent (Figs. A.6 and A.7 (n-o)). Thus, as shown by Losada et al. [2012] an increased rainfall over the Sahel associated with those La-Niña conditions cancels the negative rainfall anomaly induced in the Sahel by warm phases of ATL3 [Losada and Rodríguez-Fonseca, 2015]. Regarding the tropical Pacific, Indian Ocean and the Mediterranean Sea, their influences are more concentrated on the Sahelian area in JJAS over the 1901 - 2016 period (Fig. 2.7 (k-m)).

The strong positive correlation between the Guinean Coast rainfall index and both ATL3 and SAOD is stationary over the whole period investigated and at all timescale (Figs. 2.8 (n,o) and 2.9 (n,o)). Significant correlation values are generally between 0.5 and 0.8, and exceed 0.8 in several cases. Additionally, the Niño4 and Guinean Coast indices have a non-stationary linkage, with significant negative correlations in the early and last decades of the 1901 - 2016 period. As for the Indian Ocean, a coupling is found in the 1970s, with negative correlations (Fig. 2.8 (i)).

Lagged correlations between the JJAS rainfall index of the coast of Guinea and SST indices averaged over the successive 3-month prior to JJAS are displayed in Fig. A.1 (c). Results show a robust negative (positive) correlation with the meridional SST gradient in the tropical Atlantic (tropical South Atlantic) 8 (7) months in advance. Indices of the South Atlantic ocean dipole and Atlantic Niño display the highest correlation (around 0.6) for the first 3 lags and their influence is notable 5 months in advance.

During the monsoon season, ATL3 positive phases lead to a warming of the eastern equatorial Atlantic, which decreases the surface pressure gradient between West Africa and the Atlantic Ocean (Fig. A.8 (c)). Trade winds are subsequently weakened, altering the northward penetration of the monsoon flow into West Africa. Moist air converges over the warm water at the equator, rises and is advected to coastal Guinea, as shown in Fig. 2.10 (c). The convergence of the moisture flux over the Guinean Coast (Fig. A.9 (c)) contributes to increase its rainfall [Giannini, 2003; Polo et al., 2008; Rodríguez-Fonseca et al., 2015; Suárez-Moreno et al., 2018]. In sum, the JJAS rainfall variability in the Gulf of Guinea is

more driven by SST changes in the equatorial and tropical South Atlantic due to the Atlantic Equatorial mode and the South Atlantic Ocean Dipole. A similar circulation feature is obtained when surface and atmospheric fields are regressed onto the Guinean Coast rainfall index (Fig. A.12 (c)-(d)).

2.4.4 Oceanic forcing of the post-monsoon (ON) Guinean Coast rainfall

The post-monsoon season (October and November) accounts for 15% of the Gulf of Guinea annual rainfall (Table 2.2). The rainfall mean vary between 1 and 12 mm d⁻¹ over the coastal region, with maxima located in its easternmost and westernmost sectors (Fig. 2.4 (d)). The center of action of the rainfall correlation pattern related to Guinea Coast rainfall mode covers all the West African region, with positive values everywhere. However, correlation maxima stretch across 7 and 12°N (Fig. 2.4 (h)). The variability of the Guinean Coast rainfall in ON is in phase with SST changes in the eastern Mediterranean basin, the Red Sea and in very limited tropical South Atlantic areas (Fig. 2.4 (i)). Overall, the tropical basins are weakly connected with the ON rainfall changes in Guinea Coast.

For the 1901-2016 period, the eastern Mediterranean SST and Guinean Coast rainfall indices present the highest correlation value, 0.37 (Fig. 2.5). Warm (cold) phases of this oceanic basin cause more (less) rainfall over West Africa, as illustrated in Fig. 2.7 (r). During the first half of the 20th century, the coupling between MEDE and GCR is weak and significant for only long timescales (above 35 years). This connection strengthens from the 1950s, where their correlation values are higher than 0.4 for all timescales above 10 years (Figs. 2.8 (r) and 2.9 (r)). During the 1950-1995 period, warm phases of the eastern Mediterranean Sea in ON are associated with a warming of the equatorial and eastern tropical South Atlantic and an increase of the rainfall in the Gulf of Guinea (Fig. A.6 (r), Fig. A.4 (r,s)). Nevertheless, prior to the 1950s, SST changes in the eastern Mediterranean Sea are weakly and negatively correlated with SSTAs in the tropical South Atlantic. A warming of the eastern Mediterranean Sea is associated with a cooling of the eastern equatorial Atlantic (Fig. A.7 (r)), potentially canceling their respective effect on the rainfall in the coastal Guinea. As a consequence, the connection between the eastern Mediterranean Sea and the Guinean Coast vanishes over that period. In addition, the correlation between the GCR index and SST anomalies over the 1901 - 1950 period suggests a strong coupling with SST in the Maritime Continent area, where significant negative values appear. Elsewhere in the tropical basins, the connections are weak, while positive correlations are seen in the eastern Mediterranean basin. Those changes in the covariability between the different oceanic basin have been shown to produce non-stationarities in the teleconnections processes [Suárez-Moreno et al., 2018].

Furthermore, the Atlantic Equatorial mode displays positive correlations with rainfalls located along the coast of West Africa, while the indices of Niño3 and CINDO show significant positive correlations stretching from the northern limit of the Guinean Coast to Sahel (Fig. 2.7 (p,q,s)). The influence of ATL3 is noticeable only for shorter timescales over the 1995 - 2016 period (Fig. 2.8 (s)). The influence of Niño3 is significant for longer samples (above 35 years) starting before 1970 (Fig. 2.8 (p)) whereas for shorter samples, no robust link is found.

Several studies showed that warm phases of the eastern Mediterranean Sea in summer are associated with a rising of moist air over the basin, which in turn is transported southward by the local wind circulation. This leads to a moisture convergence in the Sahel and Coast of Guinea, with a subsequent increase in rainfall [Fontaine et al., 2009; Gaetani et al., 2010]. Regression analysis over the 1950 - 1995 period during ON reveals that warm situations of the eastern Mediterranean Sea (Fig. A.8 (d)) lead to a rising of moist air over the eastern basin, which is then transported by low-level northeasterlies (between 700 and 850 hPa) toward West Africa (Fig. 2.10 (d)). In addition, moist air from the Atlantic Ocean is brought by the low-level westerlies, southeasterlies and northerlies (in the West flank of the North Africa). Those winds increase the moisture supply to West Africa. Analogously, the regression of the circulation fields onto the Guinean Coast rainfall shows an important warming in the eastern Mediterranean basin (Fig. A.13 (a)), and the presence of a cyclonic circulation over northern Africa which transports equatorward moisture by its western flank into West Africa (Figs. A.13 (b),(d)). A low-level eastward component of the moisture flux is also present in Fig. A.13(c), resulting in more rainfall over West Africa.

In conclusion, only changes of the eastern Mediterranean SST have a strong and stationary correlation with the interannual Guinean Coast rainfall in the post-monsoon season. Anomalous warming over the Mediterranean Sea strengthens the Saharan Heat Low (SHL) which is in accordance with Diakhaté et al. [2019] hypothesis. Moreover, the maritime extension of the SHL induces over the tropical North Atlantic, a cyclonic circulation that provokes in its southern flank a strong low-level westerly jet [Grodsky, 2003; Pu and Cook, 2010; Liu et al., 2019] which transports humidity toward the coastal Guinea, leading therefore to an increased rainfall. This Mediterranean influence is noticeable two months in advance, as shown in Fig. A.1 (d), and is the most robust of the lagged correlations between the SST indices and the ON GCR index. Note that, in this season, the linkages between the Niño3, MEDE and the Guinean Coast rainfall are similar to the one obtained with the Sahelian rainfall. Moreover, Sahelian rainfall linkages with the Indian Ocean and Niño3 are more significant than in the Guinean Coast case (Fig. 2.9 (p,q,r)).

2.5 Conclusion

This study has analyzed the interannual variability of rainfall in the Guinean Coast region and its links with SST changes over the 1901 - 2016 period. Unlike numerous studies, our analysis is not restricted to the monsoon season (which only accounts for 58% of the annual rainfall in this region) and highlights how the links between rainfall and SST vary according to the season.

Our results show that the JJAS Atlantic influence on the Guinean Coast rainfall is stronger than the impact of all other basins and for all other seasons. The Harmattan (DJF) Guinean Coast rainfall interannual variability has been highly influenced by SST changes in the tropical basins over 1925 - 1970. Variability of the Indo-Pacific warm pool, El-Niño and Atlantic equatorial mode are anticorrelated with the interannual Guinean Coast rainfall. In DJF, years of larger than average SST in the IPWP are associated with positive SST anomalies in the eastern tropical Pacific. The SST imprint of those conditions appears like an El-Niño event which leads to an eastward/westward propagation of Kelvin/Rossby waves through the atmosphere. This enhances a high level wind convergence over the Guinean Coast, followed by a downward motion of the air column [Rowell, 2001], stabilizing the Guinean Coast atmosphere and reducing its rainfall. Moreover, after the 1980s, the connection between the IPWP and the equatorial Atlantic has weakened. This might have reduced the remote effect of IPWP SST changes on the Guinean Coast rainfall.

The pre-monsoon (MAM) and the monsoon (JJAS) rainfall variability of the Guinean Coast are mainly associated with the South Atlantic Ocean Dipole and the Atlantic Niño. Those SST modes are negatively correlated with the pre-monsoon Guinean Coast rainfall, with the strongest impact displayed by the SAOD. Their linkages with the rainfall fluctuations in coastal Guinea are stationary from 1901 to 1980. Afterwards, an inversion of the correlation occurs and is potentially due to the strengthening of the El-Niño remote effect. Contrary to the MAM season, the Atlantic Niño and the South Atlantic Ocean Dipole are both positively correlated with the JJAS Guinean Coast rainfall index. Those SST-rainfall interannual interrelations are stationary over the 1901 - 2016 period, and stronger than in the previous season. The Atlantic Niño and the Guinean Coast rainfall indices have the highest correlation value for that period. Positive phases of the SAOD and ATL3 modes are associated with negative (positive) Guinean Coast rainfall anomalies in the pre-monsoon (monsoon) season. Similar physical mechanisms explain the linkages between those Atlantic modes and the Guinean Coast rainfall. In summer, when the SAOD or ATL3 index is above normal, positive SSTAs in the eastern tropical Atlantic lead to a surface winds convergence at the equator and a strong ascent. Moist air is then advected into the Guinean Coast, increasing the rainfall. In the case of a positive phase of the SAOD, the surface wind divergence in the

South West Pole strengthens the convergence in the ATL3 area [Nnamchi and Li, 2011]. In MAM, the surface convergence zone due to the anomalous warming of the tropical South Atlantic in positive phases of the SAOD, is shifted south of the equator. This leads to anomalous subsidence over the coast of Guinea, and a reduction of its rainfall.

The post-monsoon Guinean Coast rainfall regime is more correlated with SST changes in the eastern Mediterranean Sea on interannual timescales. The analysis of their indices shows that they have a positive correlation which is strong and stationary over the 1950 - 2016 period. The Guinean Coast rainfall increase (decrease) coincides with a zonal incoming (outgoing) low-level moisture flux induced by a strengthening (weakening) of the low level westerly jet in the southern flank of a cyclonic circulation over the tropical North Atlantic, driven in turn by a decrease (increase) of the mean sea level pressure anomalies. This cyclonic circulation over the tropical North Atlantic appears as a maritime extension of the strengthened Saharan Heat Low associated with the anomalous warming in the eastern Mediterranean sea. Interestingly, the robust linkages drawn out provide notable different lagged correlations which are important for the seasonal rainfall forecasting in the coastal Guinea. For DJF, the influence of the IPWP can be predicted two months in advance. For the MAM rainfall, the tropical South Atlantic and the meridional SST gradient in the tropical Atlantic present noticeable correlations two months in advance. Five months early, the JJAS rainfall is in phase with SST changes in the Atlantic Niño area, while for the ON rainfall, the influence of the eastern Mediterranean Sea can be forecasted two months ahead.

Further analyses are still needed to better assess the large-scale oceanic influences on the interannual, decadal to multidecadal variability of the Harmattan, pre-monsoon and post-monsoon rainfalls in the coastal Guinea. It would be particularly valuable to further investigate the time-lagging influence of the oceanic basins, which could be done e.g. via statistical methods like the extended maximum covariance analysis. Finally, our results provide observationally-based diagnostics that could be used for the evaluation of historical simulations performed with climate models. Conversely, SST sensitivity experiments are needed to identify the mechanisms behind those statistical relationships. From this perspective, the Global Monsoons Model Intercomparison (GMMIP) simulations performed in CMIP6 [Zhou et al., 2016] should help address this issue.

3

Weakened impact of the Atlantic Niño on the future equatorial Atlantic and Guinea Coast rainfall

Abstract

The Guinea Coast is the southern part of the West African continent. Its summer rainfall variability mostly occurs on interannual timescales and is highly influenced by the sea surface temperature (SST) variability in the eastern equatorial Atlantic, which is the center of action of the Atlantic Niño mode. Using both historical and scenario (SSP5-8.5) simulations from 31 General Circulation Models (GCMs) participating in the sixth phase of the Coupled Model Intercomparison Project (CMIP6), we first show that these models present a wet bias during boreal summer. This bias is associated with too high mean boreal summer SSTs in the eastern equatorial and south Atlantic regions. Next, we analyze the near-term, mid-term and long-term changes of the Atlantic Niño relative to the present-day situation, in a climate with a high anthropogenic emission of greenhouse gases. We find a gradual decrease of the equatorial Atlantic SST anomalies associated with the Atlantic Niño in the future. This result reflects a possible reduction of the Atlantic Niño variability in the future due to a weakening of the Bjerknes feedback over the equatorial Atlantic. In a warmer climate, an anomalous higher sea level pressure in the western equatorial Atlantic relative to the eastern equatorial Atlantic weakens the climatological trade winds over the equatorial Atlantic. As a result, the eastern equatorial Atlantic thermocline is deeper and responds less to the Atlantic Niño events. Among the models that simulate a realistic rainfall pattern

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associated with the Atlantic Niño in the present-day climate, there are 12 GCMs which project a long-term decrease of the Guinean Coast rainfall response related to the Atlantic Niño. In these models, the zonal 850 hPa wind response to the Atlantic Niño over the equatorial Atlantic is strongly attenuated in the future climate. We also find that 12 other GCMs show no robust change in the patterns associated with the Atlantic Niño. There is a higher confidence in the mid and long-term reduction of the rainfall associated with the Atlantic Niño over the Atlantic Ocean than over the Guinea Coast. We also found a projected decrease of the convection associated with the Atlantic Niño in the majority of the models.

3.1 Introduction

The West African Monsoon (WAM) generally begins in mid-June and is characterized by a rapid shift of the rain band from the coastal areas to the Sahel region [Hansen, 2002; Sultan et al., 2005]. In July and August, the development of the Atlantic cold tongue is at its maximum (due to upwelling over the eastern equatorial Atlantic), which leads to higher pressure over the Gulf of Guinea and the coastal regions of West Africa. As a consequence, during July and August, the atmospheric circulation above Guinea Coast is characterized by a downward flow at upper levels. This leads to the so-called "little dry season" [Adejuwon and Odekunle, 2006; Fink et al., 2017; Wainwright et al., 2019], during which the Guinean coast rainfall decreases in magnitude compared to its mean over the April-May-June season.

Over the 20th century, the interannual variability of the Guinean coast rainfall was strongly influenced by the sea surface temperature (SST) fluctuations in the eastern equatorial Atlantic [Giannini, 2003; Polo et al., 2008; Suárez-Moreno et al., 2018]. This oceanic area corresponds to the center of action of the leading oceanic mode of variability in the tropical Atlantic [Zebiak, 1993]. Several terms are used to define this oceanic mode of variability, namely, Atlantic equatorial mode (hereafter referred to as AEM), Atlantic Niño and Atlantic zonal mode [Servain et al., 2000; Murtugudde et al., 2001; Ruiz-Barradas et al., 2000]. We will use both the Atlantic Niño and AEM terms independently in this study, without any difference between them. In positive phases of the AEM, the eastern equatorial Atlantic is warmer than in a normal situation and the anomaly is amplified through the Bjerknes feedback [Keenlyside and Latif, 2007; Lübbecke et al., 2018]. The positive SST anomalies lower the surface pressure gradient over the equatorial Atlantic, and thus weaken the equatorial trade winds. This leads to westerly anomalies to the west of the abnormal warm surface area, which in turn deepens the thermocline to the east. In consequence, the cooling due to the oceanic upwelling is reduced, which reinforces the initial warming. Moreover, another mechanism involving the southward advection of warm seawater from the subtropical North Atlantic can lead to the de-

velopment of some AEM events [Richter et al., 2012]. Other theories suggest an important contribution of the equatorial oceanic Rossby and Kelvin waves [Foltz and McPhaden, 2010b,a; Burmeister et al., 2016] and thermodynamic processes [Nnamchi et al., 2015] to the AEM variability. In order to quantify the relative contributions of the different processes driving the AEM variability, Jouanno et al. [2017] highlighted the dominant role of the dynamical forcing (i.e the Bjerknes feedback) relative to the thermodynamic processes (i.e air-sea heat flux exchanges). They argued that biases in the atmospheric components of most of the GCMs participating in the CMIP project lead to the underestimation of the dynamic part of the Atlantic Niño forcings.

The general response of the atmosphere to Atlantic Niño positive phases is a modification of the Walker circulation, characterized by rising motion and upper-level divergence in the Atlantic region. In the central Pacific, a compensating upper-level convergence and sinking motion also triggers a Gill-type response in vorticity [Gill, 1980]. The Gill-type response is characterized by a pair of upper-level anticyclones to the west and a pair of upper-level cyclones to the east of the abnormal warm oceanic regions. These upper-level divergence and vorticity responses are related to each other by Sverdrup balance [Kucharski et al., 2009; Losada et al., 2010b; Hamouda and Kucharski, 2019]. The divergence and vorticity responses are generally baroclinic, and are of opposite sign at low levels, meaning in the Indian region a low-level anticyclone is present, which leads to reduced Indian Monsoon rainfall [Kucharski et al., 2007, 2008, 2009]. On the other hand, the sinking motion in the central Pacific can lead to easterly surface winds in the central-western Pacific, which could in turn potentially induce a La-Niña event [Rodríguez-Fonseca et al., 2009]. Finally, the development of a La-Niña event due to a warm phase of the Atlantic Niño would favor positive rainfall anomalies over the Indian Peninsula, which would counteract the negative rainfall anomalies associated with the Atlantic Niño [Ding et al., 2012].

Conversely, Tokinaga et al. [2019] found that the multi-year La Niña conditions could trigger a positive Atlantic Niño. Considering the tropical basins separately, an anomalous warming of the eastern equatorial Atlantic induces a dipolar rainfall response over West-Africa in boreal summer: a decrease of the rainfall in the Sahel region and an increase of the rainfall over Guinea Coast [Losada et al., 2010a]. However, below normal sea surface temperatures in the eastern tropical Pacific lead to an increase of the rainfall in the Sahel. After the 1970s, the coupling between the eastern equatorial Atlantic and the eastern tropical Pacific has strengthened, and the two basins are characterized by an opposite phase relationship. Therefore, a positive phase of the Atlantic Niño is associated with negative SST anomalies in the eastern tropical Pacific. This results in rainfall anomalies of opposite signs over the Sahel, which damps the West-African dipolar rainfall response associated with the Atlantic Niño [Losada et al., 2012]. By contrast,

statistical analyses have demonstrated that the relation between SST changes in eastern equatorial Atlantic and the rainfall in the Guinea Coast during the boreal summer is stationary [Diatta and Fink, 2014; Rodríguez-Fonseca et al., 2015; Worou et al., 2020].

Results from the General Circulation Models (GCMs) participating in the Coupled Model Intercomparison Project (CMIP) show that from the fifth phase (CMIP5) to the sixth phase (CMIP6) of the Coupled Model Intercomparison Project, the surface temperature biases have been reduced over the tropical Atlantic, as pointed out by Richter and Tokinaga [2020] in an analysis of the pre-industrial control experiment performed with 33 models. Nevertheless, they found that in many models a warm bias remains in the east of the basin and along the western coast of Africa (Gulf of Benguela, western boundaries of West Africa), while cold biases exist in the west of the basin. Despite the presence of these biases, the models have improved their representation of the spatial patterns of the Atlantic Niño mode. Several years earlier, Kucharski and Joshi [2017] showed that the South Atlantic Ocean Dipole (SAOD) was already well represented in CMIP5 models and the new analyses thus confirm this good performance. The SAOD involves the SST variability in the eastern equatorial Atlantic and in the southwestern Atlantic, off the Argentina–Uruguay–Brazil coast [Nnamchi and Li, 2011; Nnamchi et al., 2011].

Tokinaga and Xie [2011] have revealed a weakening of the AEM variability during the last decades. They argued that this decrease over the period 1950 - 2009 was caused by the anthropogenic aerosol forcing, whose cooling effect in the northern Atlantic was greater than in the southern Atlantic. As a result, the SST meridional gradient over the tropical Atlantic weakened, and led to a relaxation of the equatorial trade winds and a deepened thermocline to the east. It was followed by a reduction of the upwelling of cold water from the deepest levels, which reinforces the positive SST trend and reduces the AEM variability. On the other hand, the exchange of the turbulent latent heat flux between the surface and the atmosphere represents an important negative feedback in the growing of the SST anomalies during the AEM events. Prigent et al. [2020] attributed the decreased variability of the AEM since 2000 to a weakening of the Bjerknes feedback in the equatorial Atlantic (like Tokinaga and Xie [2011]) and also to an increased cooling of the sea surface due to an increased latent heat flux release to the atmosphere. These authors found a reduced sensitivity of winds to the AEM in the western equatorial Atlantic basin, potentially due to a northward migration of the mean intertropical convergence zone (ITCZ), and a westward migration of the Walker circulation rising branch in the tropical Atlantic. In the eastern equatorial Atlantic, they found a reduced coupling between the surface and the thermocline.

Rainfall changes over the Guinea Coast during the past decades follow the observed decrease of the AEM variability [Tokinaga and Xie, 2011; Worou et al., 2020], underlying the strength of the AEM impact on the rainfall activity in this area

[Nnamchi et al., 2021a]. Thus, the question arises whether the Atlantic Niño will change in the future and what the implications for the tropical hydroclimate will be, particularly over the Guinea Coast. Mohino and Losada [2015] have shown an eastward displacement of the positive rainfall anomalies induced by the Atlantic Niño in the tropical Atlantic under a warmer climate. Another study has revealed a weakening of the relation between Atlantic Niño and El-Niño events in a warmer climate, due to a faster warming of the mid-troposphere relative to the low levels, leading to a more stabilized atmosphere and a reduced convection [Jia et al., 2019].

In the present study, we provide a more detailed analysis of the future changes in the Atlantic Niño and their impact on the rainfall over the tropical Atlantic and land masses in Guinea Coast by using results of GCMs obtained within CMIP6. The near-term, mid-term and long-term changes are analyzed separately. In Section 3.2, we describe the data and analysis strategy used in our study. Section 3.3 focuses on the performance of the models in simulating the Guinea Coast rainfall and the SST in the eastern equatorial Atlantic regions, as well as the SST pattern associated with the Atlantic Niño. In Section 3.4, we evaluate the rainfall patterns associated with the Atlantic Niño in the CMIP6 models over the last 30 years of the historical simulations conducted with these models. We highlight the group of models that simulate a realistic rainfall pattern associated with AEM over the Guinea Coast. Section 3.5 discusses the modelled future changes of the Atlantic Niño and their impact on the rainfall over the tropical Atlantic and Guinea Coast, as well as the associated mechanisms. In Section 3.6, we draw the main conclusions.

3.2 Data and methods

3.2.1 CMIP6 data

This study focuses on the July-August-September (JAS) season, during which the rainfall variability over the Guinea Coast and the entire West Africa peaks [Nnamchi et al., 2021a]. Outputs from 31 GCMs participating in CMIP6 are analyzed. Two kinds of simulations are considered: historical and high-emission scenario climate experiments. The historical climate simulations cover the 1850-2014 period, with observed natural and anthropogenic forcings. The climate projections are based on the Shared Socioeconomic Pathway 5, with a global radiative anthropogenic forcing that reaches $8.5 \text{ W} \cdot \text{m}^{-2}$ at the end of the 21st century (SSP5-8.5). The SSP5-8.5 simulations start in 2015 and assume a world with an intense industrial activity based on the fossil fuel and a large economic growth [O'Neill et al., 2016]. Only one member of each model is considered for each period. The description of the data is available in Table 3.1. The CMIP6

data are available on the Earth System Grid Federation (ESGF) portals (e.g. <https://esgf-node.llnl.gov/search/cmip6/>, last access: 18 March 2021).

3.2.2 Reanalyses data

The ERA5 dataset, a reanalysis product from the European Centre for Medium-Range Weather Forecasts [ECMWF, Hersbach et al., 2020], is used as a reference to evaluate the performance of the CMIP6 models in simulating the surface and atmospheric fields under present-day conditions. ERA5 has replaced ERA-Interim [Dee et al., 2011], which has been widely used to study the characteristics of the West African climate [Manzanas et al., 2014; Lavaysse et al., 2015; Kebe et al., 2016; Raj et al., 2018; Maranan et al., 2018; Diakhaté et al., 2019, 2020; Wainwright et al., 2019]. ERA5, for instance, shows an improvement of the rainfall representation over Burkina Faso [Tall et al., 2019], which is a part of the Sahelian region. ERA5 atmospheric and oceanic variables can be retrieved from the Copernicus Climate Data Store (<https://cds.climate.copernicus.eu/cdsapp#!/home>, last access: 30 July 2020). We also use the monthly sea surface height (SSH) data from the Ocean ReAnalysis System 5 [ORAS5, ECMWF, Zuo et al., 2019]. These SSH data can be retrieved from the Integrated Climate Data Center - ICDC at University of Hamburg (<https://icdc.cen.uni-hamburg.de/thredds/catalog/ftpthredds/EASYInit/oras5/catalog.html>, last access: 10 May 2021).

3.2.3 Analysis strategy

The evaluation of the present-day climate simulations is based on the last 30 years of the historical climate experiments (1985-2014). Three different periods are considered for the future climate analyses: the near-term (2015-2039), mid-term (2040-2069) and long-term (2070-2099) periods. Prior to any diagnostic over each period, the data are interpolated to the same grid of 1° resolution. Except for the analysis of the annual cycles and the seasonal means of the various surface and atmospheric fields, the climatological monthly mean is first removed from each dataset, for each considered period. The resulting anomalies are then linearly detrended and averaged over three months, July-August-September (JAS). We also found that there is no substantial difference between the residuals of the SST and rainfall indices, when the linear or the quadratic trends are removed (Figs. B.2, B.3, B.4, B.5). Therefore, for simplicity, we considered detrending linearly the different datasets. The choice of the JAS season is based on the fact that in the CMIP6 models, the Guinea Coast rainfall response to the June-July-August or July-August-September SST variability in the eastern equatorial Atlantic is maximum over JAS (Fig. B.6). Much of the statistics in this paper are based on linear

regressions onto the standardized JAS SST index of the Atlantic Niño. This index is defined as an average of SST anomalies over the ATL3 region [Table 3.2, Zebiak, 1993]. The GISS-E2-1-G model is discarded from the analyses. This model shows no significant rainfall response to the Atlantic Niño over the tropical Atlantic and the West-Africa, in contrast to ERA5. Overall, our results remain unchanged if we apply a composite analysis instead of linear regressions (Fig. B.1, compared to Fig. 3.8).

The difference of the wind divergence between 200 hPa and 850 hPa levels (DIV200/850) is used to evaluate the direction and intensity of the vertical air motion [Gómara et al., 2017; Diakhaté et al., 2019]. Positive values of this difference (upper level minus low level) mean that there is a low level convergence of the wind, an ascent and a divergence at high level. The spread of the 30 GCMs is evaluated with the Interquartile range (IQR) metric. Similar to Jia et al. [2019], a sign-dependent average (SDA) of the rainfall regression pattern is utilized to select the models, based on their robust rainfall response to the Atlantic Niño over the Guinea Coast. This technique is described in three steps. First, the JAS rainfall anomalies are regressed onto the standardized JAS ATL3 SST index. Secondly, the rainfall regression coefficients that are significant at 95% confidence level (two-sided Student test) are averaged over the Guinea Coast. This gives a whole average coefficient. Thirdly, over Guinea Coast, only significant regression coefficients that have the sign of the whole average coefficient are averaged. This gives the sign-dependent average coefficient. The sign-dependent average is set to zero if there is no significant regression coefficient.

For a given period, when the regression patterns are averaged over some subsets of models, we determine the robustness of the result by adapting the method of Tebaldi et al. [2011]. The signal in areas where at least 50 % of the models show a significant regression (two-sided Student test at 95 % confidence level) and where at least 80 % of the models agree on the sign of the mean is considered as robust. When we consider the change between two periods, the sign of the change averaged over a subset of models is considered as robust if at least two thirds of the models agree on the sign of the multimodel mean [Rehfeld et al., 2020], and at least the change of the multimodel mean is significant at 95 % confidence level according to a two-sided Welch t-test. Moreover, the percentage of change of the ATL3 SST index standard deviation between two periods is computed as $100 \times \frac{\sigma_{fut} - \sigma_{his}}{\sigma_{his}}$, where σ_{his} is the standard deviation of the JAS ATL3 SST index in the 1985-2014 period, and σ_{fut} the standard deviation of the JAS ATL3 SST index in a future period (the near-term, mid-term and long-term periods). Finally, coordinates of different geographic areas used in the present study are defined in Table 3.2. In particular, the tropical Atlantic box 1 (TAB1) and 2 (TAB2) are large areas over which we validate the SST and rainfall patterns related to the Atlantic Niño.

Moreover, we define different groups of models based on their rainfall SDA over a region. We aim to understand if different groups of models simulate the rainfall pattern related to the Atlantic Niño over the equatorial Atlantic and the Guinea Coast in different ways and in particular if a different simulation of the current state has some implications on the simulated future changes in rainfall patterns. We also aim to highlight the differences in the key physical mechanisms between the groups. Focusing on the Guinea Coast for example, we first identify the climate models which are able to realistically simulate the observed rainfall pattern related to the Atlantic Niño in the Guinea Coast over the past decades (the group GC+). In observations, a positive rainfall anomaly over Guinea Coast is related to a warm phase of the Atlantic Niño (and vice versa). The group GC- indicates models which present a negative rainfall pattern associated with a warm phase of the Atlantic Niño. For the future changes in the AEM-related rainfall pattern over Guinea Coast, we separated models which present an enhancement of the positive rainfall pattern (GC++) from models which present a weakening of the rainfall pattern (GC+-). Then we tried to understand the reasons of these changes and the differences between the different categories. A similar argument is applied to the ocean.

3.3 Evaluation of the performance of the GCMs in simulating the rainfall in Guinea Coast, SST in ATL3 region, and JAS SST pattern related to the Atlantic Niño

3.3.1 Rainfall in Guinea Coast

The observed Guinea Coast rainfall annual cycle is characterized by a bimodal structure with two maxima in June and September respectively. Much of the CMIP6 models overestimate the magnitude of the rainfall and rainfall standard deviation in this region during the boreal summer (Fig. B.7). Unlike ERA5, the CMIP6 ensemble mean (EnsMean) depicts a plateau of $9 \text{ mm} \cdot \text{day}^{-1}$ in July and August, when much of the wet biases occur. Over the Guinea Coast, the positive JAS mean rainfall bias is present in 19 GCMs. More specifically, there are 24 models with a wet bias between 10°W and 0°E , against 14 models in the western (-20°W to -10°W) and eastern (0°E - 10°E) Guinea Coast (not shown).

The CMIP5 models are also known to overestimate the rainfall in coastal areas of West Africa during the "little dry season" [Sow et al., 2020]. Wainwright et al. [2019] have demonstrated that the misrepresentation of the July-August Guinean coast rainfall in these models comes from the positive SST biases in the Atlantic Ocean, which strengthen the rising motions over the Guinea Coast and increase the rainfall. Consistently, the CMIP6 ensemble mean exhibits a positive (negative)

Table 3.1: CMIP6 models and members of historical and SSP5-8.5 simulations. (*) indicates models for which the sea surface height variable was not yet available at the time of this study. Models for which, on the one hand, the sea level pressure and the 10 m horizontal wind variables, and, on the other, the mixed layer depth variable were not used for the mean state change analyses are indicated by (**) and (***), respectively. The other variables used are: rainfall, sea surface temperature, 850 hPa specific humidity, 850 hPa and 200 hPa horizontal wind components.

CMIP6 model	Historical member	SSP5-8.5 member
ACCESS-CM2	r1i1p1f1	r1i1p1f1
ACCESS-ESM1-5	r1i1p1f1	r1i1p1f1
BCC-CSM2-MR	r1i1p1f1	r1i1p1f1
CAMS-CSM1-0	r1i1p1f1	r1i1p1f1
CanESM5	r1i1p1f1	r1i1p1f1
CanESM5-CanOE	r1i1p2f1	r1i1p2f1
CESM2 (*, **, ***)	r1i1p1f1	r2i1p1f1
CESM2-WACCM (**)	r1i1p1f1	r1i1p1f1
CNRM-CM6-1	r1i1p1f2	r1i1p1f2
CNRM-CM6-1-HR	r1i1p1f2	r1i1p1f2
CNRM-ESM2-1	r1i1p1f2	r1i1p1f2
EC-Earth3	r1i1p1f1	r1i1p1f1
EC-Earth3-Veg	r1i1p1f1	r1i1p1f1
FGOALS-f3-L (*)	r1i1p1f1	r1i1p1f1
FIO-ESM-2-0 (**, ***)	r1i1p1f1	r1i1p1f1
GFDL-ESM4 (*, ***)	r1i1p1f1	r1i1p1f1
GISS-E2-1-G	r1i1p1f2	r1i1p1f2
HadGEM3-GC31-LL	r1i1p1f3	r1i1p1f3
INM-CM4-8 (***)	r1i1p1f1	r1i1p1f1
INM-CM5-0 (***)	r1i1p1f1	r1i1p1f1
IPSL-CM6A-LR	r1i1p1f1	r1i1p1f1
KACE-1-0-G (*, **, ***)	r1i1p1f1	r2i1p1f1
MCM-UA-1-0 (*, ***)	r1i1p1f2	r1i1p1f2
MIROC6 (**)	r1i1p1f1	r1i1p1f1
MIROC-ES2L (***)	r1i1p1f2	r1i1p1f2
MPI-ESM1-2-HR	r1i1p1f1	r1i1p1f1
MPI-ESM1-2-LR	r1i1p1f1	r1i1p1f1
MRI-ESM2-0	r1i1p1f1	r1i1p1f1
NorESM2-LM (**)	r1i1p1f1	r1i1p1f1
NorESM2-MM (**)	r1i1p1f1	r1i1p1f1
UKESM1-0-LL (**)	r1i1p1f2	r1i1p1f2

Table 3.2: Coordinates of different domains.

Domain	Lon min (°)	Lon max (°)	Lat min (°)	Lat max (°)
West Africa box (WAB)	−20	10	4	20
Guinea Coast box (GCB)	−20	10	4	10
ATL3	−20	0	−3	3
Equatorial Atlantic box (EAB)	−30	10	−5	5
Tropical Atlantic box 1 (TAB1)	−70	10	−20	20
Tropical Atlantic box 2 (TAB2)	−70	10	−5	15

mean SST bias in the eastern (western) part of the tropical Atlantic, as depicted in Fig. 3.1. This figure also shows low-level anomalous northerlies that reinforce the convergence over the Guinea Coast. Analysis of the 10 m wind biases shows anomalous westerlies to the west of the warm bias area (Fig. B.8). This result is consistent with the conclusions obtained from Richter and Tokinaga [2020], who analyzed the CMIP6 pre-industrial control simulations.

In general, the JAS spatial structure of the West African rainfall is well reproduced by the models (Fig. B.9 (a)). The spatial correlation between the mean JAS rainfall in the 30 GCMs and that of ERA5 lies between 0.68 and 0.96 (Fig. B.9 (b)). The multimodel ensemble mean of the JAS rainfall also shows a good performance, with a correlation of 0.92 with ERA5. Moreover, the standard deviation of the JAS West-African rainfall ranges between 2.79 and 5.75 mm · day^{−1} for the 30 GCMs. The corresponding value for the multimodel ensemble mean is 3.76 mm · day^{−1}, which corresponds to 93% of the spatial rainfall variability in ERA5.

3.3.2 SST in the eastern equatorial Atlantic

In ERA5, the annual cycle of the SST in the ATL3 area (Fig. B.10 (a)) shows a cold tongue which develops from April when the mean SST over the ATL3 region is about 29 °C and reaches its lowest value in August, around 24 °C. The CMIP6 EnsMean overestimates the SST annual cycle in this region, with a pronounced warm bias in June-September (Fig. 3.1). The correlation between the annual cycle values in each model with ERA5 ranges from 0.88 to 0.99 (Fig. B.10 (c)). This result indicates that the phasing of the SST in the ATL3 region is relatively well simulated by the models. Moreover, the CMIP6 ensemble mean of the monthly variability of the ATL3 SST index peaks in June-July-August, and is delayed one month relative to ERA5 (Fig. B.10 (b)). There is a second peak in November, corresponding to the winter Atlantic Niño which may have impacted the ENSO events and the rainfall in South America, mainly during a few decades in the mid-

20th century [Okumura and Xie, 2006; Hounsou-Gbo et al., 2020]. Throughout the year, the monthly SST standard deviation in the models covers a wide range of values, with different phasings (Fig. B.10 (d)).

3.3.3 JAS SST pattern associated with the Atlantic Niño

The spatial SST pattern characteristic of the summer Atlantic Niño is derived by regressing the JAS SST anomalies onto the standardized JAS ATL3 SST index (Fig. 3.2). The SST pattern in ERA5 shows positive anomalies along the equator and off coastal areas in the southeastern Atlantic Ocean. Most of the positive SST anomalies vary between 0.1 and 0.4°C. The CMIP6 models exhibit various SST imprints related to the Atlantic Niño over the tropical Atlantic (the TAB1 region, see Table 3.2). Some models display a uniform sign of the significant SST anomalies over TAB1 (e.g., CAMS-CSM1-0, CESM2, FIO-ESM-2-0). In addition to the warming in the equatorial Atlantic, other models present negative SST anomalies in the tropical north Atlantic (e.g., ACCESS-ESM1-5), southwest Atlantic (e.g., HadGEM3-GC31-LL), or both north and southwest Atlantic (e.g., ACCESS-CM2, EC-Earth3, MIROC-ES2L).

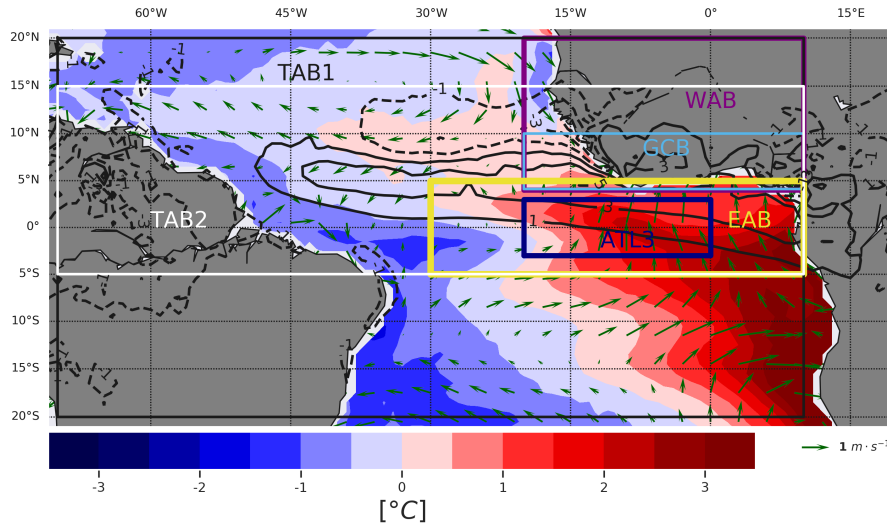


Figure 3.1: Mean biases (relative to ERA5) of the ensemble mean of the 30 GCMs for the JAS SST (in colors), rainfall (in black contours) and 850 hPa wind (arrows) over 1985-2014. Boxes in black, white, yellow, blue, cyan and magenta corresponds to the TAB1, TAB2, EAB, ATL3, GCB and WAB regions, respectively.

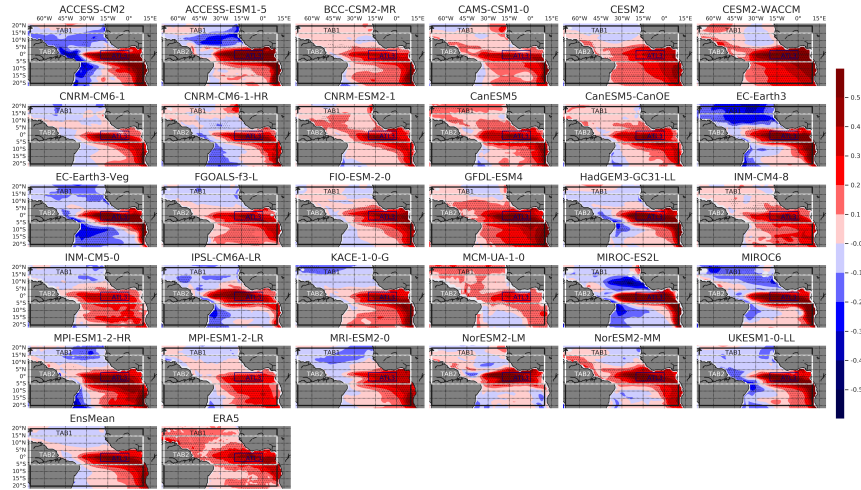


Figure 3.2: Regression maps of the JAS SST anomalies onto the standardized JAS ATL3 SST index. Stippling in the EnsMean indicates grid points where more than 50 % of the models show significant regression at 95 % confidence level and more than 80 % of the models agree on the sign of the regression coefficient. Stippling in each model and ERA5 indicates significant regression coefficients at 95 % level.

Overall, the performance of the models in representing the AEM SST pattern is relatively good (Fig. 3.3). The EnsMean SST pattern is highly correlated with the SST pattern in ERA5, with a correlation value of 0.8. Except for GISS-E2-1-G (not shown), the spatial correlation between the AEM SST pattern in each model and ERA5 over the TAB1 region ranges from 0.5 to 0.9. The standard deviation of the spatial SST anomalies related to the AEM in the TAB1 region amounts 0.1°C in ERA5. In 29 models, this spatial variability of the anomalous SST pattern is 0.01 to 0.2°C higher than that in ERA5. In contrast, the spatial variability of the AEM-SST pattern in MCM-UA-1-0 is 0.02°C lower than that in ERA5.

3.4 Evaluation of the performance of the GCMs in simulating the JAS rainfall pattern related to the Atlantic Niño

Figure 3.4 displays the regression maps of the JAS rainfall anomalies onto the standardized JAS ATL3 SST index, computed for the 1985-2014 period. In the reanalysis ERA5, the Atlantic Niño positive phases limit the northward progression of the West African monsoon flow, which leads to an anomalous increase of the rainfall over the Guinea Coast. The positive rainfall pattern associated with the Atlantic Niño extends to the ocean, from the equator up to 10°N and from the

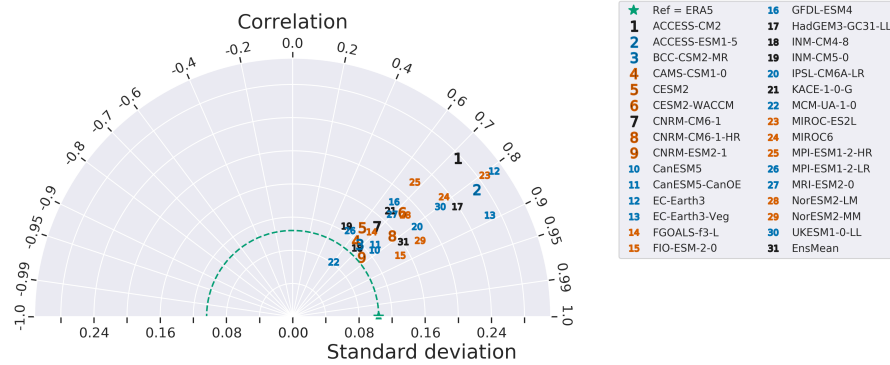


Figure 3.3: Taylor diagram of the JAS SST pattern in Fig. 3.2 over the tropical Atlantic box 1, where ERA5 is used as the reference. Models in blue (orange) simulate a long-term increase (decrease) in the Guinea Coast rainfall response to the AEM.

West African coast to 45°W . Moreover, while the rainfall anomalies in ERA5 are generally positive over the ocean in the TAB2 area (see Table 3.2), a majority of the models depict a dipolar structure of the rainfall pattern, consistent with Mohino and Losada [2015], among others. In these models, negative rainfall anomalies are present north of the band of positive values, between 5 and 15°N .

Over the West African region, the latitudinal position of the maximum of the rainfall anomalies related to the AEM and averaged between 20°W and 10°E is 6.5°N in ERA5 (Fig. B.11). In the CMIP6 Ensmean, this position is relatively good and is located at 4.5°N . Over West Africa (the WAB, see Table 3.2), the performance of the GCMs is poor to modest (Fig. 3.5 (a)). The spatial correlation with ERA5 lies between -0.5 and 0.5 (0.5 for the EnsMean). Over the oceanic area (the TAB2), the position of the maximum of the AEM related rainfall anomalies averaged between 70°W and 10°E in the CMIP6 EnsMean is located at 3.5°N , which is two degrees below the position in ERA5. Figure 3.5 (b) indicates that the models show a poor to modest spatial correlation with ERA5, which ranges from -0.3 to 0.5 (0.3 for the EnsMean).

The sign-dependent average of the 1985-2014 rainfall anomalies related to the AEM over the Guinea Coast box is positive in the reanalysis ERA5 (Fig. B.12 (a)). Two groups of GCMs are considered, according to their sign-dependent average over the GCB. The first group is termed GC+, and defines models with a positive sign-dependent average. It includes 24 models which are able to reproduce an increased rainfall associated with the Atlantic Niño positive phases (Fig. 3.6 (a)). In contrast, there are 6 models in the second group, termed GC-, which show

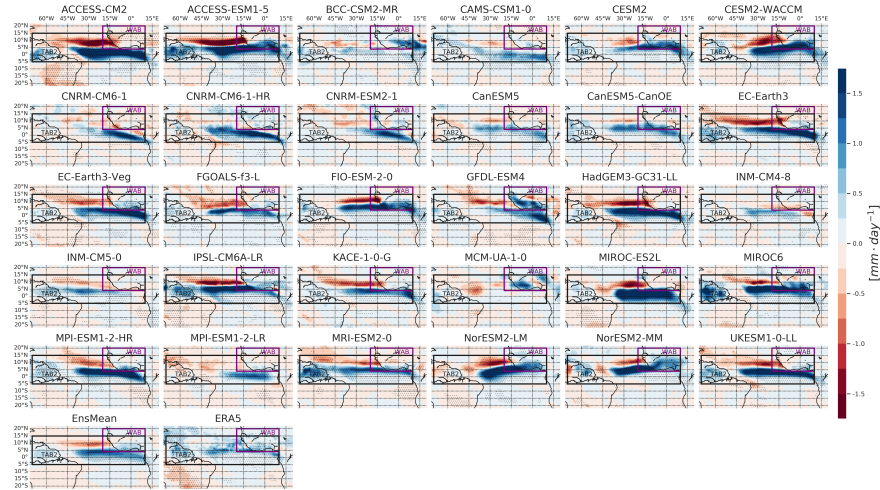


Figure 3.4: (a) Regression maps of the JAS rainfall anomalies onto the standardized JAS ATL3 SST index over 1985-2014. Stippling in the EnsMean indicates grid points where more than 50 % of the models show significant coefficients at 95 % level and more than 80 % of the models agree on the sign of the regression coefficient. Stippling in each model and ERA5 indicates significant regression coefficients at 95 % level.

negative correlations between the rainfall over the Guinea Coast and the Atlantic Niño (Fig. 3.6 (b)). The GC- ensemble mean rainfall response over the GCB is weak in magnitude, with less model agreement compared to the GC+ ensemble mean. In both groups, the dipolar structure of the AEM rainfall pattern is similar over the tropical Atlantic (the TAB2 area). However, this dipolar structure is not present in ERA5 (Fig. 3.6 (d)).

During positive phases of the Atlantic Niño, warmer than normal sea surface in the eastern equatorial Atlantic weakens the zonal surface pressure gradient over the equatorial Atlantic. This in turn weakens the prevailing trade winds. The regression of the low-level zonal component of the wind onto the ATL3 SST index is used to evaluate the first component of the Bjerknes feedback, which is the forcing of the surface wind in the western basin of the Atlantic Ocean by SST in the eastern basin [Keenlyside and Latif, 2007]. Next, these anomalous westerlies increase the surface convergence above the warm waters in the east, which leads to a rising of the sea surface height, an increased heat content and a deepening of the thermocline. This is the second component of the Bjerknes feedback. Hence, the deepening of the eastern equatorial Atlantic thermocline reduces the influence of the upwelling of cold subsurface water on the surface temperature, which then reinforces the initial surface warming. This is the third component of the Bjerk-

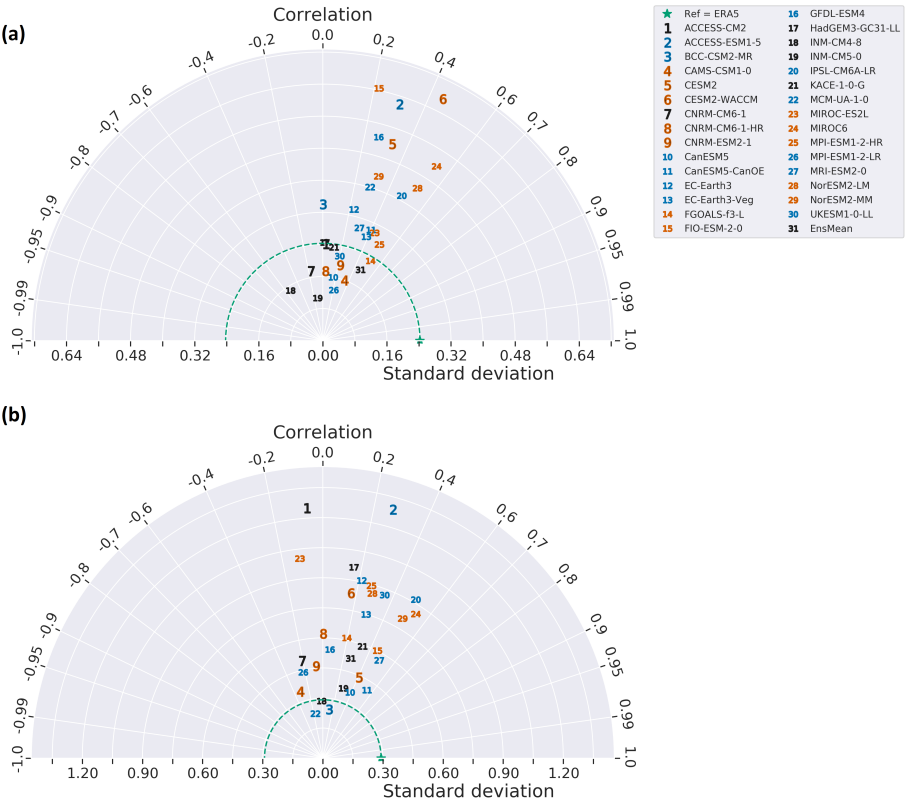


Figure 3.5: (a) Taylor diagram of the JAS spatial rainfall pattern obtained in Fig. 3.4 over the West African box (magenta box in the maps in Fig. 3.4). (b) Same as in (a) but for the tropical Atlantic box 2 (black box in the maps in Fig. 3.4). ERA5 is used as the reference in (a) and (b). Models in blue (orange) simulate a long-term increase (decrease) in the Guinea Coast rainfall response to the AEM.

nes feedback. The regression of the sea surface height anomalies, a proxy for the thermocline depth, onto the ATL3 SST index is a measure of the third component of the Bjerknes feedback. This diagnostic is also influenced by the second component, but this point is not explicitly analysed in this study.

The JAS sea surface temperature and sea surface height imprints related to the standardized JAS ATL3 SST index are similar in the GC+ and GC- ensemble means (Fig. 3.6 (e)-(f), (u)-(v)). In the eastern equatorial Atlantic, the core of the anomalous warming is relatively well represented in both GC+ and GC-, compared to ERA5 (Fig. 3.6 (h)). Contrastingly, the positive SSH anomalies in GC+ and GC- are stronger than in ORAS5 (Fig. 3.6 (x)). In the western boundaries of the basin and along the 5°N - 10°N band, the negative SSH anomalies are higher in GC+ and GC- than in ORAS5. In association with the AEM, there is a stronger zonal moisture flux from the equatorial Atlantic toward the Guinea Coast in GC+ than in GC-. This could partially explain the rainfall response difference between the two groups (Fig. 3.6 (i)-(j)). The northward extension of the positive zonal moisture flux is limited to 10°N in GC+ and 5°N in GC-, while it reaches 15°N in ERA5 (Fig. 3.6 (l)). The positive anomalous zonal moisture flux is limited to the oceanic area in GC-, whereas it extends to the Guinea Coast in GC+ and covers both the Guinea Coast and part of the Sahelian region in ERA5. Off the Guinea Coast, the meridional component of the moisture flux is also slightly greater in GC+ than in GC-, but the difference is lower than that in the zonal component (Fig. 3.6 (m)-(n)). However, the positive meridional moisture flux over the equatorial and South Atlantic is weaker in both GC+ and GC-, compared to ERA5 (Fig. 3.6 (p)).

Finally, during positive phases of the Atlantic Niño, over the Guinea Coast and the equatorial Atlantic, the atmospheric convection is more enhanced in the GC+ group than in GC-. This is indicated by the more pronounced wind divergence difference between the 200 hPa and 850 hPa levels in Fig. 3.6 (q)-(r). Moreover, in GC+ and GC-, the oceanic areas where the atmosphere is destabilized are located 5° south of their positions in ERA5 (Fig. 3.6 (t)). This explains why the positive rainfall anomalies associated with the positive phases of the Atlantic Niño in the GCMs are located south of the obtained position in ERA5. In conclusion, the combination of a large moisture flux from the equatorial Atlantic toward the Guinea Coast, and a more destabilized atmosphere over the Guinea Coast leads to an enhanced rainfall response to AEM in the GC+ models compared to GC- models. Moreover, according to Nnamchi et al. [2021b], the rainfall response to the AEM is maximum at the location of the mean Intertropical convergence zone (ITCZ). The CMIP6 models present a spurious southward position of the mean ITCZ position relative to the observations during the boreal summer. This induces an enhancement of the coupling between the atmosphere and the ocean during the growing phase of the Atlantic Niño in the models, which is located south of the maximum rainfall response in ERA5. Note that the aspects of the 30 GCMs

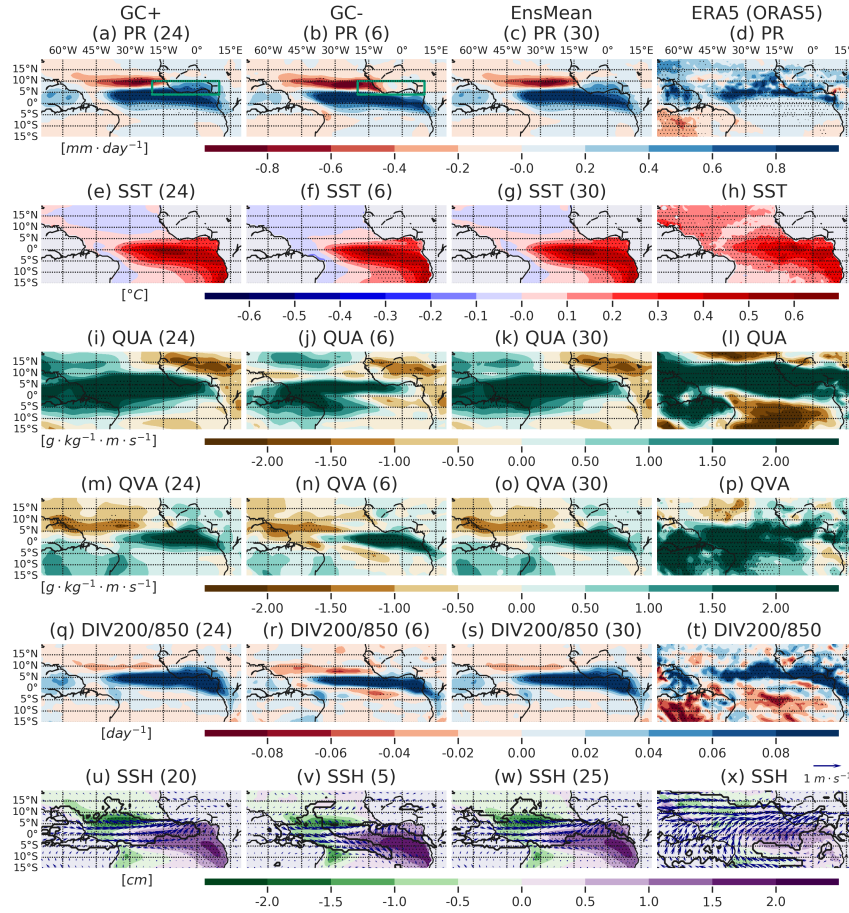


Figure 3.6: Regression maps of the JAS rainfall (a,b,c,d), SST (e,f,g,h), 850 hPa zonal moisture flux (i,j,k,l), 850 hPa meridional moisture flux (m,n,o,p), horizontal wind divergence difference between 200 hPa and 850 hPa (q,r,s,t) and sea surface height (u,v,w,x) anomalies associated with the standardized JAS ATL3 SST index over the 1985–2014 period. Results for the GC+, GC- and EnsMean categories are displayed, and the number of models in each category is indicated in parentheses. GC+ (GC-) is a group of models with a positive (negative) sign-dependent average of the rainfall associated with the Atlantic Niño over the Guinea Coast (green boxes in (a) and (b)). Results for the reanalyses are displayed in the rightmost column, and stippling represents significant regressions at 95 % level. Stippling in the other maps indicates areas where the regression coefficients are significant at 95 % confidence level for at least 50 % of the models in each group, and where more than 80 % of the models agree on the sign of the regression coefficient. In (u)–(x), the JAS 850 hPa horizontal wind related to the standardized JAS ATL3 SST index is superimposed on the SSH pattern, and contours represent areas where at least the zonal or the meridional component of the wind is significant.

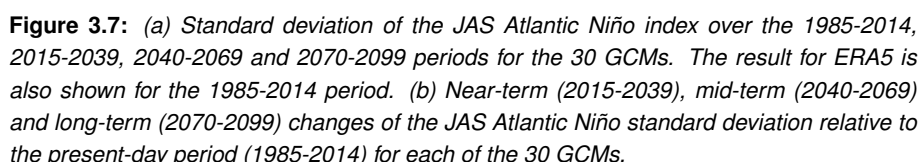
EnsMean patterns are very close to the ones of the GC+ (Fig. 3.6 (c), (g), (k), (o), (s), (w)).

3.5 Future impact of the Atlantic Niño on the rainfall over the tropical Atlantic and Guinea Coast

3.5.1 Overview of the near-term, mid-term and long-term Atlantic Niño changes

In this section, the impact of the Atlantic Niño on the tropical Atlantic and Guinea Coast is evaluated in a climate with a high anthropogenic emission of greenhouse gases. The standard deviation of the Atlantic Niño index in the present-day climate simulations varies between 0.24 and 0.65 °C (Fig. 3.7 (a)), against 0.38 °C in ERA5. Relative to the 1985-2014 period, the standard deviation of the Atlantic Niño index shows a change that ranges from -44 % to 48 % for the near-term period, -41 % to 16 % for the mid-term period and -49 % to 11 % for the long-term period (Fig. 3.7 (b)). The average of the relative changes in the 30 GCMs amounts to -6 %, -13 % and -22 % for the near-term, mid-term and long-term periods, respectively. Among the 30 GCMs considered, 18, 23, and 27 agree on the reduction of the Atlantic Niño variability for the near-term, mid-term and long-term periods, respectively. Interestingly, these results are opposite to the findings of Brierley and Wainer [2018], who evaluated the change of the AEM variability in a quadrupled CO₂ experiment with CMIP5 models. They found that 3 (12) GCMs out of 15 depict a decrease (an increase) of the Atlantic Niño variability. Among other reasons, we can postulate on the differences in the models between the CMIP5 and CMIP6 to explain the differences between the two results. Brierley and Wainer [2018] compared a 1% per year quadrupled CO₂ experiment to a pre-industrial control simulation of CMIP5, which is different from the simulations compared in our analysis (historical and SSP5-8.5 simulations). A more direct comparison between the two studies could be performed by analyzing the AEM variability changes between CMIP6 1pctCO₂ and the CMIP6 pre-industrial simulations.

In the present-day climate, the multimodel mean of the JAS SST anomalies associated with the JAS ATL3 SST index and averaged over the equatorial Atlantic (the EAB region) is 0.33 °C (Fig. 3.8 (a)). This value is consistent with the reanalysis ERA5. For the periods 2015-2039, 2040-2069 and 2070-2099, the EnsMean value is reduced to 0.30, 0.28 and 0.27 °C, respectively. This result indicates that, relative to the present-day climate, the multimodel mean of the SST response to the Atlantic Niño has gradually decreased, with a percentage of change equals to -7 %, -13 % and -18 % for the near-term, mid-term and long-term periods, respectively (Fig. 3.8 (a)). There are 73 % and 87 % of the 30 GCMs which agree on



Over the period 1985-2014, the JAS CMIP6 EnsMean rainfall anomalies related to one standard deviation of the JAS ATL3 SST index and averaged over the EAB is $0.62 \text{ mm} \cdot \text{day}^{-1}$, which is greater than the observed value ($0.21 \text{ mm} \cdot \text{day}^{-1}$ in ERA5). Subsequent to the weakening of the equatorial SST anomalies in the future, the rainfall associated with the Atlantic Niño over the EAB area decreases (Fig. 3.8 (b)). The CMIP6 EnsMean of the anomalous EAB rainfall values for the three consecutive future periods are 0.54, 0.46 and $0.40 \text{ mm} \cdot \text{day}^{-1}$, respectively. The corresponding EnsMean rainfall change relative to the present-day period is about -12 %, -25 % and -36 %, with an agreement of 63 %, 83 % and 80 % of the models on the sign of the change in the near-term, mid-term and long-term periods, respectively.

Figures B.13 (a) and (c) indicate that, over the Atlantic Ocean, in the four periods of the study, the EnsMean position of the maximum rainfall anomalies associated with AEM remains close to 4°N , but with a linear decrease of the rainfall

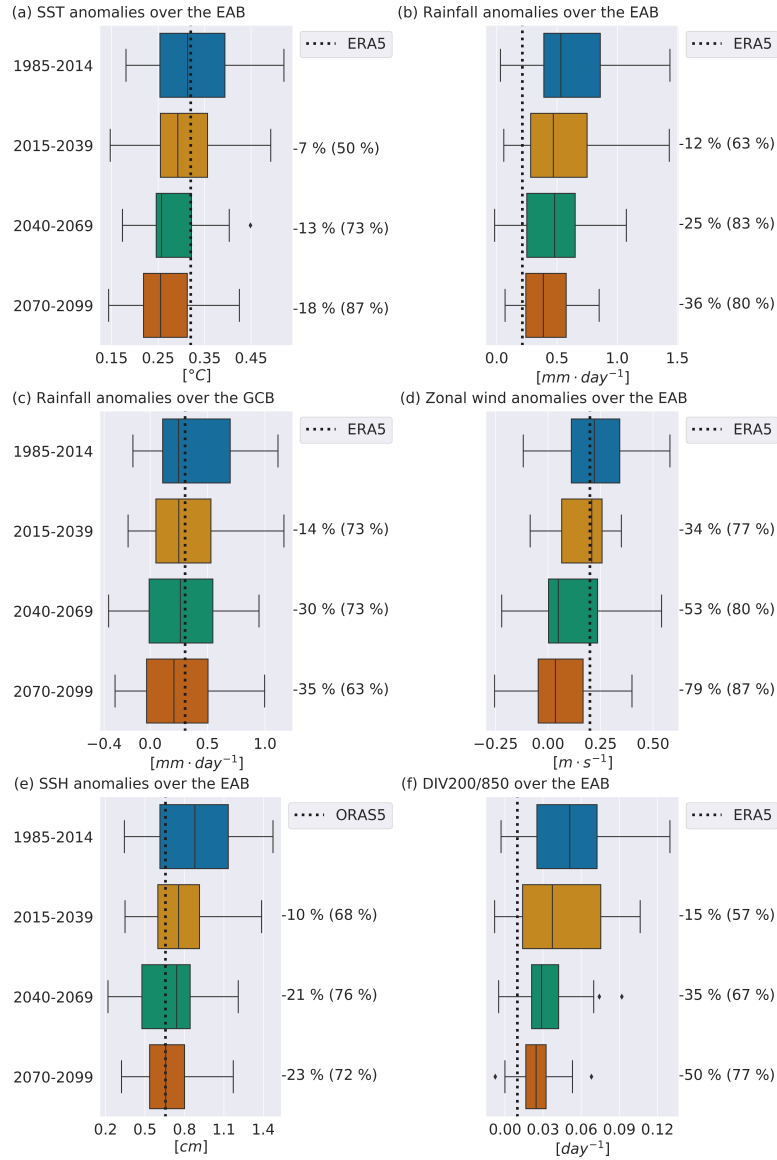


Figure 3.8: Boxplots of the JAS regression coefficients of (a) SST, (b) rainfall, (d) 850 hPa zonal wind, (e) sea surface height and (f) horizontal wind divergence difference between 200 hPa and 850 hPa (DIV200/850) anomalies averaged over the equatorial Atlantic box. Boxplot of the regression coefficients of the rainfall anomalies averaged over the Guinea Coast (c). The variables from each of the 30 models are regressed onto the standardized JAS ATL3 SST index, for four different periods: 1985-2014, 2015-2039, 2040-2069 and 2070-2099. Each boxplot represents the distribution of the 30 regression coefficients averaged over the specified area in the different models. Vertical lines represent the corresponding regression values in ERA5 (ORAS5) for the 1985-2014 period.

intensity with time. North of 5°N , an upper level subsidence produces negative rainfall anomalies in the tropical North Atlantic, which gradually weaken in the future periods. Over the EAB region, the decrease of the rainfall interquartile range in the CMIP6 models for the mid and long-term periods indicates an increased consistency of the rainfall response among the models. For the present-day, the near-term, the mid-term and the long-term periods, the IQR values are 0.47, 0.47, 0.40 and $0.34 \text{ mm} \cdot \text{day}^{-1}$, respectively.

Regarding the AEM induced rainfall over the Guinea Coast, an overall reduction of the multimodel mean is projected (Fig. 3.8 (c), Figs. B.13 (b) and (d)). The CMIP6 EnsMean rainfall averaged over the Guinea Coast (the GCB area) decreases from a value of $0.37 \text{ mm} \cdot \text{day}^{-1}$ in the period 1985-2014 to $0.24 \text{ mm} \cdot \text{day}^{-1}$ at the end of the 21st century. The corresponding values for the near-term and mid-term periods are 0.32 and $0.26 \text{ mm} \cdot \text{day}^{-1}$, respectively. The amount of the GCB EnsMean rainfall reduction reaches thus 14 %, 30 % and 35 % in the near-term, mid-term and long-term periods, respectively. The percentages of the models that agree on a reduction of the rainfall magnitude associated with AEM over GCB for the three periods are 73 %, 73 % and 63 %, respectively. This means that there is a lesser agreement on the projected mid and long-term AEM-rainfall changes over the Guinea Coast than over the equatorial Atlantic. Moreover, there is a slight increase of the inter-model consistency in the projected long-term AEM-rainfall over the Guinea Coast.

The positive loop of the Bjerknes feedback requires first the forcing of the eastern equatorial Atlantic SST on the surface winds in the western basin. A warming in the eastern equatorial Atlantic gives rise to anomalous westerlies in the western basin. In turn, these weaker trade winds induce a deepening of the eastern equatorial Atlantic thermocline, which can be measured by a rising of the sea surface height. This is the second element of the Bjerknes feedback. The increased thermocline depth in the eastern equatorial Atlantic reduces the upwelling of cold water in that area, and strengthens the initial warming. This third component of the Bjerknes is characterized by a positive correlation between the SST and SSH in the eastern equatorial Atlantic.

In the future climate, following the SSP5-8.5 scenario, the variability of the trade winds associated with the Atlantic Niño decreases, as shown in Figs. 3.8 (d) and B.13 (e), where the 850 hPa zonal winds are regressed onto the standardized ATL3 SST index and averaged over the equatorial Atlantic (the EAB region). The CMIP6 EnsMean of the 850 hPa zonal wind anomalies corresponding to one standard deviation of the ATL3 SST index and averaged over the EAB amounts to $0.24 \text{ m} \cdot \text{s}^{-1}$ in the 1985-2014 period, which is close to the value derived from ERA5 ($0.2 \text{ m} \cdot \text{s}^{-1}$). For the near-term, mid-term and long-term future, this value decreases to 0.16, 0.11 and $0.05 \text{ m} \cdot \text{s}^{-1}$, respectively. This reduction corresponds to 34 %, 53 % and 79 % of the value obtained in the present-day climate, respec-

tively, with a high agreement among the 30 CMIP6 models (77 % for the near-term, 80% for the mid-term periods, and 87 % for the long-term period). This result suggests a weakening of the first component of the Bjerknes feedback, which links anomalous westerlies to the abnormal warming in the eastern equatorial Atlantic.

In response to weakened trade winds during a positive AEM phase, the upwelling of seawater in the east is reduced, leading to an anomalous rising of the SSH in the east, and an SSH fall in the west (Fig. 3.8 (e) and Fig. B.13 (f)). The weakening of the anomalous westerlies associated with one standard deviation of the ATL3 SST index in the future reduces the positive (negative) SSH anomalies in the eastern (western) equatorial Atlantic, which in turn reduces the SST anomalies over the ATL3 area. The CMIP6 EnsMean average of the SSH anomalies related to the standardized ATL3 SST index over the EAB region for the four consecutive periods is 0.87, 0.78, 0.69 and 0.67 cm, respectively. The corresponding reduction of the SSH EnsMean in the EAB for the three consecutive future periods is 10 %, 21 % and 23 %. The CMIP6 models agreement on the sign of this change is between 68 % and 76% in the three cases. This result demonstrates a weakening of the third element of the Bjerknes feedback. In addition, the spread of the models is also reduced at the end of the 21st century, highlighting the increased consistency of the models for the SSH responses to the Atlantic Niño.

Part of the weakening of the equatorial Atlantic rainfall variability associated with the Atlantic Niño in a warmer climate has been attributed to a faster warming of the mid-tropical Atlantic troposphere compared to the surface, as pointed out by Jia et al. [2019]. This implies that the troposphere above the equatorial Atlantic becomes more stable and limits the convection. This is in agreement with Figure 3.8 (f) that shows, over the equatorial Atlantic, a gradual decrease of the strength of the rising motion associated with the Atlantic Niño warm phases, during the near-term, mid-term and long-term periods. For the three successive future periods, the relative change (and the agreement between models) of the CMIP6 EnsMean DIV200/850 averaged over the EAB area is -15 % (57 %), -35 % (67 %) and -50 % (77 %), respectively. The consistency of the DIV200/850 anomalous response in the CMIP6 models is significantly enhanced in the mid-term and long-term periods.

Moreover, as the rainfall patterns related to the Atlantic Niño could be a mix from local and remote SST drivers, the monthly Niño3 indices (average of SST anomalies over 5 °S - 5 °N, 150 °W - 90 °W) are regressed onto the standardized JAS ATL3 SST index for the models and ERA5 (Fig. B.14). In ERA5, the JAS ATL3 SST index is negatively correlated with the Niño3 index from April to December. This opposite phase relationship is stronger in November and December. From January to March, an in-phase relationship is observed. The CMIP6 models ensemble mean response shows an anticorrelation between the JAS ATL3 SST index and the Niño3 index for all the months during 1985-2014. Thus, the effect of

SST in both basins would lead to rainfall anomalies of the same sign over Guinea Coast. In the future, a general decrease of the Niño3-ATL3 relationship is obtained in the 2015-2039 and 2040-2069 period. However, the sign of the correlation between both indices is reversed in the long-term period (2070-2099). This means that in the 2070-2099 period, two rainfall anomalies with opposite signs will interact over Guinea Coast, and this might reduce the rainfall amplitude associated with the Atlantic Niño. For the 1985-2014, 2015-2039, 2040-2069 and 2070-2099 periods, there are, respectively, 9, 7, 9 and 7 models which show a significant correlation between the JAS Niño3 and ATL3 SST indices (Fig. B.15).

3.5.2 Change of the Atlantic Niño impact on the rainfall in Guinea Coast

Throughout this section, we investigate the JAS rainfall, SST, 850 hPa zonal wind and moisture flux associated with the standardized JAS ATL3 SST index. To highlight the impact of some specific differences between the models in their way of representing the spatial characteristics related to AEM in response to climate change, six different subsets of the CMIP6 models are considered. They are also based on the sign-dependent average of the AEM related rainfall pattern over, on the one hand, the Guinea Coast (the GCB area) and, on the other hand, the equatorial Atlantic region (the EAB area). First, the GC+ group (the 24 models in Sect. 3.4 which simulate a realistic JAS GCB rainfall associated with one standard deviation of the JAS ATL3 SST index) is split into two groups. One group, termed GC+-, projects a decrease of the GCB rainfall intensity associated with the Atlantic Niño. The other group, termed GC++, projects an increase of the GCB rainfall magnitude associated with the Atlantic Niño (see Fig. B.12 (a) for the values of the sign-dependent average). These two different signals are considered because we do not know exactly the realistic AEM-rainfall pattern in the future. Figure 3.5 highlights the GC++ and GC+- groups. It shows that the future changes in the rainfall response to AEM in a given model do not depend on its performance on the representation of the AEM rainfall pattern under present-day conditions. Similarly, Fig. 3.3 shows a negligible impact of the models' skill in representing the SST patterns on the long-term future change in the Guinea Coast rainfall response to the AEM. Furthermore, the six models that simulate weak or negative JAS AEM-rainfall anomalies over the GCB during the 1985-2014 period are similarly divided into two groups, GC-- and GC-+, which contain models projecting a strengthened and a decreased (or inverted) rainfall intensity related to one JAS ATL3 SST index standard deviation, respectively.

Next, the 30 CMIP6 models are able to simulate a positive rainfall anomaly over the Atlantic Ocean in the area between the equator and 5° N. The same sign-dependent average analysis performed over the equatorial Atlantic region (the

EAB area, Fig. B.12 (b)) conduces to the OC++ and OC+- groups (Fig. B.16). The GC+- and OC+- groups show an overall decrease of the rainfall magnitude related to AEM in the future, which is associated with a weakened Bjerknes feedback. By contrast, in GC++ and OC++ groups, weak changes of the zonal wind and sea surface height anomalies indicate that the Bjerknes feedback acts similarly in the present and future climates. Moreover, the spatial characteristics of the changes are similar for the near-term, mid-term and long-term periods, with an enhancement of the pattern of change with time (as summarized in Fig. B.13). Therefore, we will focus the next discussion on the 2070-2099 changes relative to the 1985-2014 period, as these changes are the most amplified. The detailed lists of the models in the different categories for the three future periods are available in Tables B.2, B.3 and B.4.

Among the 24 models in the GC+ group, 12 (12) project a decrease (an increase) of the rainfall magnitude associated with AEM (Fig. 3.9 (a)-(b)). The long-term projection of the AEM-rainfall signal in the GC+- group is limited to the Atlantic Ocean, and is hardly robust over the Guinea Coast (Fig. B.18). However, relative to the period 1985-2014, models in the GC++ group show an eastward and northward shift of the anomalous positive rainfall belt over the Atlantic Ocean at the end of the 21st century. This results in an enhancement of the rainfall teleconnection pattern over the Guinea Coast, mainly over the Cameroon mountains and Guinean highlands.

Regarding the SST patterns related to the Atlantic Niño, the projected changes display a decrease in SST over the equatorial Atlantic and off the Angola-Benguela Coast in both GC+- and GC++ (Fig. 3.9 (d)-(e)). However, the SST patterns of change in GC+- and GC++ are different in the north and south of the 5° N - 5° S band. Compared to the 1985-2014 period, there is a poleward (in both directions) and westward extension of the anomalous warming associated with AEM in GC+-, which directly results in a decreased zonal and meridional surface pressure gradients over the Atlantic Ocean, and a reduced surface convergence over the equatorial Atlantic (Fig. 3.9 (m)). Regarding the 850 hPa zonal wind anomalies, their magnitude associated with AEM in the GC+- group is projected to decrease in the future (Fig. 3.9 (g)), while this change is very weak and unrobust in GC++ group (Fig. 3.9 (h)). As a consequence, in the GC+- group, the projected equatorial Atlantic zonal moisture flux and moisture flux convergence related to AEM are weaker, and their contribution to intensify the Guinea Coast rainfall is limited (Fig. 3.9 (m)).

The SSH response to AEM in GC+- is also projected to weaken in the future, compared to the present-day situation (Fig. 3.9 (j)). Together with the important change of the low-level zonal wind, these results evidence a weakened Bjerknes feedback over the equatorial Atlantic in a warmer climate. In the GC++ group, there is no substantial difference between the 1985-2014 and 2070-2099 SSH

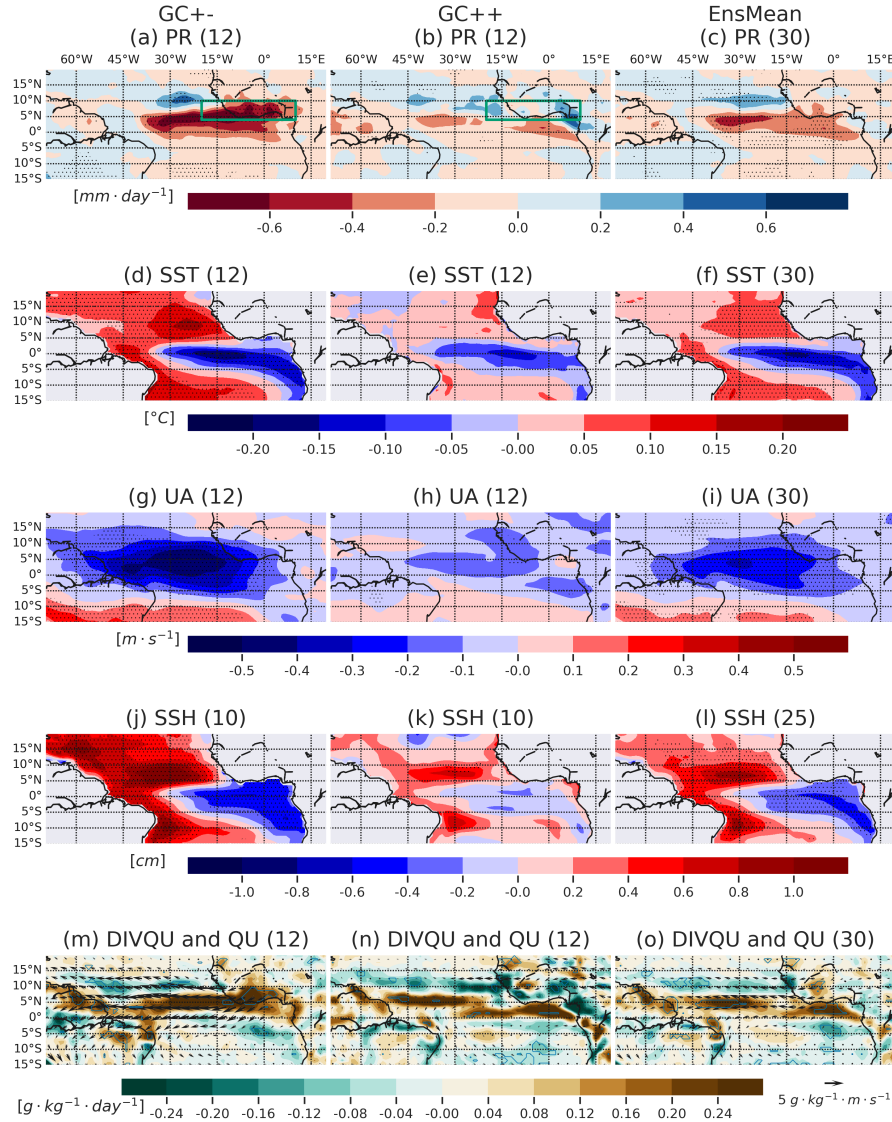


Figure 3.9: Long-term changes of the JAS rainfall (a-c), SST (d-f), 850 hPa zonal wind (g-i), sea surface height (j-l), 850 hPa moisture flux (vectors) and moisture flux divergence (in colors) (m-o) regression patterns associated with the standardized JAS ATL3 SST index, relative to the present-day climate (2070-2099 minus 1985-2014). Stippling in (a)-(l) and contours in (m)-(o) indicate areas where the mean change (in colors) is significant at 95 % level according to a two-sided Welch t-test and where at least two thirds of the models agree on the sign of the change. The number of models in each group is indicated in parentheses.

patterns related to the Atlantic Niño (Fig. 3.9 (k)). This means that the intensity of the Bjerknes feedback remains quite similar in the GC++ models over the two periods. Furthermore, compared to the present-day situation, in GC++, the surface moisture convergence related to the Atlantic Niño is displaced eastward and northward over the equatorial Atlantic and Guinea Coast in the future (Fig. 3.9 (n)). This is consistent with the projected AEM-rainfall increase over the Guinea Coast in GC++, in agreement with Mohino and Losada [2015]. Besides, over the equatorial Atlantic, there is an overall decrease of the convection associated with AEM in both GC+- and GC++ models (Figs. B.17 and B.18).

From the six models in the GC- group, four (the GC+- group) project an eastward shift of the positive AEM-rainfall anomalies over the tropical Atlantic, leading to an increased rainfall intensity over the Guinea Coast (Fig. B.19). In addition, the negative AEM-rainfall located north of the positive AEM-rainfall band weakens in these models. This future response is, however, unrobust over the Guinea Coast. In the other two models (the GC-- group), over the tropical Atlantic, the dipolar rainfall response to the AEM remains present in the future period. Moreover, the band of negative rainfall anomalies which extends from the ocean to the westernmost region of the Guinea Coast is shifted equatorward. Overall, the patterns of change of the OC+- group (Fig. B.16), and the 30 CMIP6 EnsMean (Figs. 3.9 (c), (f), (i), (l), (o)) are similar to those in the GC+- group. Table B.1 summarizes the main results for the GC++, GC+- categories and the EnsMean.

In a warmer climate, over the equatorial Atlantic, the models project an increase of the mean JAS sea level pressure which is greater in the western basin than in the eastern area. This zonal surface pressure gradient drives anomalous westerlies over the equatorial Atlantic (Fig. B.20). As a consequence, the weakened trade winds cause a deeper mean thermocline in the eastern equatorial Atlantic. This is indicated by the increased oceanic mixed layer depth, here used as a proxy for the top of the thermocline depth. As the thermocline becomes deeper, it responds less to the SST changes at the surface during the Atlantic Niño phases, and leads to a weakened Bjerknes feedback.

3.6 Discussion and Conclusion

In this study, the JAS Atlantic Niño impact on the JAS rainfall in Guinea Coast and over the tropical Atlantic has been evaluated. First, the 1985-2014 period of the historical simulations has been considered. The spatial patterns of the JAS SST anomalies related to the standardized JAS ATL3 SST index in the GCMs show good correlations with the pattern in ERA5. However, the maximum rainfall position associated with the Atlantic Niño in the 30 GCMs ensemble mean is displaced south of the observed position over the tropical Atlantic. All the models manifest an increased rainfall associated with positive AEM phases over the

equatorial Atlantic. In the case of the Guinea Coast region, 24 models exhibit a realistic increased rainfall related to a warm phase of the Atlantic Niño, against 6 models which show a negative or weak rainfall response. In connection with the AEM, there is a weaker zonal wind response and a weaker zonal moisture flux toward the Guinea Coast, as well as a weaker deep convection in these 6 models (the GC- group), relative to the 24 models (the GC+ group).

Secondly, the change of the Atlantic Niño and its impact on the rainfall over the equatorial Atlantic and the Guinea Coast in a climate with a high anthropogenic emission of greenhouse gases has been evaluated. Analyses of the Shared Socioeconomic Pathway 5-8.5 simulations demonstrate a gradual decrease of the SST anomalies associated with one standard deviation of the Atlantic Niño index for the near-term, mid-term and long-term periods, relative to the present-day climate. The ensemble mean change of the standard deviation of the ATL3 SST index relative to the present-day period is -6% , -13% and -22% for those three periods, respectively. Among the 30 GCMs, 18, 23 and 27 agree on the sign of the ATL3 SST index STD change, for the three future periods respectively. Much of the models also indicate a gradual weakening with time of the rainfall magnitude related to the Atlantic Niño over the Guinea Coast and the equatorial Atlantic in the future.

More specifically, focusing on the Guinea Coast, among the 24 models which present a relatively good AEM-rainfall signal in the 1985-2014 period, there are 12 GCMs (the GC+- group) which project a relative decreased rainfall related to AEM in the 2070-2099 period. Analyses of the JAS SST, 850 hPa zonal wind and SSH anomalies indicate that the magnitude of their variations associated with one standard deviation of the Atlantic Niño decreases, suggesting a weakening of the Bjerknes feedback in a warmer climate. By contrast, 12 other models show weak differences in the patterns associated with the Atlantic Niño, and a little increase in the rainfall pattern over elevated areas in Guinea Coast (the GC++ group). While the GC+- group exhibits an important decrease of the zonal wind variability associated with the Atlantic Niño in the future, this variability remains similar in the GC++ group. In these models, we argue that the Bjerknes feedback should be of the same intensity in the present-day and at the end of the 21st century. Alongside, there is a projected westward and a poleward extension of the anomalous warm water associated with the Atlantic Niño in the GC+- group. This contributes to the reduction of the equatorial Atlantic surface convergence associated with AEM.

In the GC+- group, the equatorial Atlantic convection associated with the Atlantic Niño is projected to decrease, as indicated by the weakened wind divergence difference between 200 hPa and 850 hPa. This decrease is weaker in the GC++ group. This result adds support to Jia et al. [2019], who found a more stabilized troposphere over the tropical Atlantic in the future, which in turn reduces the convection during the positive phases of the Atlantic Niño. Their analysis was focused

on the link between the Atlantic Niño and La-Niña, while our analysis is focused on the Atlantic Niño teleconnection with the tropical Atlantic and Guinea Coast.

In the last part of this study, we linked the background JAS mean state change to the change of the Atlantic Niño variability. In a warmer climate, we found that over the equatorial Atlantic, the mean sea level pressure change is positive and presents an East-West gradient. This contributes to decreasing the intensity of the mean easterlies, thus limiting the equatorial upwelling. Subsequently, the thermocline becomes deeper along the equator, and the deepening is more important in the eastern equatorial Atlantic. For this reason, the influence of the eastern equatorial Atlantic thermocline on the surface during the Atlantic Niño phases is weakened in the future. This explains the projected decreased variability of the Atlantic Niño in the SSP5-8.5 emission scenario.

Future studies are needed to assess the impact of the warm biases in the eastern equatorial Atlantic on the teleconnection patterns of the Atlantic Niño, and could help to increase the reliability of the projected changes. Moreover, the reduction of these biases is crucial for the rainfall projections over West Africa. Further analyses are also needed to understand the multidecadal modulation of the Guinean coast rainfall and extreme rainfall by oceanic internal modes of variability in climate models for both present-day and future climate conditions.

4

Future changes in the mean and variability of extreme rainfall indices over the Guinea Coast and role of the Atlantic equatorial mode

Abstract

The occurrence of climate extremes could have dramatic impacts on various sectors such as agriculture, water supply, and energy production. This study aims to understand part of the variability in the extreme rainfall indices over Guinea Coast that can be related to the Atlantic equatorial mode (AEM) whose positive phases are associated with an increase in the intensity and frequency of rainfall events. We use six extreme indices computed from six observed rainfall databases and historical and SSP5-8.5 simulations from 24 general circulation models (GCMs) that participate in the sixth phase of the Coupled Model Intercomparison Project (CMIP6) to study changes in extreme rainfall events over Guinea Coast during July-September. We found that current GCMs clearly overestimate the frequency of wet events and the maximum number of consecutive wet days. The magnitude of the other extreme indices simulated is within the range of the observations which, moreover, present a large spread. Our results confirm existing studies. However, less attention has been paid to the evaluation of the modelled rainfall extremes associated with the AEM under different climate conditions, while the variability of the AEM is expected to decrease in the future with a potentially significant impact on the extreme events. Here, we use six (one) observed rainfall (sea surface temperature) data and 24 GCMs outputs to investigate the present-day, near-term, mid-term, and long-term future links between the AEM and the extreme rainfall events over Guinea Coast. The biases in the extreme rainfall re-

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sponses to the AEM are subject to a large spread across the different models and observations. For the long-term future (2080-2099), less intense and more frequent rainfall events are projected. As an illustration, the multimodel ensemble median (EnsMedian) maximum rainfall during five consecutive wet days (RX5day) would be 21 % higher than under present-day conditions. Moreover, the variability of the majority of the extreme indices over the Guinea Coast is projected to increase (48 % for RX5day in the long-term future). By contrast, the decreased variability of the AEM in a warmer climate leads to a reduced magnitude of the rainfall extreme responses associated with AEM over the Guinea Coast. While under present-day conditions the AEM explains 18 % of the RX5day variance in the EnsMedian, this value is reduced to 8 % at the end of 21st century. As a consequence, in absolute, there is a projected increase in the total variability of most of the extreme rainfall indices, but the contribution of the AEM to this variability weakens in a warmer future climate.

4.1 Introduction

In Chapter 3, we demonstrated that the Atlantic equatorial mode (AEM) impact on the Guinea Coast rainfall (GCR) during the boreal summer will weaken in a future warmer climate, according to the SSP5-8.5 scenario. As the AEM phases also impact extreme rainfall events, it is likely that future changes in its variability would affect its influence on extreme occurrences.

Severe flash flood events were reported over West and Central Africa during the last decades, affecting millions of people, destroying houses, buildings and left dozens of deaths [United Nations Office for the Coordination of Humanitarian Affairs (OCHA), 2012, 2021]. Some of these events were caused by severely abnormal rainfall events, which often contribute to increasing the rivers levels, favouring flooding of the surrounding areas [Elagib et al., 2021; Fofana et al., 2022]. According to the climate scenarios, for which future global warming is limited to 1.5 °C, 2.4 °C and 3.5 °C by 2100, people born in 2021 in Sub-Saharan Africa will experience, during their lifetime, 4.6, 8, and 8.6 times more river flooding events than without climate change, respectively (<https://myclimatefuture.info/>, last access: 11-08-2022). This increase is two to fourfold higher than the flooding events experienced by people in the same area, born in 1960, and highlights the climate urgency in reducing our greenhouse gas emissions for future generations [Thiery et al., 2021].

During the last decades, the Guinea Coast has experienced changes in its rainfall characteristics, with conclusions that depend on the season considered, the period of study, the source of observations and the extreme rainfall index used [Bichet and Diedhiou, 2018; New et al., 2006; Odoulami and Akinsanola, 2017; Kpanou et al., 2020; Dike et al., 2020]. It is important to highlight that these

changes are not homogenous over the whole region. Over the period 1961-2000, the observations indicate a decrease in the annual rainfall in southern Nigeria, with, an increase in the intensity of the daily rainfall and the annual maximum 1 day rainfall [New et al., 2006]. Odoulami and Akinsanola [2017] showed, however, a significant negative trend in the June-September mean daily rainfall over the Guinea Coast during the 1998-2013 period. Nevertheless, they found a decrease in the frequency of rainfall events higher than the 95th percentile. This result is opposite to Kpanou et al. [2020], who, on an annual base, found an increasing trend in the number of days with rainfall exceeding the 95th percentile over coastal areas in the southwestern Côte d'Ivoire, Togo, and Bénin. Odoulami and Akinsanola [2017] also observed an overall upward and insignificant trend in the number of dry days over that area and a significant decreasing trend in the number of wet days. As an exercise, and to be consistent with our current study we show in Fig. C.1 the observed trends in some extreme indices for the July-September season (JAS) and the 1995-2014 period. The figure shows an overall intensification of rainfall during wet days over the southern part of Guinea Coast (between 7.5°W and 7.5°E), a decrease in the frequency of wet events, a decrease in the maximum consecutive wet days, and an increase in the amount of rainfall during wet days. The Sixth Assessment Report (AR6) of the Intergovernmental Panel on Climate Change (IPCC) stated that there is a low confidence in the observed heavy precipitation trend over West Africa for the last decades, as well as in the contribution of human influence to that trend [Seneviratne et al., 2021].

Several studies have shown that climate General Circulation Models (GCMs) participating in the sixth phase of the Coupled Models Intercomparison Project (CMIP6) simulate reasonably well the spatial distribution of the rainfall extreme indices in West Africa, with however different levels of magnitude [Faye and Akinsanola, 2021; Klutse et al., 2021]. However, Regional Climate Models (RCMs) forced with CMIP5 GCMs outputs show an added value in simulating the spatial distribution of the present-day mean rainfall, maximum number of consecutive dry days and the total rainfall exceeding 95th percentile over West Africa [Akinsanola et al., 2020]. In addition these models project an enhancement of extreme rainfall events over Guinea Coast in the future, under the Representative Concentration Pathway (RCP) 4.5 and 8.5 forcing scenarios [Akinsanola and Zhou, 2019]. Furthermore, Akinsanola et al. [2020] demonstrated from RCP4.5 and RCP8.5 projections with RCMs an increase in the West African rainfall variability, which is associated with enhanced mean rainfall and rainfall extremes in the Guinea Coast. This increase in variability was explained by the change in the water vapour in a warmer world following the Clausius Clapeyron equation [Akinsanola et al., 2020].

The first mode of covariability between the sea surface temperature in the tropical Atlantic and the rainfall over West Africa during the boreal summer indicates a strong connection between the eastern equatorial Atlantic SST and the Guinea

Coast rainfall, and explains 31% of the total covariability [Polo et al., 2008]. These local SST changes are related to the Atlantic equatorial mode, also named Atlantic zonal mode [Zebiak, 1993]. Positive phases of the AEM are characterized by above normal SST conditions in the eastern equatorial Atlantic. The enhanced low-level convergence over the warm oceanic area is accompanied by an ascent of moist air, which is then advected by the low-level circulation toward the Guinea Coast. This provides favourable conditions for the rainfall occurrence over this region [Polo et al., 2008; Rodríguez-Fonseca et al., 2015; Schubert et al., 2016; Lübbecke et al., 2018; Worou et al., 2020]. Worou et al. [2022] have demonstrated that the covariability between the Guinea Coast rainfall and the AEM is maximum in JAS, in the simulations performed within CMIP6 models (June–September in observations), under present-day conditions. Over the 1901–2016 period, in observations, the eastern equatorial Atlantic positive correlation with the rainfall over the Guinea Coast was clear during the boreal summer [Losada et al., 2012; Diatta and Fink, 2014; Worou et al., 2020]. Nevertheless, the variability of the AEM is projected to decrease under future global warming [Worou et al., 2022; Crespo et al., 2022; Yang et al., 2022]. This would potentially reduce the role of the AEM in the Guinea Coast rainfall variability. Moreover, Diatta et al. [2020] showed that the Guinea Coast rainfall extremes are strongly correlated with the AEM index. Atiah et al. [2020] also showed an increase (a decrease) in the total annual wet day rainfall, heavy rainfall and very heavy rainfall over Ghana during warmer (colder) sea surface conditions in the eastern equatorial Atlantic. Therefore, it is important to understand and quantify the projected potential changes in the extreme rainfall events in Guinea Coast that are associated with the AEM.

Heavy rainfalls in West Africa could occur under organized mesoscale convective systems such as squall lines, enhanced convection triggered by the propagation of the African Easterly Waves (AEWs) or moisture advection from the Atlantic Ocean [Thorncroft and Hoskins, 1994a,b; Gu and Adler, 2004; Maranan et al., 2019; Baidu et al., 2022]. A combination of these different factors could also lead to high impact weather events, such as the very heavy rainfall that occurred in Ouagadougou on the 1st of September 2009 [Engel et al., 2017; Lafore et al., 2017; Beucher et al., 2019]. More than one third of the annual rainfall was recorded within 24 hours, and reached 263 mm. In addition, the large-scale sea surface temperature (SST) conditions during this event showed an anomalous warming of the Mediterranean Sea, a negative North Atlantic Oscillation (NAO) -like pattern and a pronounced Atlantic cold tongue. This pronounced cold tongue corresponds to a negative phase of the AEM, which is favourable to an increased rainfall over the Sahel.

Previous studies suggested a link between the AEM and the Guinea Coast extreme rainfall variability in observations. However, this link in the current GCMs and the impact of the projected reduction in the AEM variability on the extreme rainfall

events over Guinea Coast have not yet been assessed. This study aims to further explore the relationship suggested in previous studies between the AEM and the extreme rainfall events over the Guinea Coast under present-day and future climate conditions by using outputs from CMIP6 models and to determine how those changes in AEM contribute to the simulated changes in rainfall extremes. The Section 4.2 is dedicated to the data used in this work and the methods adopted. Section 4.3 describes the present-day statistics of the rainfall extremes over the Guinea Coast, and the near-term, mid-term and long-term changes in the mean and variability of these extremes. In Section 4.4, the impact of the AEM on the Guinea Coast rainfall extremes under present-day conditions is evaluated, as well as the future changes in the extreme indices responses related to the AEM. We conclude with Section 4.5, where our findings are summarized.

4.2 Datasets and methods

The present study focuses on the impact of the Atlantic equatorial mode on the rainfall extreme events over the Guinea Coast under different climate conditions. Four 20-year periods are considered: the present-day (1995 - 2014), the near-term future (2021 - 2040), the mid-term future (2041 - 2060) and the long-term future (2080 - 2099) periods. The choice of these periods is based on the definition provided by IPCC [2021] (see their Table SPM.1).

4.2.1 CMIP6 data

Daily rainfall and monthly SST data from 24 climate models participating in CMIP6 were retrieved from one of the Earth System Grid Federation (ESGF) portals (e.g. <https://esgf-node.llnl.gov/search/cmip6/>, last access: 15 June 2022) and analysed. The analysis of the present-day period is based on the historical simulations, which covered the 1850-2014 period and were forced with observed natural and anthropogenic forcings [Eyring et al., 2016]. The analyses for the future periods rely on the shared socioeconomic pathway with a high greenhouse gas emission [SSP5-8.5, O'Neill et al., 2016]. These simulations are started from the year 2015. We choose the SSP5-8.5 scenario to get the clearest signal of climate change.

The majority of the studies analysing the internal modes of variability in the tropical Atlantic only used one member for each available model [e.g., Kucharski and Joshi, 2017; Richter and Tokinaga, 2020; Worou et al., 2022; Crespo et al., 2022; Yang et al., 2022]. We rely here on the same approach and use one realization for each model in our study, assuming that all realizations are equivalent for the analyses we performed. It could be interesting to test the limitations of this hypothesis using the models that provide a relatively large ensemble for both

historical and future simulations. However, this is out of the scope of this present study, and we expect that the differences between ensemble members of a specific model are smaller than the differences between models for the diagnostics analysed here. Table 4.1 provides a list of the different models used, their corresponding ensemble member, and their resolution.

Table 4.1: List of the analysed 24 CMIP6 models in this study, their ensemble member considered in the historical and SSP5-8.5 simulations, and their resolutions. In the variant label for each model member, "r", "i", "p" and "f" represent the realization index, the initialization method, the physics, and the forcing, respectively. Note that in the SSP5-8.5 outputs, the parent variant label does not necessarily correspond to the variant label. We then read thoroughly metadata in future simulation outputs and associate them to their corresponding parents, from which they were branched. For example, the daily precipitation file from the SSP5-8.5 simulation r2i1p1f1 performed with CESM2 was branched on the historical r11i1p1f1 output.

CMIP6 model	Historical member	SSP5-8.5 member	Resolution ($^{\circ}\text{lat} \times ^{\circ}\text{lon}$)
ACCESS-CM2	r1i1p1f1	r1i1p1f1	1.25 x 1.88
ACCESS-ESM1-5	r1i1p1f1	r1i1p1f1	1.25 x 1.88
CESM2	r11i1p1f1	r2i1p1f1	0.94 x 1.25
CESM2-WACCM	r1i1p1f1	r1i1p1f1	0.94 x 1.25
CNRM-CM6-1	r1i1p1f2	r1i1p1f2	1.40 x 1.41
CNRM-CM6-1-HR	r1i1p1f2	r1i1p1f2	0.50 x 0.50
CNRM-ESM2-1	r1i1p1f2	r1i1p1f2	1.40 x 1.41
CanESM5	r1i1p1f1	r1i1p1f1	2.79 x 2.81
EC-Earth3	r1i1p1f1	r1i1p1f1	0.70 x 0.70
EC-Earth3-Veg	r1i1p1f1	r1i1p1f1	0.70 x 0.70
GFDL-ESM4	r1i1p1f1	r1i1p1f1	1.00 x 1.25
HadGEM3-GC31-LL	r1i1p1f3	r1i1p1f3	1.25 x 1.88
INM-CM4-8	r1i1p1f1	r1i1p1f1	1.50 x 2.00
INM-CM5-0	r1i1p1f1	r1i1p1f1	1.50 x 2.00
IPSL-CM6A-LR	r1i1p1f1	r1i1p1f1	1.27 x 2.50
KACE-1-0-G	r1i1p1f1	r2i1p1f1	1.25 x 1.88
MIROC-ES2L	r1i1p1f2	r1i1p1f2	2.79 x 2.81
MIROC6	r1i1p1f1	r1i1p1f1	1.40 x 1.41
MPI-ESM1-2-HR	r1i1p1f1	r1i1p1f1	0.94 x 0.94
MPI-ESM1-2-LR	r1i1p1f1	r1i1p1f1	1.87 x 1.88
MRI-ESM2-0	r1i1p1f1	r1i1p1f1	1.12 x 1.13
NorESM2-LM	r1i1p1f1	r1i1p1f1	1.89 x 2.50
NorESM2-MM	r1i1p1f1	r1i1p1f1	0.94 x 1.25
UKESM1-0-LL	r1i1p1f2	r1i1p1f2	1.25 x 1.88

4.2.2 Observations

To evaluate the performance of the models in simulating different aspects of the extreme indices, and to take into account uncertainties in observations, we use daily rainfall data from six different sources: CHIRPS, ARCV2, PERSIANN-CCS-CDR, REGEN LongTermStns, TAMSAT v3, and GPCC-FDD-v2022. These data are described in Table 4.2. We show in Fig. C.2 three additional observed rainfall data: GSMAP, IMERG and GPCP [Kubota et al., 2007; Huffman et al., 2015; Becker et al., 2013]. These data are not considered in our current work, as they do not cover the present-day period (1995-2014). They show, however, a consistent annual cycle in the extreme rainfall indices with the other observational rainfall data (Fig. 4.1). A comparison between some of those datasets is also given in Sanogo et al. [2022]; Ageet et al. [2022].

Observed monthly SST data are derived from the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST). This dataset is available at a spatial resolution of 1° [Rayner et al., 2003] and covers the 1870-2022 period. We do not include additional observed SST data, as we consider that this is not a critical point for our analyses.

Table 4.2: List of the six observed rainfall data considered in this study.

Data	Title	Availability	Resolution ($^{\circ}\text{lat} \times ^{\circ}\text{lon}$)	Reference
CHIRPS	Climate Hazards Group Infrared Precipitation with Station Data	1981-2023	0.25 x 0.25	Funk et al. [2014]
ARCV2	African Rainfall Climatology Version 2	1983-2023	0.1 x 0.1	Novella and Thiaw [2013]
PERSIANN-CCS-CDR	Precipitation Estimation from Remotely Sensed Information using Artificial Neural Networks-Cloud Classification System-Climate Data Record	1983 - 2023	0.04 x 0.04	Sadeghi et al. [2021]
REGEN LongTermStns	Rainfall Estimates on a Gridded Network based on long-term station data v1-2019	1950-2016	1 x 1	Contractor et al. [2020]
TAMSAT v3	Tropical Applications of Meteorology using SATellite data and ground-based observations, version 3.0	1983-2023	0.0375 x 0.0375	Maidment et al. [2017]
GPCC-FDD-v2022	Global Precipitation Climatology Centre, Full Data Daily Version 2022	1982-2020	1 x 1	Markus et al. [2022]

4.2.3 Choice of the season

This study is focused on the July-September season (JAS), following Worou et al. [2022]. This season contributes to 46 % of the total annual rainfall over the Guinea Coast. It is also dominated by the monsoon system of West Africa, which is characterized by an abrupt shift of the rainfall band from the coastal areas in the end of June to the Sahel region. Sanogo et al. [2022] found a strong coupling between the West African monsoon rainfall and the occurrence of extreme precipitation events, mainly during the boreal summer. Note that although mesoscale convective systems have an important contribution to the GCR during April-May-June and September-October-November [Maranan et al., 2018], the annual cycle of the very wet days index (R95p index) across different observations indicates the high-

est values during July, August, September and October (Fig. C.2 (b)). The total wet day precipitation (PRCPTOT) and the contribution of the monthly rainfall to the annual rainfall (ANNPCT) also show their highest values over these months (Fig. C.2 (a),(c)). As we do not focus our analysis on specific contributions of mesoscale convective systems to the rainfall over Guinea Coast, we will keep the JAS season as the focus of our study.

4.2.4 Methods

4.2.4.1 Definition of the AEM index

Before computing the AEM index, all the monthly observations and models SST data are remapped on the HadISST grid, at 1° of resolution. We used a bilinear interpolation method with the climate data operator routine (CDO, <https://code.mpimet.mpg.de/projects/cdo>). The index of the AEM is defined over the Atlantic Niño 3 region [ATL3, Zebiak, 1993] which extends between $20^\circ W - 0^\circ E$ and $3^\circ S - 3^\circ N$. There are models with a low resolution that only have a few grid points within the ATL3 region. If the AEM index was computed on each model grid, this would imply to use different regions for different models. This motivates the choice of a common grid of 1° of resolution, which requires interpolating the data. A similar procedure has been applied in Kucharski and Joshi [2017]; Worou et al. [2022] for instance.

Over every 20-year study period, the monthly SST data is linearly detrended at each grid point to remove any drift due to climate change. The resulting monthly SST anomalies are then averaged over the ATL3 region. Next, we compute the seasonal (JAS) mean of the obtained area-averaged monthly SST anomalies. This gives one value per year for the AEM index. The AEM index time series are then standardized, by dividing them by their standard deviation over the entire period considered.

4.2.4.2 Computation of the extreme rainfall indices

First, all the daily observational and model rainfall data are remapped on the grid of the model which has the lowest resolution. In our study, this corresponds to the grid of MIROC-ES2L model, with a resolution of 2.8° . We use a first order conservative remapping method with the CDO routine to perform the remapping of the rainfall data. We define the Guinea Coast region as the area extending between $18^\circ W - 15^\circ E$, and $4^\circ S - 10^\circ N$. There are 32 grid points inside this area on the MIROC-ES2L grid (Fig. C.3).

In this work, we analyse a set of six extreme rainfall indices defined by the Expert Team on Climate Change Detection and Indices (ETCCDI) (http://etccdi.pacificclimate.org/indices_def.shtml, last access: 16 June 2022):

- the SDII (simple daily intensity index), which describes the intensity of wet rainfall events;
- the R95p (very wet day), which describes the intensity of rainfall events exceeding the 95th percentile;
- the CWD (consecutive wet days), which describes the maximum duration of a wet event;
- the RX5day (maximum 5 days precipitation), which is the intensity of an event over the duration of five days;
- the FRQW (frequency of wet days), which describes the frequency of wet events;
- R20mm (very heavy precipitation days), which is a measure of the frequency of rainfall events exceeding 20 mm, and which could have a high socio-economic impact.

The details of these indices are provided in Table 4.3. The indices recommended by the ETCCDI are widely used to monitor and detect changes in drought and wet conditions over different regions. Some applications can be found in several studies such as New et al. [2006]; Sillmann et al. [2013a,b]; Mouhamed et al. [2013]; Diedhiou et al. [2018]; Faye and Akinsanola [2021]; Delhaye et al. [2022], among others. We computed these indices in two ways. Firstly, we computed monthly indices over the whole period of study, and perform the monthly averages to get the mean annual cycle. In a second way, we considered for each year in the period of study, daily rainfall values over the JAS season (92 values at each grid point). We obtain one value per year for each extreme index. The corresponding series provide the information needed to study the seasonal climatology and interannual variability of the extreme rainfall indices.

4.2.4.3 Links between the AEM and the extreme rainfall indices

The analysis of the JAS rainfall extreme patterns associated with the JAS AEM is completed through linear regressions of the extreme anomalies at each grid point onto the standardized JAS AEM SST index. The patterns are computed for each GCM. The regression patterns in the observations are computed by regressing the rainfall indices from the six different observational data onto the standardized AEM index from HADISST. In both cases, we will consider the multi-model ensemble median and multi-observation ensemble median to resume the common characteristics. To identify the most robust changes among the members of those observations and model ensemble median, we will only consider a two-third sign-agreement approach among the set of data to show the robust signals related to

Table 4.3: List and definition of the six rainfall extreme indices selected for this study. These indices are based on the definition provided by the ETCCDI.

Index Label	Index Name	Index Definition	Index Unit
SDII	Simple daily intensity index	For each year, compute the average of wet days daily rainfall in a month/season. A wet day is defined as a day when the rainfall is greater or equal to 1 mm	$mm \cdot day^{-1}$
R20mm	Very heavy precipitation days	For each year, count the number of days in the month/season when the daily rainfall is greater or equal to 20 mm	days
R95p	Very wet days	Let PR95 be the 95 percentile of the wet-day daily rainfall timeseries in the month/season over the 1995-2014 period. For each year, sum of rainfall over days in the month/season, when the rainfall amount is greater or equal to PR95	mm
RX5day	Maximum 5 days precipitation	For each year, compute the maximum of the rainfall sum over 5 consecutive days in the month/season.	mm
CWD	Consecutive wet days	For each year, compute the largest number of consecutive wet days in the month/season. A wet day is defined as a day when the rainfall is greater or equal to 1 mm	days
FRQW	Frequency of wet days	For each year, compute the number of wet days (when the rainfall is greater or equal to 1 mm) in the month/season. The result is divided by the total number of days in the month/season, and multiplied by 100	%

AEM: at a grid point, a regression coefficient value is robust if more than 66% of the models agree on the sign of the multimodel ensemble median (EnsMedian) [Rehfeld et al., 2020]. Similarly, the robustness of the models' EnsMedian change in the regression patterns between two periods will be based on the two-third sign agreement approach: at a grid point: the change in the regression slope between two periods is robust if at least two-third of the models agree on the sign of the EnsMedian.

4.2.4.4 Robustness of the mean state changes

We use the methodology suggested in the Cross-Chapter Box Atlas 1 of the IPCC AR6 [Gutiérrez et al., 2021; Dosio et al., 2021] to disentangle the forced changes signal from the internal variability. We will mainly indicate regions where the forced response (the EnsMedian change in the climatology) is robust, or conflicting or no robust. A change in the mean state is considered as robust if two thirds of the models show changes greater than the interannual variability (IAV) and 80% of the models agree on the sign of the change. The IAV is computed for each model following the equation:

$$IAV = \sqrt{\frac{2}{20}} \times 1.645 \times \sigma \quad (4.1)$$

where σ is the standard deviation of the linearly detrended variable for the present-day period (1995-2014). The factor $\sqrt{2/20}$ takes into account the variability of the difference between two 20-year periods. The factor 1.645 considers a confidence interval of 90% for the change signal to exceed the IAV. Unrobust change signal is also defined for grid points where more than 80% of the models show a small

change compared to their IAV. A change signal is considered non-robust if less than two-third of the models present a change greater than the IAV. Finally, a change signal is conflicting if more than two-third of the models project a change greater than the IAV, and less than 80% agree on the sign of the change.

Additionally, we test another approach used in Monerie et al. [2017] (but we keep the IPCC approach), to determine the similarities in the robustness of the change signal. In this approach, a signal-to-noise ratio (SNR) is defined as:

$$SNR = \frac{\overline{\Delta X}}{\sigma \Delta X} \quad (4.2)$$

where:

- ΔX is the change in a variable X in one model,
- $\overline{\Delta X}$ is the multimodel ensemble mean (EnsMean) of the change in a variable X , which is termed the forced signal.
- $\sigma \Delta X$ is the spread of the change signal (the standard deviation) among the different models, which is due to the internal variability and the different responses of the GCMs.

Therefore, regions where the SNR is greater than one are supposed to experience a robust change signal. Qualitatively, both approaches give a similar result for the robust long-term changes in the extreme indices, except for R95 (Fig. C.5). In the IPCC approach, the projected long-term changes in R95p are non-robust, whereas the second approach shows robust changes over a large area of West Africa. Moreover, the IPCC approach gives more information on grid points where the change is not robust, whereas the second approach only provides information on the robust forced changes. Hereafter, we consider the IPCC approach.

4.2.4.5 Performance metrics for the of the models

The evaluation of the models' performance in representing the spatial distribution of the different extreme indices relative to observations is based on four metrics which have been applied in different studies [e.g., Akinsanola and Zhou, 2019; Faye and Akinsanola, 2021; Akinsanola et al., 2021; Li et al., 2021] :

- the percentage of bias (%BIAS)

$$\%BIAS = 100 \times \frac{\sum_{i=1}^n (M_i - O_i)}{\sum_{i=1}^n O_i} \quad (4.3)$$

- the normalized root mean square error (NRMSE)

$$NRMSE = \frac{\sqrt{\frac{1}{n} \sum_{i=1}^n (M_i - O_i)^2}}{\frac{1}{n} \sum_{i=1}^n O_i} \quad (4.4)$$

- the pattern correlation coefficient (PCC)

$$PCC(M, O) = \frac{Cov(M, O)}{\sqrt{Var(M)Var(O)}} \quad (4.5)$$

- a variant of the Taylor skill score [Taylor, 2001]

$$TSS = \frac{4(1 + PCC)^2}{\left(\frac{\sigma_{cmip6}}{\sigma_{observation}} + \frac{\sigma_{observation}}{\sigma_{cmip6}} \right)^2 (1 + PCC_0)^2} \quad (4.6)$$

where M and O are model and observation values, respectively, i is the index of a grid point, n is the number of grid points over which the data are compared, Cov and Var are the covariance and variance, respectively, PCC_0 is the maximum correlation reachable, set to 1 in our case, and σ_{cmip6} and $\sigma_{observation}$ are the standard deviations of the model and observation patterns (mean state pattern or teleconnection pattern) over the Guinea Coast, respectively. TSS and PCC (%BIAS and NRMSE) values are close to one (zero) for a very good representation of the observations by the model. The TSS is close to zero for no match between the model and observation. Moreover, one can choose to penalize models with low spatial correlation or low spatial variability by modifying the power of the different terms in the TSS [e.g., equations 4 and 5 in Taylor, 2001]. In our case, both the variability and correlation terms are to the same power. We do not choose to reward models that better represent the spatial pattern correlation than the spatial variability and inversely.

4.3 Present and future rainfall extremes over the Guinea Coast

4.3.1 Annual cycle of extreme rainfall indices over Guinea Coast

The mean annual cycle of the Guinea Coast extreme rainfall indices is shown in Figure 4.1. In the GCMs, the magnitude of those indices increase in general from January, reach a maximum during the boreal summer and decrease afterward. In the observations, a similar behaviour is present in the indices SDII, R20mm, R95p and RX5day (Fig. 4.1 (a)-(d)). It is not the case for the observed CWD and FRQW.

The CWD increases during the first months of the year, reaches a plateau in May (between 10 and 11 days for the EnsMedian) until July. Then it increases again and reaches a maximum in September (14 days for the EnsMedian). Interestingly, in the observations, the FRQW presents a bimodal structure in the observations, with its highest values in May and September (76 % and 79 % for the EnsMedian, respectively).

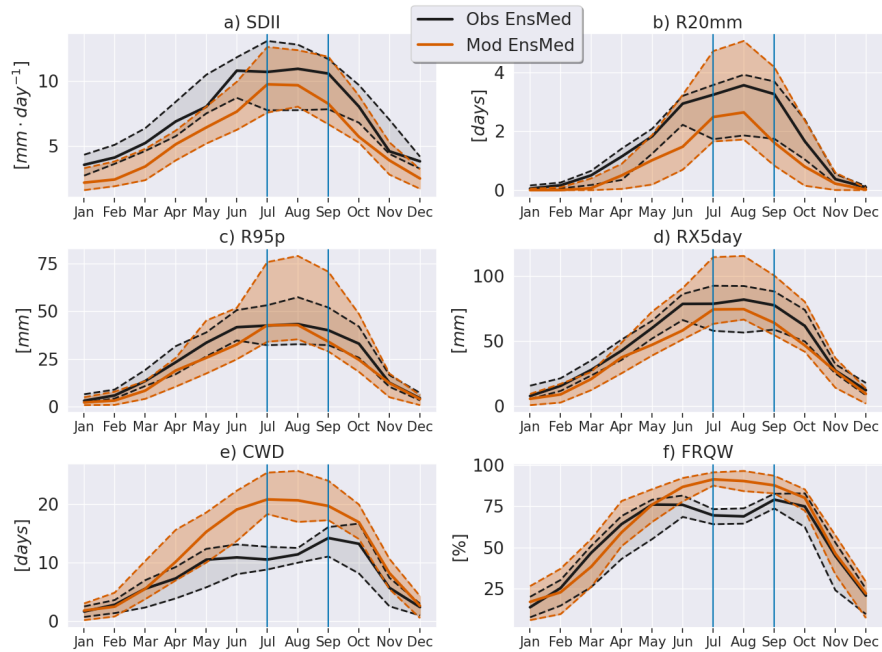


Figure 4.1: Annual cycle of the six extreme rainfall indices over the Guinea Coast during the 1995 – 2014 period. The black (orange) curves indicate the median values of the 6 (24) observations (models). The gray (orange) shading indicates the 10th - 90th percentile range for 6 (24) observations (GCMs). These percentiles are marked with the dashed curves (in black for observations, and in brown for models). The two vertical lines indicate the JAS season.

In the observations, the uncertainties in the SDII, R20mm, R95p and RX5day are larger for the boreal spring, summer and fall, compared to their values during the boreal winter. The GCMs spread for SDII is comparable to the observations' spread in JAS, despite the lower models' EnsMedian value compared to observations' EnsMedian. For example, in July, the SDII 10th-90th percentile values ranges between 7.5 and 13 mm · day⁻¹. For R20mm, R95p and RX5day, the models' spreads are larger than the spreads in the observations.

In JAS, there is a difference of one day between the R20mm observations and models EnsMedian. The 10th - 90th percentile values range between 1.7 and 5 days for the models. Interestingly, in July and August, both models and observations EnsMedians have the same value, 42 mm. The maximum difference among them in the annual cycle is of the order 8.5 mm. In August, the 90th (10th) percentile value of R95 is 79(35) mm for the models, against 57(33) mm in observations.

For the RX5day, the maximum difference between the models EnsMedian and observations EnsMedian is 20 mm (found in June). The maximum EnsMedian value in observations (models) is reached in August, 82(74) mm.

Finally, the models simulate too many consecutive wet events compared to observations, mainly during the boreal summer. There is no overlapping of the spread of the models and observations in JAS, for CWD and FRQW. The models EnsMedian CWD reaches 20 days in July, which is two-times the value of the observations EnsMedian (Fig. 4.1 (e)). Remarkably, the spread of the FRQW is reduced in JAS, compared to the other seasons, in both models and observations. The models EnsMedian FRQW value in July is 91%, against 69% for the observations EnsMedian (Fig. 4.1 (f)).

In summary, there is a large variability among the different observed datasets, in the annual cycle for most of the extreme indices. The uncertainties in the observed SDII are comparable to the uncertainties in the models, although the models EnsMedian indicate weaker intensities compared to the observations' EnsMedian. For the other indices, the spread of the models is higher than the spread of the observations.

4.3.2 Mean JAS rainfall extremes in Guinea Coast under present-day conditions

First, we present the observed and simulated JAS spatial distribution of rainfall extremes over West Africa in the present-day period (Fig. 4.2). The EnsMedian JAS spatial distribution of the indices in the six observational data is shown in contours in Fig. 4.2 (and in colours in Fig. C.6). Most of the indices show maximum values over the Guinean Highlands (in the western Guinea Coast) and the Cameroon mountains (east of 5°E), and moderate values in the center Guinea Coast (between 7.5°W and 5°E). From 10°N to the North, there is a gradual decrease in the wet indices. Focusing on Guinea Coast, the simple daily intensity index (SDII) ranges between 6(6) and 10(8) mm · day⁻¹ over the center of the region in the models (observations) EnsMedian (Fig. 4.2 (a)). Figure 4.3 (a) indicates a robust EnsMedian wet bias of 1 to 2 mm · day⁻¹ in the SDII over the center of Guinea Coast, and non-robust dry biases elsewhere. Over the Sahel, there is a homogeneous dry bias stretching from the West to the East, with robust values exceeding 2 mm · day⁻¹ in some locations.

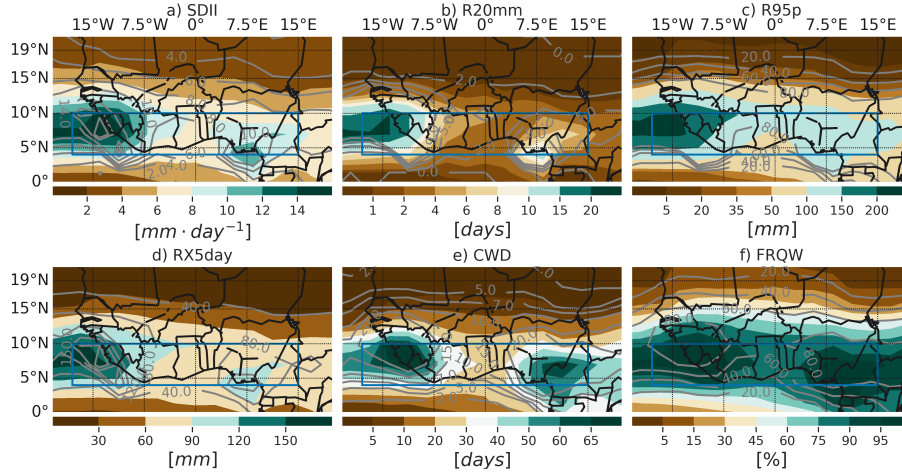


Figure 4.2: *EnsMedian* distribution of the mean JAS extreme rainfall indices over the 1995-2014 period from 24 CMIP6 models. The blue rectangles indicate the Guinea Coast region. Contours in gray show the observations *EnsMedian* values.

The R20mm index presents a non-robust *EnsMedian* overestimation of 1 to 2 days over the center Guinea Coast (Fig. 4.3 (b)). The R20mm bias is, however, negative and robust over the eastern Cameroun and the western Sahel. For the R95p and RX5day variables, the *EnsMedian* biases are positive and robust among the different observations, mainly over the center Guinea Coast. The bias values can reach and 50 mm and 30 mm for R95p and RX5day, respectively (Fig. 4.3 (c)-(d)). For these two variables, Sahel models *EnsMedian* exhibits weaker values compared to the observations, and this bias is robust north of 15°N in general.

The CWD and FRQW summer values in the northern hemisphere are uniformly overestimated by the GCMs *EnsMedian* over the Guinea Coast, and this bias is coherent among the different observations (Fig. 4.3 (e)-(f)). The spatial distribution of CWD biases indicates values between 10 and 30 days in the center Guinea Coast, and these values exceed 30 days east of 7.5°W. The CWD biases are weaker over Sahel (between 0 and -5 days). The FRQW positive biases range between 10 and 40 % over Guinea Coast, and exceed 40 % along the coast (between 7.5°W and 0°E). In the Sahel, FRQW is uniformly underestimated.

Figure 4.4 presents a summary of the models' performances in the simulating the extreme characteristics over Guinea Coast. It shows that the relative bias (in %) is higher in the CWD than in the other variables. Compared to the observations, all the CWD values in the models are higher in magnitude, with a %BIAS median value of 108 % (Fig. 4.4 (a)). This means, in other words, that the CWD values in

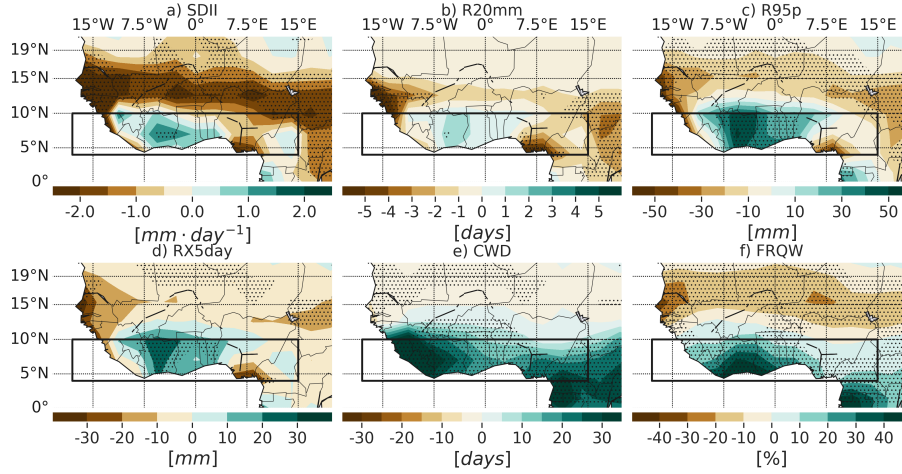


Figure 4.3: JAS mean biases of the 6 rainfall-based extreme indices over the 1995-2014 period. The biases for each model are computed relative to each of the 6 observations. In total, there are 6 observations $\times$ 24 models = 144 different biases. The stippling indicates regions where two-third of the biases agree on the sign of the EnsMedian of the 144 biases. The black rectangles indicate the Guinea Coast region.

the GCMs are two orders of magnitude greater than in the observations. This can also be seen in Fig. 4.4 (b), where the EnsMedian value of the NRMSE is 1.3, which is the highest value compared to the other extreme rainfall indices. This overestimation of the CWD by the GCMs over the coastal regions has already been reported by Faye and Akinsanola [2021]. Despite the substantial differences in magnitude compared to the observations, the CWD spatial correlation between models and observations gives a median value of 0.5. The corresponding TSS amounts to 0.4.

Compared to the other extreme indices, the models present a consistent positive bias in the JAS FRQW (Fig. 4.4 (a)). The EnsMedian %BIAS amounts to 23 %, which also confirms results in Fig. 4.1 (f). Likewise, the spread of the NRMSE is small, and its median value is 0.3 (Fig. 4.4 (b)). However, there is a large spread in the spatial correlations between the mean JAS values in the models and the values in observations (Fig. 4.4 (c)). The PCC and TSS values of the FRQW present similar characteristics with median values of 0.4 and 0.3, respectively.

R95p and RX5day are indices of intensity and duration of extreme rainfall events. The performance statistics of these indices indicate no clear representation compared to the different observations in term of %BIAS (Fig. 4.4 (a)). The median values of this latter are 2.7 % and -2.4 % for R95p and RX5day, respec-

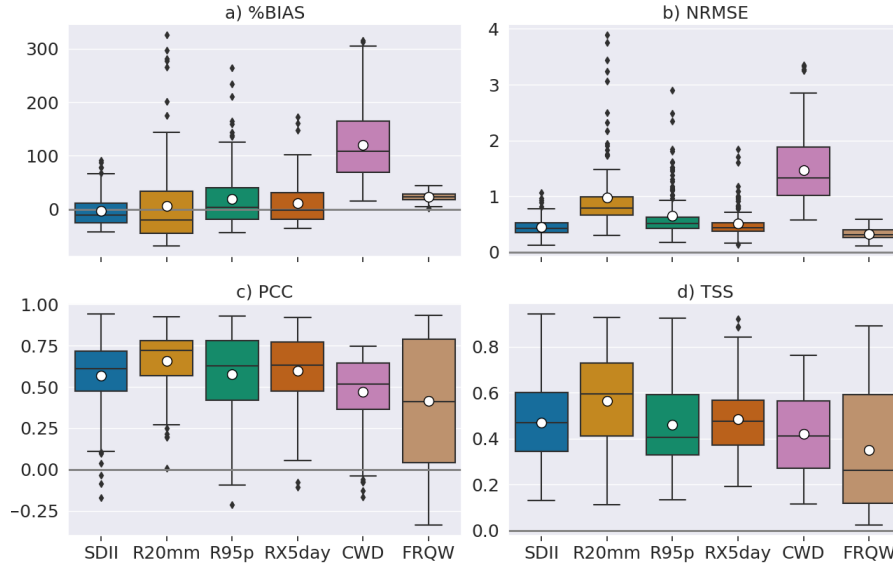


Figure 4.4: Performance of the 24 GCMs relative to the 6 rainfall observations, in representing the spatial distribution of the rainfall extreme indices over the Guinea Coast. Each boxplot indicates the distribution of $24 \times 6 = 144$ different values. The median (mean) values of the statistics are represented by the horizontal bar (white circle). Outliers are plotted individually with the black markers.

tively. Although the sign of the bias is not coherent among the different observations, more than 75 % of the NRMSE are below 0.63 (0.52) for R95p (RX5day) (Fig. 4.4 (b)). The EnsMedian values of PCC and TSS are 0.6(0.6) and 0.4(0.5), respectively for R95p (RX5day) (Fig. 4.4 (c)-(d)).

Finally, more than half of the model's biases in SDII and R20mm (over Guinea Coast) are negative, relative to the observations (Fig. 4.4 (a)). The EnsMedian value of these two variables amounts to -11% and -20% , respectively. As in the case of R95p and RX5day, there are more than 25 % cases where the SDII and R20mm are overestimated. Moreover, 75 % of the models present an NRMSE lower than 0.5(1), for the SDII (R20mm) index. The EnsMedian value of the PCC and TSS are 0.6 and 0.5 for SDII, respectively. For R20mm, the EnsMedian value of PCC and TSS corresponds to 0.7 and 0.6, respectively, and they are the highest values compared to the other extreme indices. It is important to note that the area defining the Guinea Coast is large, and the sign of the biases is not uniform over the region for all the variables. Thus, the area average over the region is subject to compensating effects.

In summary, over Guinea Coast, it rains too frequently and during too many consecutive days in the GCMs, compared to the observations. However, the intensity of daily wet events is weaker, and the number of very heavy precipitation days (with an intensity above 20 mm) is lower than observed. In contrast, very wet day intensity (with rainfall exceeding 95th percentile) is higher than observed. The R20mm, according to the TSS metric has the highest EnsMedian values, and can be considered as the index which is better represented in the GCMs. Moreover, the biases in the indices can be as large as the spread of the observations, mainly for R95p, CWD, and FRQW (Fig. C.4). The performance metrics for the individual models across the six different observations are shown in Figs. C.7 - C.10. Next, we will describe the changes of the rainfall extremes over the Guinea Coast under global warming, as projected by the climate models. All the indices will be studied, regardless of their good or poor representation by the GCMs, but the biases identified here will be accounted for in our discussion.

4.3.3 Changes in the mean and variability of the JAS rainfall extremes over the Guinea Coast

Figure 4.5 displays the projected EnsMedian long-term changes (2080-2099 minus 1995-2014) in the rainfall extreme indices over West Africa. The spatial patterns of the changes (relative to the present-day period) in the extreme rainfall indices over West Africa are similar for the near-term, mid-term and long-term periods, with, in general, a gradual increase in magnitude over the three future periods. However, the near-term and mid-term changes (Figs. C.11-C.12) are often less robust than the long-term changes.

In particular, over the majority of West Africa, the forced change in the different extreme indices does not emerge from the IAV in the near-term future period, as also shown by Monerie et al. [2017]. These authors found that the interannual forced changes in the central (western) Sahel in August-October (June-August) becomes stronger than the IAV only from the 2060s (2080s). In agreement with Monerie et al. [2017], we found an increase in the SDII over the central to eastern regions of Guinea Coast and Sahel, which becomes stronger than the IAV from 2040s (Fig. C.12 (a) and 4.5 (a)). Our results also show over the western areas of the Sahel, a long-term decrease in the SDII, which is stronger than the IAV, but with less agreement on the sign of the change among the models.

Analysing the various indices specifically, we found that the changes in the R95p are not robust over Guinea Coast, even at the end of the 21st century. This suggests a strong influence of IAV on the intensity of rainfall events exceeding the 95th percentile. We note that the long-term projection of these events is robust over the easternmost Sahelian region (Fig. 4.5 (c)). The R20mm is projected to increase over several areas of West Africa in the long-term future: easternmost

Sahel and Guinea Coast, central Guinea Coast (Fig. 4.5 (b)), which is consistent with Akinsanola and Zhou [2019].

The FRQW index exhibits a robust decrease of 5 % over Guinea Coast and over western Sahel (Fig. 4.5 (f)). Additionally, a robust decrease in the CWD is projected in the westernmost and easternmost region of the Guinea Coast. Over the Sahel, only the decrease in the western areas is robust for the CWD index (Fig. 4.5 (e)). These changes in CWD are consistent with the findings of Klutse et al. [2018] under global warming of 1.5 °C and 2 °C. Consistently, Wainwright et al. [2021] found an increase (a decrease) in the mean length of dry (wet) spells over the Guinea Coast during the wet season in a future warmer climate.

There is also a robust increase in RX5day over the Guinea Coast and the central to eastern Sahel (Fig. 4.5 (d)). These results show a tendency to less frequent and more intense rainfall over the Guinea Coast in a warmer climate, which happens over shorter duration. Our results are in accordance with Dosio et al. [2021], who found an increase in the mean SDII, and a decrease in the number of wet days in the JJA long-term projections for the Guinea Coast.

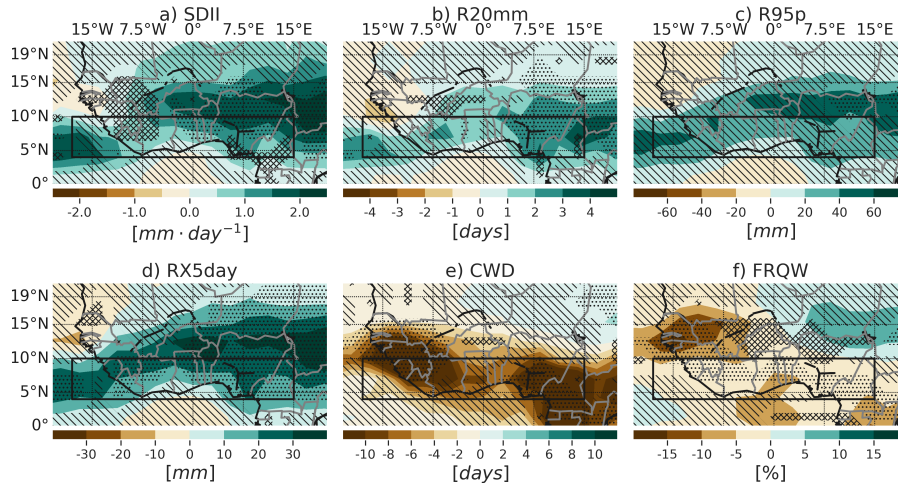


Figure 4.5: Projected multi-model ensemble long-term median change in the JAS rainfall extreme indices over West Africa, relative to the present-day period (2080-2099 minus 1995-2014). The stippling indicates regions where the change robustly emerges from internal variability (at least 66% of the models show a change greater than the IAV and at least 80% of the models agree on the sign of change); hatching (\\) indicates regions where the change is non-robust (fewer than 66% of the models show change greater than the IAV); crossed lines (X) indicate conflicting signals where at least 66% of the models show change greater than the IAV, with less than 80% agreement on the sign of the change.

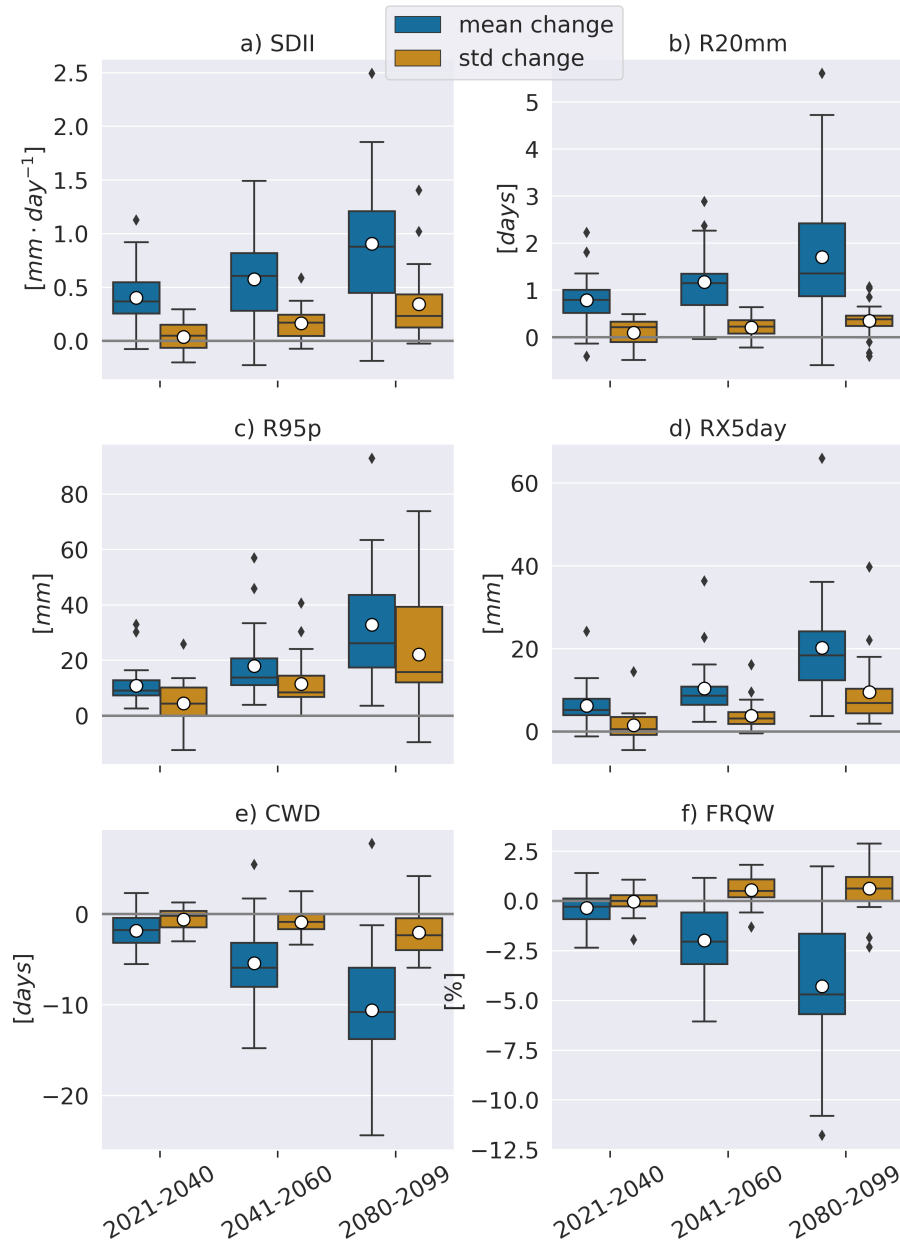


Figure 4.6: The boxplot of the average of the near-term, mid-term and long-term changes (relative to the present-day period) in the mean and standard deviation of the rainfall extreme indices over the Guinea Coast. The averages were computed from 24 different GCMs.

Figure 4.6 displays the average of the change in mean and standard deviation of the JAS extreme rainfall indices over the Guinea Coast for the 2021-2040, 2041-2060 and 2080-2099 periods, relative to 1995-2014. There is a tendency toward an increase in magnitude of SDII, R20mm, R95p and RX5day averages over Guinea Coast, which is gradual from the near-term to the long-term future periods (Figs. 4.6 (a)-(d)). The variability in these variables also exhibits a gradual increase with time in the future periods. Specifically, the changes in the mean and standard deviation of the R95p are of the same order of magnitude.

Consistent with the previous discussion, there is projected decrease in the CWD and FRQW indices over Guinea Coast (Figs. 4.6 (e)-(f)). The variability of the CWD will also decrease over the three future periods, with higher magnitudes in the long-term period compared to the others. For the FRQW, however, no change in the variability is projected in the near-term period, while there is a slight increase from the mid-term to the long-term future periods.

In summary, under future global warming, rainfall events over the Guinea Coast are projected to intensify (increase in the daily rainfall intensity over wet days), to be less frequent, to happen over shorter periods, and to be more extreme (increase in the amount of rainfall during very wet and extremely wet days). The overall signs of the long-term changes in SDII, R20mm, R95p, RX5day, and CWD over Guinea Coast the extreme indices are also in accordance with [Akinsanola and Zhou, 2019]. However, the results of these authors are obtained from RCM-CMIP5 projections and show more robust changes over West Africa, without taking into account the IAV. Moreover, there is a projected increase in the variability of most of the rainfall-based extreme indices, which is consistent with Akinsanola et al. [2020]. The multi-model median long-term percentage of changes in mean (variability) averaged Guinea Coast, for SDII, R20mm, R95p, RX5day, CWD and FRQW correspond to 11 (29), 35 (22), 20 (29), 21 (48), -23 (-12) and -5 (37)%, respectively (Fig. C.13-C.14).

Wet and dry extreme events over Guinea Coast could also happen under positive or negative phases of the Atlantic equatorial mode, respectively. As forced changes induced by future global warming can be exacerbated or damped by internal climate variability, study of the impacts of known internal climate mode of variability becomes crucial. It is the objective of this study, to understand the future changes in the rainfall extremes indices over the Guinea Coast that are connected with the AEM. The impact of the AEM on the extreme rainfall variability during the boreal summer is assessed here by regressing the JAS extreme indices onto the standardized JAS AEM index.

4.4 Impact of the AEM on the rainfall extreme events over the Guinea Coast

4.4.1 AEM impact on the rainfall extreme events under present-day climate conditions

Over the Guinea Coast, positive anomalies of the extreme rainfall indices are associated with positive phases of the AEM (Figs. 4.7, C.15). In GCMs, the extreme responses to this oceanic mode of variability are robust, according to the two-third sign agreement metric, except for the CWD index. In all cases, there is an anomalous increase (decrease) in the wet indices, in association with warm (cold) phases of the AEM over the Atlantic oceanic region (between the equator and the southern limit of Guinea Coast). A noteworthy feature of the anomalous patterns is the lack of robust signals over the majority of the Sahelian region. Particularly, the westernmost area of Sahel exhibits an opposite phase relationship with the AEM, for CWD and FRQW indices (Fig. 4.7 (e)-(f)), compared to Guinea Coast.

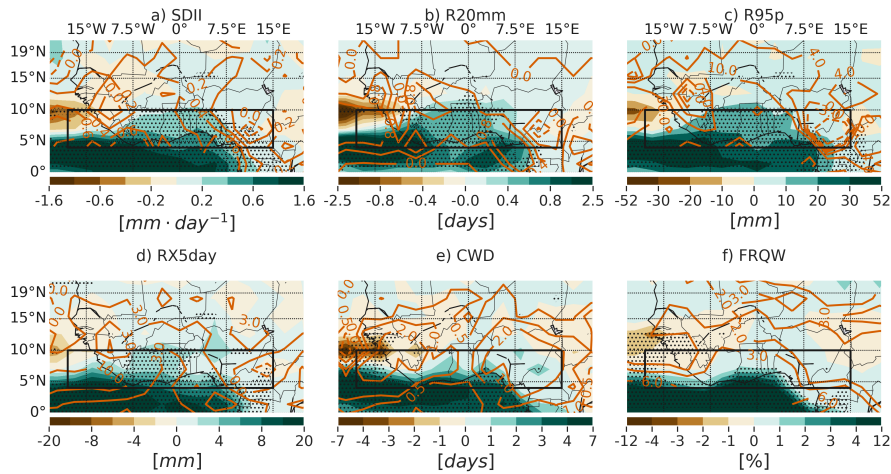


Figure 4.7: Maps of the JAS rainfall extreme indices regressed onto the standardized JAS AEM SST index over the 1995-2014 period. The stippling represents grid points where two thirds of the models agree on the sign of the EnsMedian regression coefficient across 24 GCMs. Contours indicate the multi-observation median values of the regression coefficients from six different rainfall data. The black rectangles indicate the Guinea Coast region.

The GCMs EnsMedian regression patterns associated with one standard deviation of the AEM index show an increased intensity of rainfall over wet days (SDII) (0.2 to $0.6 \text{ mm} \cdot \text{day}^{-1}$) over the Guinea Coast (Fig. 4.7(a)). There is about one-

day increase in the number of days with very heavy precipitation (R20mm), during positive AEM events (Fig. 4.7(b)). Warmer than average sea surface conditions in the eastern equatorial Atlantic favour an increased amount of rainfall during wet days (R95p) (Fig. 4.7(c)). Over the Guinea Coast, the anomalies related to one standard deviation of the AEM index range between 10 and 30 mm for R95p. The maximum precipitation over five consecutive days (RX5day) is also intensified over Guinea Coast (Fig. 4.7(d)), with positive anomalies ranging between 2 to 6 mm. The anomalous EnsMedian patterns of the maximum consecutive wet days (CWD) related to a warm phase of the AEM shows a non-robust increase of one to three days in wet spells duration (Fig. 4.7(e)). Similarly, the frequency of wet days over the Guinea Coast is higher up to 4% under warm sea surface conditions in the eastern equatorial Atlantic (Fig. 4.7(f)).

Fig. 4.8 displays the biases in the regression patterns associated with the AEM, for the different extreme indices. Over Guinea Coast, SDII and R20mm show no systematic biases relative to the different observations (Fig. 4.8 (a)-(b)). In the eastern Guinea Coast, there is an overestimation (underestimation) of the R95p and RX5day (CWD) anomalies by the GCMs (Fig. 4.8 (c)-(e)). Regarding the FRQW, there is a robust underestimation of its anomalies related to one standard deviation of the AEM index over West Africa (Fig. 4.8 (f)).

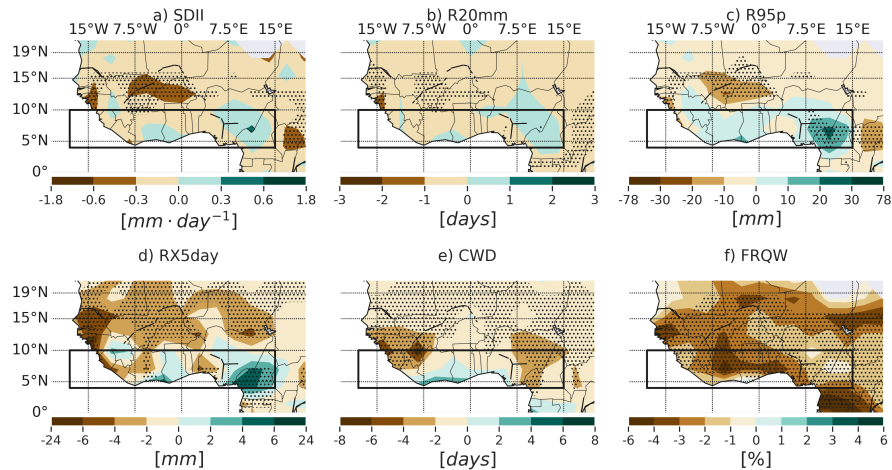


Figure 4.8: Spatial distribution of the EnsMedian bias relative to six observations for the JAS regression pattern of the extreme rainfall indices related to one standard deviation of the Atlantic equatorial mode. In total, there are six observations $\times$ 24 models = 144 different biases. The stippling indicates regions where two-third of the biases agree on the sign of the EnsMedian of the 144 biases. The black rectangles indicate the Guinea Coast region.

Figure 4.9 displays the performance metrics for the GCMs extreme rainfall indices related to the AEM. The EnsMedian values of the %BIAS correspond to -17 , -30 , 34.5 , 6 , -31 and -75 % for SDII, R20mm, R95p, RX5day, CWD and FRQW, respectively (Fig. 4.9 (a)). The biases for FRQW are the largest. Moreover, the NRMSE of R95p has the highest values, compared to the other variables (Fig. 4.9 (b)) with an EnsMedian value reaching 4. In more than 75 % of the cases, the PCC between the individual models and the different observations is lower than 0.5 (Fig. 4.8(c)). Particularly, the CWD anomalies are negatively correlated with the observation patterns in much of the models (the correlations are weak, however). According to the TSS, the response of the SDII index to AEM is better represented by the models, and this is indicated by a median TSS value of 0.4 among the set of $24 \text{ GCMs} \times 6 \text{ observations} = 144$ values (Fig. 4.8 (d)). This can also be seen from Fig. C.19, where much of the good performances of the different models relate to SDII. The individual %BIAS, NRMSE and PCC statistics for the individual models are available in Figs. C.16-C.18.

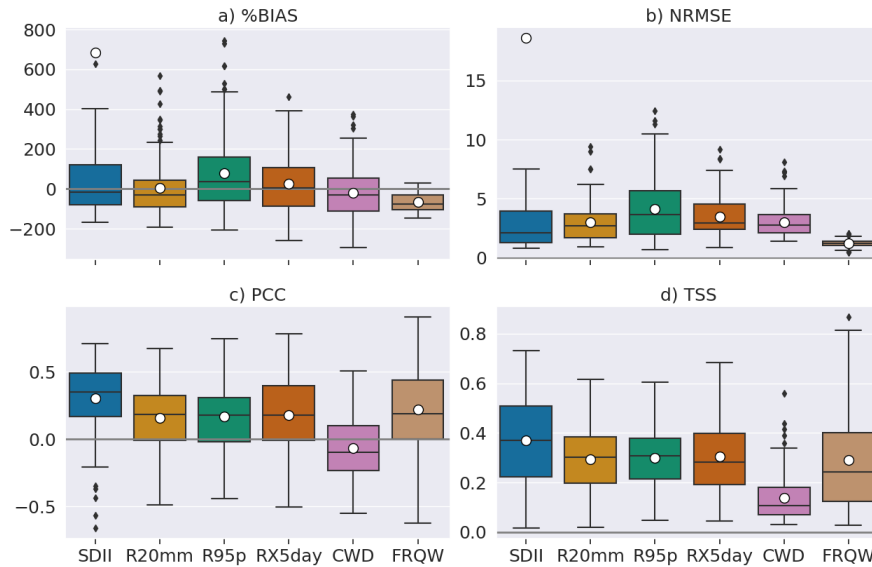


Figure 4.9: Performance of the 24 GCMs relative to the six rainfall observations, in representing the spatial distribution of extreme rainfall responses to AEM over the Guinea Coast. Each boxplot indicates the distribution of $24 \times 6 = 144$ different values. The median (mean) values of the statistics are represented by the horizontal bar (white circle). Outliers are plotted individually with the black markers. Note that all the outliers for SDII are not plotted, to improve the readability of the figure.

In summary, the anomalous warming of the eastern equatorial Atlantic during positive phases of the AEM leads to positive anomalies in the wet extreme indices over the Guinea Coast. This result is consistent with Atiah et al. [2020]; Diatta et al. [2020]. However, according to the Taylor skill score, the performance of the models in representing the anomalous responses in the different extreme indices is poor to modest, although some indices exhibit a good pattern correlation coefficient. Over Guinea Coast, the biases in the indices (except FRQW) are relatively small compared with the spread in the regression patterns of the observations (Fig. C.20). In the next section, the future changes in the Guinea Coast extreme rainfall indices related to the AEM will be assessed.

4.4.2 Future changes in rainfall extreme events associated with the AEM

The changes of the equatorial Atlantic mean state under the highest greenhouse gas emission scenario (SSP5-8.5) lead to a weakened variability of the AEM in the future [Worou et al., 2022; Crespo et al., 2022; Yang et al., 2022]. Worou et al. [2022] showed a projected weakening of the trade winds climatology over the equatorial Atlantic and a deeper thermocline in the eastern equatorial Atlantic which would lead to a reduction of the coupling between the surface and the thermocline depth. This implies a future weakening of the Bjerknes feedback [Bjerknes, 1969] in the equatorial Atlantic that explains the reduced variability of the AEM under global warming, and the weaker future impact of the AEM on the rainfall over the equatorial Atlantic and the Guinea Coast. These results are confirmed by Crespo et al. [2022], who found a future reduction of the AEM variability in a warmer climate, mainly due to a weakening of the third component of the Bjerknes feedback (the SST response to the variations of the thermocline depth). Yang et al. [2022] also found a reduced AEM variability. They underlined a greater role of a more stable tropical Atlantic atmosphere background in a future warmer climate [Jia et al., 2019], in reducing the AEM variability, compared to the weakening associated with the deepening of the eastern equatorial Atlantic thermocline. Subsequently, the variability of the JAS rainfall over the equatorial Atlantic and Guinea Coast that is related to the different phases of AEM is reduced [Worou et al., 2022].

The magnitude of the change in the Guinea Coast SDII-AEM EnsMedian is about $-0.1 \text{ mm} \cdot \text{day}^{-1}$ in the three future periods (Fig. 4.10(a)). The EnsMedian values of the percentage of changes in the average of SDII response to AEM over Guinea Coast correspond to -37% , -32% and -87% in the near-term, mid-term and long-term periods, respectively (Fig. C.21). However, less than two-third of the models agree on the sign of these changes, whatever the future period considered. The spatial distribution of the long-term change in the SDII anomalies over the Guinea Coast indicates values that reach $-0.4 \text{ mm} \cdot \text{day}^{-1}$, which is robust only

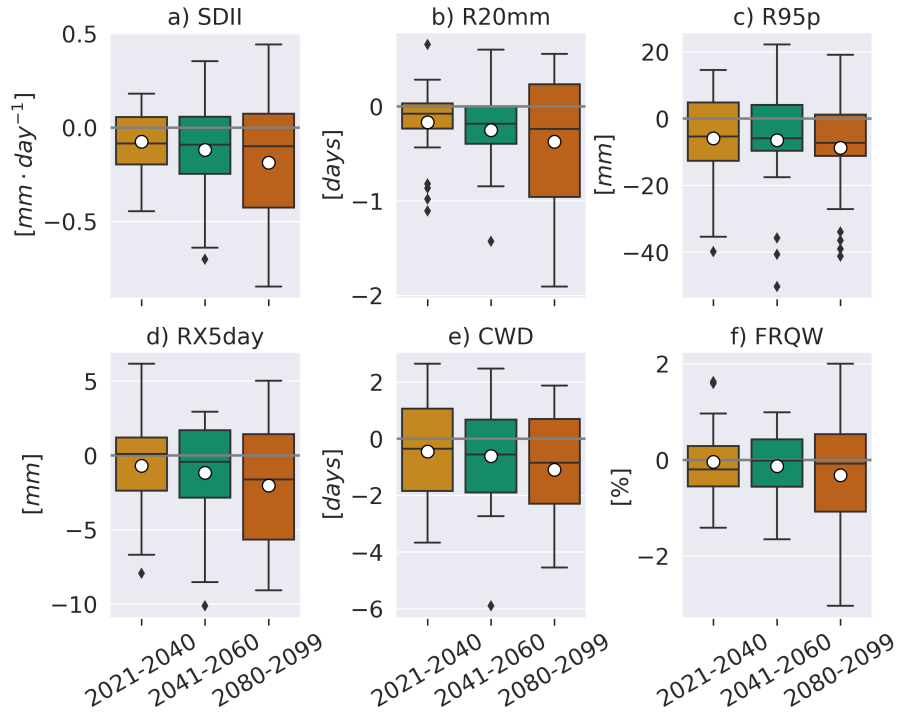


Figure 4.10: Near-term, mid-term and long-term changes in the Guinea Coast area average of the JAS extreme rainfall responses to one standard deviation of JAS AEM index. Each boxplot represents the distribution of 24 GCMs. The median (mean) value of each distribution is indicated by a black horizontal bar (a white circle).

over the central Guinea Coast (Fig. 4.11(a)). The maps corresponding to the near-term and mid-term changes in the regression patterns are available in Figs. C.22-C.23. These maps show weak changes in regression patterns over Guinea Coast, compared to the long-term future changes. The regression maps corresponding to the four different periods are also displayed in Fig. C.24.

The number of days with heavy and very heavy precipitation associated with one standard deviation of the AEM also decreases in the future periods compared to the present-day period (Fig. 4.11(b)-(c)). The average of the R10mm (R20mm) changes relative to 1995-2014 over the Guinea Coast equals -0.28 (-0.18), -0.29 (-0.22) and -0.61 (-0.33) days in the 2021-2040, 2041-2060, 2080-2099 periods, respectively (Fig. 4.10 (b)-(c)). The multi-model median values corresponding to the percentage of changes in the area averaged R10mm (R20mm) regression coefficients over the Guinea Coast lie between -28% and

−72% (−37% and −86%) over the three future periods.

The spatial patterns of the future changes in the variability of rainfall during the very wet days (R95p) that is related to the warm and cold AEM events indicate a reduction in R95p magnitudes over most of the Guinea Coast areas west of 5°E (Fig. 4.11(c)). Elsewhere over Guinea Coast, there is no evidence of a robust change in the R95p pattern. The EnsMedian change in the average of R95p over the Guinea Coast ranges from −5.3 mm to −7.3 mm over the three future periods (Fig. 4.10(c)). The multi-model median values of the Guinea Coast R95p change percentage (and the agreement among the models) corresponds to −41% (62%), −38% (67%) and −50% (67%) in the near-term, mid-term and long-term future periods, respectively, relative to the present-day period (Fig. C.21 (c)).

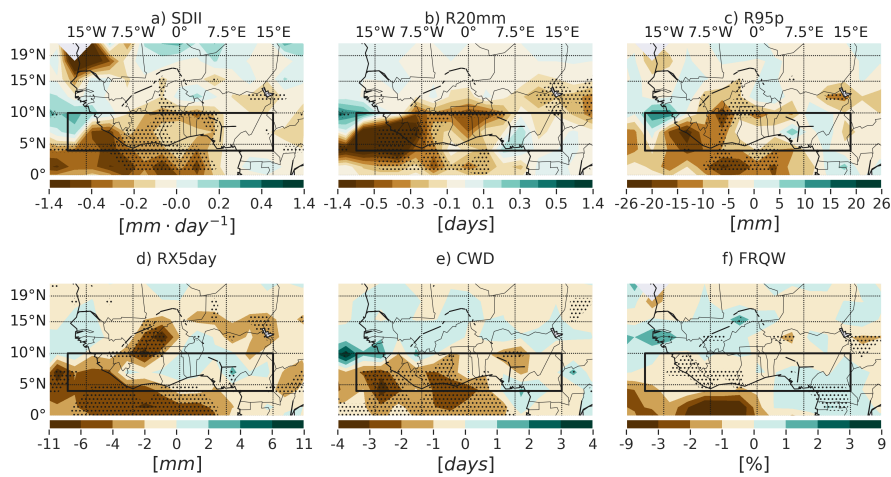


Figure 4.11: Long-term changes in the regression patterns of the JAS extreme rainfall indices associated with the standardized JAS AEM SST index (2080-2099 minus 1995-2014). The stippling indicates grid points where two thirds of the models agree on the sign of the change.

Lower maximum five days precipitation related to one standard deviation of the AEM index averaged over the Guinea Coast are projected in the 2021-2040, 2041-2060 and 2080-2099 periods, compared to their magnitude in the 1995-2014 period (Fig. 4.10(d)). Over the mid-term and long-term future periods, the RX5day EnsMedian reduction is between −0.4 and −1.6 mm. The near-term EnsMedian change is positive but very close to zero (0.1 mm). In the three cases, there is a low agreement among the models in the sign of the change (less than 55%).

The multimodel EnsMedian values of the percentage of change in the RX5day averaged over the Guinea Coast range between -24% and -58% over the three future periods (Fig. C.21 (d)). Furthermore, Fig. 4.11(d) shows no robust change signal in the spatial distribution of the RX5day anomalies over Guinea Coast.

The projected decrease in the AEM variability in the future leads to a decrease in the magnitude of the wet spells corresponding to one standard deviation of the AEM index. However, the EnsMedian change in the anomalous responses of these indices averaged over the Guinea Coast is less than one day in the different three future periods (Fig. 4.10(e)). The EnsMedian value of the CWD change percentages averaged over the Guinea Coast (as well as the percentage of agreement) approximate $-56\%(54\%)$, $46\%(67\%)$ and $-71\%(71\%)$ for the near-term, mid-term and long-term future periods, relative to the present-day period (Fig. C.21 (e)). These area-average statistics are robust for the mid-term and long-term future changes, according to the two-third sign agreement metric. Figure 4.11(e) displays the spatial distribution of the long-term changes in the CWD response to AEM. It also shows a robust decrease in CWD magnitudes over Guinea Coast, which is between one and three days over the central and western Guinea Coast.

Finally, a reduction in the positive anomalies of the frequency of wet days over the Guinea Coast that are linked with one standard deviation of the AEM is projected in the future, under the SSP5-8.5 scenario. Robust long-term reduction up to 1% are detected in the western Guinea Coast. In the eastern part of the region, non-robust weak long-term changes are projected (Fig. 4.11(f)). The averages of the change patterns over the Guinea Coast are equal to -0.2% , -0.0% and -0.1% , for the three consecutive future periods, respectively, relative to the present-day period (Fig. 4.10(f)). The multi-model median value of the percentage of the change in the FRQW-AEM index averaged over the Guinea Coast ranges between -20% and -38% over the three future periods (Fig. C.21 (f)), but these values are non-robust. About half of the 24 GCMs agree on the sign of the EnsMedian change, which makes the result uncertain.

To quantify the contribution of AEM in the total variability of the rainfall extremes over Guinea Coast, we correlate the extreme index averaged over Guinea Coast onto the standardized AEM index. The square of the correlation coefficient gives the variance explained fraction (VEF). Figure 4.12 shows the distribution of the VEF across 24 GCMs, for each extreme index and for each period of the current study. For each of the extreme index, there is a clear decrease of VEF by the AEM between the present-day period and the long-term future. However, change in the VEF is not linear with time, and this could be due to chance in the small set of models used in our study. Under present-day conditions, the EnsMedian VEF ranges between 13 % and 28 %, for the six indices. In the long-term future, the VEF EnsMedian does not exceed 8 %. This indicates a weaker role of AEM in the last decades of 21st century, under SSP5-8.5 scenario.

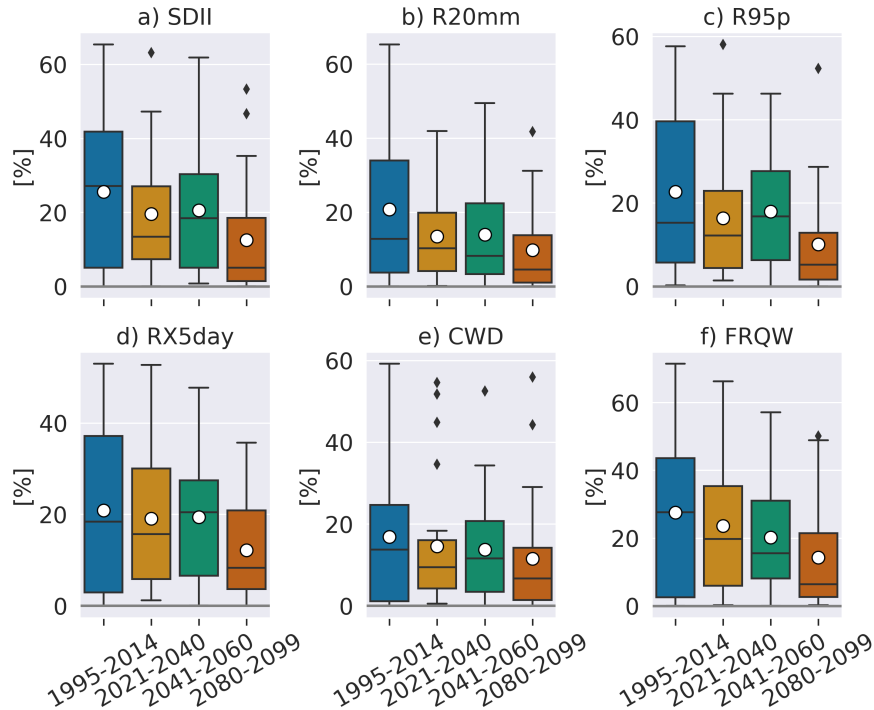


Figure 4.12: Regression of the GCR extreme indices regressed onto the standardized AEM index: proportion of the variance explained by the AEM over four different periods (in %). Each boxplot represents the distribution of 24 GCMs. The median (mean) value of each distribution is indicated by a black horizontal bar (a white circle).

In summary, subsequent to the decrease in the AEM variability in a future warmer climate relative to the present-day conditions, there is a decrease in the variability of the extreme rainfall indices associated with the AEM. For the majority of the indices, the average of the changes in the teleconnection amplitude between the Guinea Coast and the eastern equatorial Atlantic indicates an overall enhancement in magnitude with time. Crespo et al. [2022] demonstrated that the models with the strongest warm sea surface biases exhibit the weakest change in both the AEM variability and the third component of the Bjerknes feedback. Given the positive SST biases in most of the CMIP6 models in the eastern Atlantic Basin [Richter and Tokinaga, 2020; Worou et al., 2022; Crespo et al., 2022], improving these biases would likely increase the change in the AEM variability, suggesting that the projected weakened AEM variability is robust despite these biases. Moreover, despite the increase in the total variability of the majority of the extreme

indices under global warming in the future (Section 4.3.3), there is a reduction in the proportion of variance explained by the AEM for these extreme indices. This result highlights a weaker role played by the AEM in explaining the future rainfall extreme variability over the Guinea Coast. Furthermore, there is more confidence in the results averaged over the Guinea Coast than in the spatial distribution of the changes in the teleconnection patterns.

4.5 Summary and conclusions

In this study, the performance of 24 GCMs participating in CMIP6 in simulating six JAS rainfall-based climate extremes indices over the Guinea Coast was assessed as well as the projected mean changes in these extremes under global warming. We used six different observation database to evaluate the models. Historical and SSP5-8.5 simulations were considered to study changes in the near-term (2021-2040), mid-term (2041-2060) and long-term (2080-2099) future periods, compared to the present-day climate period (1995-2014).

Under present-day conditions, we found that the GCMs EnsMedian simulates less intense rainfall events. All the models exhibit too frequent rainfall and longer wet spells over the Guinea Coast, compared to the observations. These models show a wide range of FRQW spatial correlation with the different observations, which makes this variable uncertain in the models, as indicated by its poorest skill score (0.26 for the EnsMedian), compared to the other variables. On the other hand, the R20mm is better represented in the models, according to the distribution of the PCC and TSS values in the GCMs. The EnsMedian values for these metrics are 0.7 and 0.6, respectively. For the other variables, the EnsMedian PCC (TSS) values are between 0.5 and 0.6 (0.4 and 0.5). Those results indicated that despite the good performance of the models for some aspects of the rainfall characteristics, there is still a need to continue the effort toward an improvement of the GCMs to better represent the mean state of the West African hydroclimate. This would lead to a more reliable use of climate models for climate services over West Africa, in order to implement better mitigation and adaptation strategies to climate change.

Under the SSP5-8.5 scenario, changes in the near-term, mid-term and long-term (2080-2099) periods are evaluated, relative to the present-day climate conditions. Consistent with previous studies [Dosio et al., 2021; Wainwright et al., 2021], results of the average of the mean changes in the extreme indices over the Guinea Coast indicate an intensification of the daily rainfall, together with a decrease in the frequency of wet days and the duration of wet spells. The number of days with precipitation exceeding 20 mm is increased, as well as the rainfall amount during the very wet days and the maximum rainfall amount over five consecutive days.

The area average of the extremes indices over Guinea Coast indicates an intensification of extreme conditions in the projections, with a gradual increase in magnitude over the three future periods. However, it is important to note that the spatial distribution of the changes is not uniform over the region, and the robustness of the changes at a grid point varies among different indices and the future period considered. In general, internal variability is large enough to mask the forced signal during the first decades of the projection over Guinea Coast. In the mid-term changes, only a few areas in the western Guinea Coast exhibit a robust change in SDII, R20mm and RX5day. In the long-term projections, the R95p projections remain uncertain over Guinea Coast, while robust change signal with a uniform sign can be found for the other extreme indices. These conclusions are in overall agreement with Monerie et al. [2017], who estimated the different numbers of models ensemble members needed to resolve the forced signal in the projections over Sahel. This kind of study needs to be performed over Guinea Coast. It is, however, beyond the aim of our current study, which is rather focused on teleconnection processes.

Anomalous warming (cooling) of the eastern equatorial Atlantic in positive (negative) phases of the AEM is associated with above (below) normal values of extreme rainfall indices over Guinea Coast in the current climate. Accordingly, our results indicate an in-phase relationship between the wet extreme indices of the Guinea Coast and the AEM SST index. However, over Guinea Coast, the multi-model EnsMedian spatial distributions of the extreme rainfall anomalies related to one standard deviation of the AEM under the present-day conditions are not significant for CWD, and limited to few areas very close to the coastal line for FRQW. These two indices present the poorest TSS values, in term of performance of the 24 GCMs relative to the six different observations, and compared to the other extreme indices. The index that has the closest pattern to the observations is SDII. The distributions of the TSS values for R20mm, R95p and RX5day indices are similar, with an EnsMedian value close to 0.3. The six observations are consistent with the sign of the regression patterns of the different extreme indices over Guinea Coast. However, the biases in the GCMs's patterns vary a lot through the 24 models and the six observations, leading to no consensus on the sign of the bias, except for the FRQW for which there is a consistent underestimation of the responses to AEM.

Additionally, there is a projected weakening of the AEM under global warming. Subsequently, there is overall a projected decrease in magnitude of the Guinea Coast rainfall extreme responses to this mode of climate variability in the three future periods, relative to the present-day situation. However, the spatial distribution of the changes in the regression patterns is different from one future period to another period and generally non-robust among the models. These results are also limited by the poor skill of the models in representing the extremes responses

to AEM under present-day conditions. Moreover, for a given future period, the sign and the significance of the change are highly heterogeneous in space. As an example, there is a mid-term robust increase in the RX5day response to AEM over the center of Guinea Coast, whereas no significant changes are found over the entire region for the near-term and long-term periods. In general, there is no robust changes in the majority of the indices for the near-term and mid-term periods. The long-term changes are, however, stronger in magnitude with more robustness over Guinea Coast, according to the sign agreement among the models. Thus, in general, it is questionable whether it makes sense to perform an average of non-robust changes over the Guinea Coast region, for a given variable. Moreover, differences among the models' responses could be highlighted by grouping them in different categories based on different criteria. According to the poorest representation (in general) of the extreme responses to AEM under present-day conditions, one could choose to separate models that project an increase in the magnitude of the responses, from the models that project a decrease in the extreme responses to AEM. This is not done in our current study. Another solution would be to use few GCMs with 10 or more ensemble members at high spatial resolution, with a good representation of the AEM impact on West Africa. The increase in the spatial resolution will allow the study of regional differences within the Guinea Coast region.

Despite the biases and uncertainties, the area average of the changes over Guinea Coast is more robust, and gives an overall tendency. However, although there is an overall increase in the total variance of most of the extreme indices over the Guinea Coast, the contribution of the AEM to explain the variance of these extreme indices is reduced. This reduction is clear if the long-term variance explained fraction is compared to its values in the present-day situation. This result suggests a weaker role of the AEM in driving the extreme rainfall variability over the Guinea Coast in a future scenario of high emission of greenhouse gases. Therefore, there is a need to identify other oceanic and atmospheric modes of variability or other processes that explain the future changes in the extremes' variability over the Guinea Coast. These processes should compensate for the decreasing role of the AEM. For instance, Akinsanola et al. [2020] argued that the changes in the circulation (the dynamics) should contribute less to the increase in the rainfall variability over West Africa, while the local thermodynamics should be the dominant factor of these changes.

Our conclusions are based on 24 CMIP6 GCMs that have clear biases and whose resolution is too coarse to represent well some important processes controlling extremes and their changes. It would thus be useful to reevaluate our results with RCMS or even convection permitted models. The processes that drive the biases in the mean state and teleconnection patterns also need to be better understood and better represented in order to gain more reliability on the projected

changes in the rainfall extremes over the Guinea Coast. Additional insights on the impact of the internal variability on the projected changes over the Guinea Coast could also be gained by using more GCMs or more ensemble members within one GCM, to assess the impact of the internal climate variability on the projected changes over Guinea Cast.

5

Conclusions and perspectives

5.1 Main achievements

The West African rainfall variability shows various spatio-temporal characteristics that need to be understood in order to provide better mitigation and adaptation strategies in a changing climate to policy-makers. While many studies are focused on the Sahelian rainfall variability, very few deals with the variability of the Guinea Coast hydroclimate. This thesis aims to put light on the dynamics at work in this coastal region of West Africa during the last century and in a warmer future. In the introduction of this manuscript, we raised this general question: **What is the role of oceanic internal modes of variability in driving the past, present and future Guinea Coast rainfall variability on the interannual timescale?** Three main points were developed in order to answer this question. Below, we summarize the main outcomes.

5.1.1 What are the past, present and future impacts of the AEM on the interannual Guinea Coast rainfall variability during the boreal summer?

During the monsoon season of West Africa, the AEM appears as the leading oceanic internal mode of variability of the rainfall fluctuations over the Guinea Coast, at least during the last century. The summer link between the AEM and the Guinea Coast rainfall (GCR) is the strongest and stable SST-GCR relationship. This result was highlighted in Chapter 2. Thus, the equatorial Atlantic appears as a precious predictor for the Guinea Coast hydroclimate, which could be of great interest for forecasters as well as for climate projections over the Guinea Coast.

Above normal surface warming in the eastern equatorial Atlantic leads to a positive rainfall anomaly over the Guinea Coast, and inversely.

The study of climate change involves not only the changes in the mean climatic variables, but also the changes in their variability around the mean. In that framework, climate models are useful to understand the changes in the climate system from the past to the future. In Chapter 3, we developed a metric, based on the sign of the correlation between SST and rainfall anomalies, to evaluate the capability of 31 GCMs participating in CMIP6 to simulate the boreal summer link between the AEM and the GCR under present-day conditions. This evaluation is necessary, as we learned from the Chapter 2 that the impact of the eastern equatorial Atlantic SST anomalies on the rainfall anomalies over the Guinea Coast dominates the influence of the other oceanic regions during the monsoon season. Analysis of the CMIP6 historical simulations indicated that the majority of current GCMs (24 models) perform well in simulating the positive correlation between the GCR and the eastern equatorial Atlantic SST fluctuations on the interannual timescale. However, the spatial variability of the West African rainfall response to the AEM is stronger in GCMs, compared to observations. Furthermore, we noted that the main differences between the GCMs that are able to simulate the good sign of the correlation between the AEM and the GCR from the others come from the atmospheric response to the AEM phases. In the GCMs which simulate the good AEM-GCR relation, there is a larger extension of the zonal and meridional moisture transport from the Atlantic Ocean toward the Guinea Coast in response to a positive AEM phase, compared to the other GCMs. In these latter, the positive moisture transport is limited to the oceanic region. These encouraging results brought us to study the likely changes in the AEM impact on the GCR in a warmer future climate.

This was done by comparing the climate change projections under the highest greenhouse-gas emission scenario SSP5-8.5 to the historical simulations conducted by the CMIP6 GCMs. Three main periods were considered for the future: the near-term (2015-2039), mid-term (2040-2069) and long-term (2070-2099) futures. Results indicate a weaker role of the AEM in driving the rainfall over Guinea Coast in a warmer future, compared to nowadays climate. This means that our first-class oceanic predictor of the interannual Guinea Coast rainfall will leave the floor to other processes such as the thermodynamic changes in specific humidity [Seager et al., 2010] over the Guinea Coast in a warmer future. We showed that the reduced GCR variability associated with the AEM is due to a reduction of the AEM variability in a warmer climate. We found a weakening of the Bjerknes feedback over the equatorial Atlantic, due to the changes of the equatorial Atlantic mean state. Mainly, under future global warming, the eastern equatorial Atlantic mean thermocline is deeper, and is less sensitive to the fluctuations of the surface temperature. It results in a weakening of the third component of the Bjerknes

Feedback, and a reduction of the AEM variability. This result was further confirmed by [Crespo et al., 2022]. However, Yang et al. [2022] evinced a greater role of a more stable tropical Atlantic atmosphere in the mean warmer future climate, compared to the change in the mean eastern equatorial thermocline, in reducing the AEM variability. At last, our analysis revealed that the magnitude of the rainfall response to the AEM decreases progressively with time in the future.

5.1.2 What are the present and future impacts of the AEM on the extreme rainfall events over the Guinea Coast during the monsoon season?

In Chapter 4, we extended the study of Chapter 3 to the impact of AEM on the extreme rainfall indices over the Guinea Coast during the boreal summer. We showed that biases exist in the representation of the JAS extreme indices climatology by the GCMs over Guinea Coast. These biases are in general of the same order of magnitude as the spread of the observations, except for CWD and FRQW which present the largest biases compared with the spread in the observations. Moreover, the models struggle to simulate the variability of the extreme indices related to the AEM over the Guinea Coast. The performance of the models in simulating the spatial distribution of the extreme indices responses to AEM is weaker than the one obtained for the spatial distribution of the climatology. The biases in the extreme responses can be as large as the spread in the observations.

From the reduction in the AEM variability under warmer future climate conditions (a result obtained in Chapter 3), we expect not only a reduction in the total rainfall response to this oceanic internal mode of variability, but also a decrease in the AEM impact on the variability of the extreme rainfall indices. This was confirmed by our analyses in Chapter 4. The variability of the extreme indices related to the AEM is reduced at the end of 21st century, following the SSP5-8.5 scenario, despite an overall increase in the total variability of these indices.

5.1.3 What is the role of the oceanic internal modes of variability on the rainfall over the Guinea Coast in the other seasons during the last century?

Beyond the monsoon season, we found in Chapter 2 some connections between the GCR and SST modes of variability. In the pre-monsoon season, the development of the cold tongue in the eastern equatorial Atlantic is favourable to wetter conditions over the Guinea Coast. In addition, over the last century, the South Atlantic Ocean Dipole (SAOD) anti-correlation with the GCR in the pre-monsoon season is the strongest. We noted a reversal of the SAOD-GCR correlation after the 1980s, when the link between the eastern equatorial Atlantic and the eastern

Pacific basin has started to strengthen. This result highlights again the important role of the adjacent equatorial Atlantic SST in the onset of the first rainy season over the Guinea Coast on the interannual timescale.

During the post-monsoon season, the eastern Mediterranean basin shows the strongest SST-GCR linkage during the 20th century, mainly after the 1950s. This finding was highlighted for the first time in this study, as well as the dynamics associated with. In particular, we pointed out the role of the oceanic extension of the Saharan Heat low in forcing the low-level westerly jet, which, in turn, advects moist air into the Guinea Coast.

Lastly, a more global tropical basin connection with the GCR was found in the harmattan season, with a stronger anti-correlation with the Indo-Pacific Warm pool (IPWP) SST index, compared to the other SST-GCR relationships. The dynamics involved is similar to the one in the forcing of an El-Niño phase on the rainfall over West Africa.

5.1.4 General conclusions

In this thesis, we showed that the Guinea Coast rainfall has been influenced by tropical and extratropical oceanic basins over the 1901-2016 period, with different SST patterns according to the season (Chapter 2). We found strong links with the equatorial and South Atlantic Ocean in the pre-monsoon and monsoon seasons, while the eastern Mediterranean Sea has been in phase with the post-monsoon rainfall over the Guinea Coast. For the harmattan season, we can consider that a wide tropical basin has influenced the GCR, with a stronger connection with the IPWP. These findings have the potential to be used for the predictability of the seasonal rainfall over the Guinea Coast. Although we found physical mechanisms to explain each strong connection during each season, we believe that these relationships could suffer from confounding effects, mainly for the harmattan and post-monsoon seasons. Moreover, our regressions were based on synchronous timeseries, without considering any delayed effect of the climate system. This could explain partly the weaker correlations obtained in the seasons other than the monsoon season, which is a limitation of our study.

For the future, we do not know exactly what will be the trajectory of our fossil fuel consumption and greenhouse-gas emissions. However, simulations performed with climate models can help to anticipate the future changes under different scenarios. Despite the biases in the CMIP6 models in the representation of the mean tropical Atlantic SST and rainfall, they perform reasonably well in simulating the AEM SST pattern as well as its impact on the Guinea Coast rainfall under present-day climate conditions. For the future, our results suggest that the role of the AEM in driving the GCR will be reduced, as well as the AEM impact on the extreme rainfall occurrence over the Guinea Coast (Chapters 3 and 4).

Compared to the existing literature, this thesis showed different teleconnection processes involving the rainfall over the Guinea Coast and oceanic basins for different seasons, which is original. We found new relevant SST-GCR linkages. In particular, we pointed out the role of the eastern Mediterranean Sea during the post-monsoon season, the influence of the IPWP during the harmattan season, and the tropical south Atlantic during the pre-monsoon season. The second originality of our thesis rely on the reduction of the AEM impact on the rainfall and extreme rainfall indices over the Guinea Coast, under future global warming compared to the present-day climate conditions.

Changes in the SST variability in the eastern equatorial Atlantic will less affect the rainfall interannual variability over Guinea Coast under the SSP5-8.5 scenario. The total forced changes under this scenario indicate, however, a northward shift of the West African monsoon system during the summer, at the end of the 21st century (Fig. 5.1). This shift is a result of the intensification of land-surface warming over North Africa, which is translated into a deepening of the low-pressure system over the region [Monerie et al., 2020]. The land-sea pressure gradient is enhanced and favors anomalous southwesterlies which advects moist air into the central and western Sahel.

Over the Atlantic Ocean, the relative warmer SST over the equator compared with the other sectors in the long-term future (Fig. 5.2) shifts the rainfall band to the south (Fig. 5.1). This leads to negative rainfall anomalies over Western Sahel. In the southern part of Guinea Coast, negative rainfall anomalies are the result of the shift of the monsoon system to the north [Monerie et al., 2020].

5.2 Perspectives

As highlighted in the previous section, our analyses in Chapter 2 were based on linear regressions that cannot eliminate confounding effects from different oceanic or atmospheric drivers. A natural extension of this work would consider, on the one hand, inferring causality from the linear regression framework [Pearl, 2013; Kretschmer et al., 2020; Adams et al., 2021], in the analysis of the past and present climate variables. This will help to understand the right pathway of the causal link between different variables: for example the link between the post-monsoon GCR, the SST in the eastern Mediterranean Sea and the low-level westerly jet. Such an analysis can identify if the low level westerly jet is a mediator of the influence of the Mediterranean Sea on the Guinea Coast rainfall. In general, this framework would help to understand if a variable is a common driver, a mediator or a common effect of other variables. Thus, the nature of the teleconnection can be more clarified with a good physical reasoning. Furthermore, the study of the non-linearity in the atmosphere-ocean interactions could be explored deeper by analysing a large ensemble of seasonal hindcasts, which offers more realizations

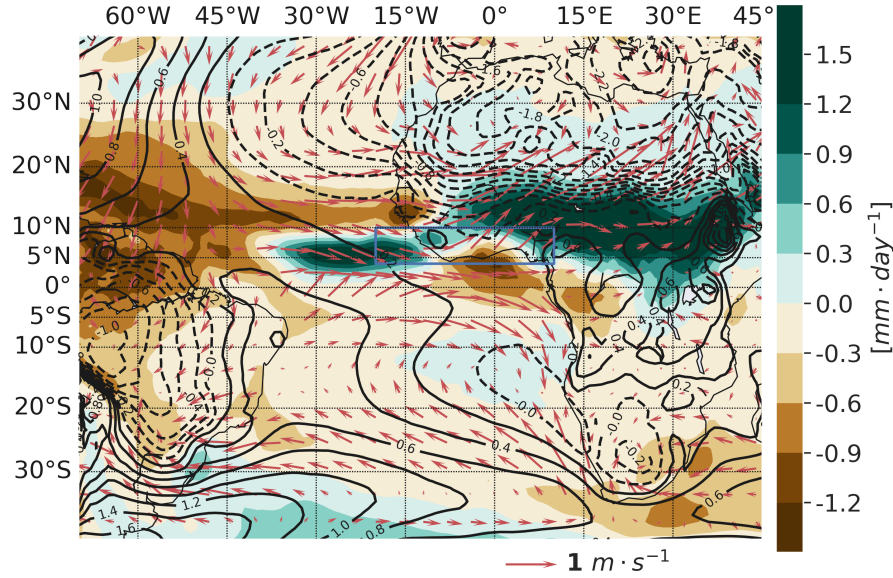


Figure 5.1: Mean long-term JAS changes (2070-2099 minus 1985-2014) in the precipitation (in colors), 10 m horizontal winds (arrows) and mean sea level pressure (contours, in hPa). The blue rectangle indicates the Guinea Coast region. Data sources are indicated in Table 3.1

of the atmosphere-ocean coupling than in the real observations of the climate system [Osman et al., 2022, e.g.,]. Studies of teleconnection processes based on observations and reanalyses alone are complicated by the limited observation records, which add a difficulty in the control of confounding effects [Byrne et al., 2019]. An ensemble of hindcast provides a forecast of the past, with different initial conditions to characterize the chaotic aspect of the climate system. Thus, increasing the sample size can help to perform composite analyses on different SST indices and to explore the effect of an in-phase or out-of-phase relationship between the main oceanic modes of variability (e.g. different combinations of El-Niño and AEM positive, negative and neutral phases) on the Guinea Coast rainfall. Such an analysis could be done for the different seasons, and could also consider some time lags between several climatic variables [Monnin et al., 2021].

In Chapter 2, the eastern Mediterranean basin was pointed out as the main oceanic driver of the post-monsoon rainfall variability over the Guinea Coast. To go forward, we could perform sensitivity experiments with GCMs to test the impact of the Mediterranean SSTs on the GCR in the post-monsoon season. One could design a set of idealized warm (cold) SST patterns restricted to the Mediter-

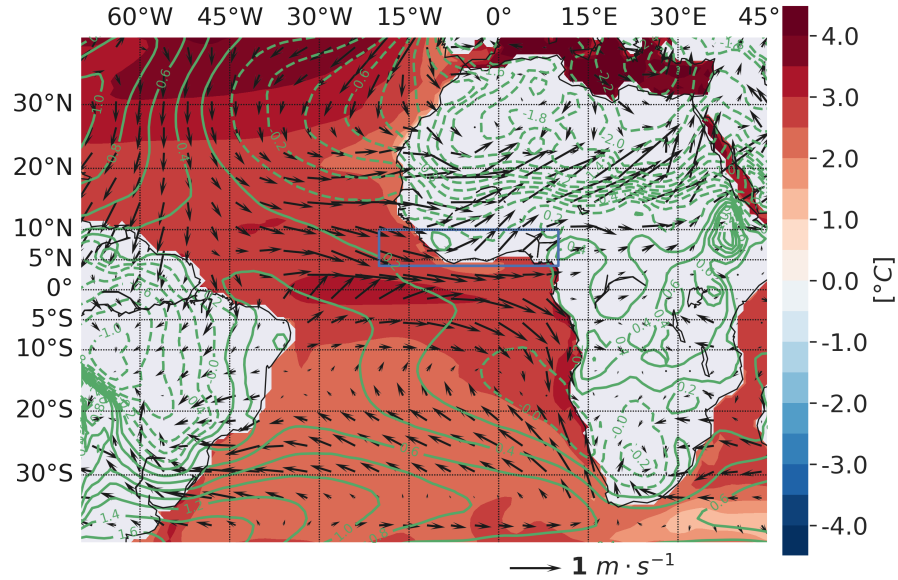


Figure 5.2: Mean long-term JAS changes (2070-2099 minus 1985-2014) in the SST (in colors), 10 m horizontal winds (arrows) and mean sea level pressure (contours, in hPa). The blue rectangle indicates the Guinea Coast region. Data sources are indicated in Table 3.1

anean Sea, which can then be used to force the GCMs. The rainfall response over the Guinea Coast could then be analysed. This work can be based on Rowell [2003]; Fontaine et al. [2009] experimental designs or others [Mohino et al., 2011b; Rodríguez-Fonseca et al., 2011]. These SST sensitivity experiments would thus increase the confidence of our findings. Similar integrations could be conducted for the harmattan season, to quantify the tropical basin influence on the GCR during that season. Furthermore, studies showed that this basin has played a major role in the recovery of the rainfall over the Sahel in the last decades [Park et al., 2016]. Therefore, a further step would be the evaluation of the remote impact of this basin on the rainfall over the Guinea Coast area with the GCMs participating in CMIP6, as well as the changes in this teleconnection under future global warming.

In the analysis of the projected changes in the AEM influence on the GCR (Chapter 3), we stopped the future period at 2100. We suggest extending the analysis to 2300, in order to verify if the AEM variability continues to decrease, or if it stabilizes, as well as its impact on the GCR. The linkage between the AEM and El-Niño could also be assessed in order to see if their covariability is enhanced or not, as well as their impacts on the West African rainfall.

A simple way to consider time-lag in the remote SST anomaly impacts on the seasonal (other than JJAS) rainfall in the Guinea Coast region would be the use of the Extended Maximum Covariance Analysis (EMCA) [Polo et al., 2008] between a chosen oceanic area and the rainfall over the Guinea Coast. One could consider for instance monthly rainfall of a given season, and SST data that cover, for each year, a period starting (ending) few months before (after) the rainfall season considered. This will help to get a time-evolving SST patterns related to the leading mode of covariability between the GCR and SSTAs. This kind of analysis has been done for the June-September rainfall over West Africa [Polo et al., 2008].

Finally, we studied the impact of the AEM on the extreme rainfall occurrence under the present-day and future climate conditions. Although GCMs simulate relatively well some of the JAS extreme indices over the Guinea Coast, there is a need to reduce the biases in the spatial distribution of these indices. This could be done by understanding and improving the land-atmosphere interactions over West-Africa, and their coupling with oceanic processes. The majority of the extreme indices show robust responses to the AEM over the oceanic region south to the Guinea Coast. This is probably related to the southward position of the mean ITCZ over the tropical Atlantic in the GCMs (during the boreal summer), an issue which needs to be addressed in the next generation of GCMs. Another way to tackle this problem is the use of outputs from Regional Climate Model (RCM) simulations, such as those in the Coordinated Regional Climate Downscaling Experiment (CORDEX). They have the advantage to resolve more small scale processes and Akinsanola et al. [2020] showed that the RCMs better represent the spatial distribution of extreme rainfall indices climatology over West Africa. Moreover, the spatial distribution of the extremes responses to AEM simulated by the GCMs is robust over Guinea Coast, but the performance compared to the observations is modest.

A

Supplement to Chapter 2

A.1 Supplementary figures

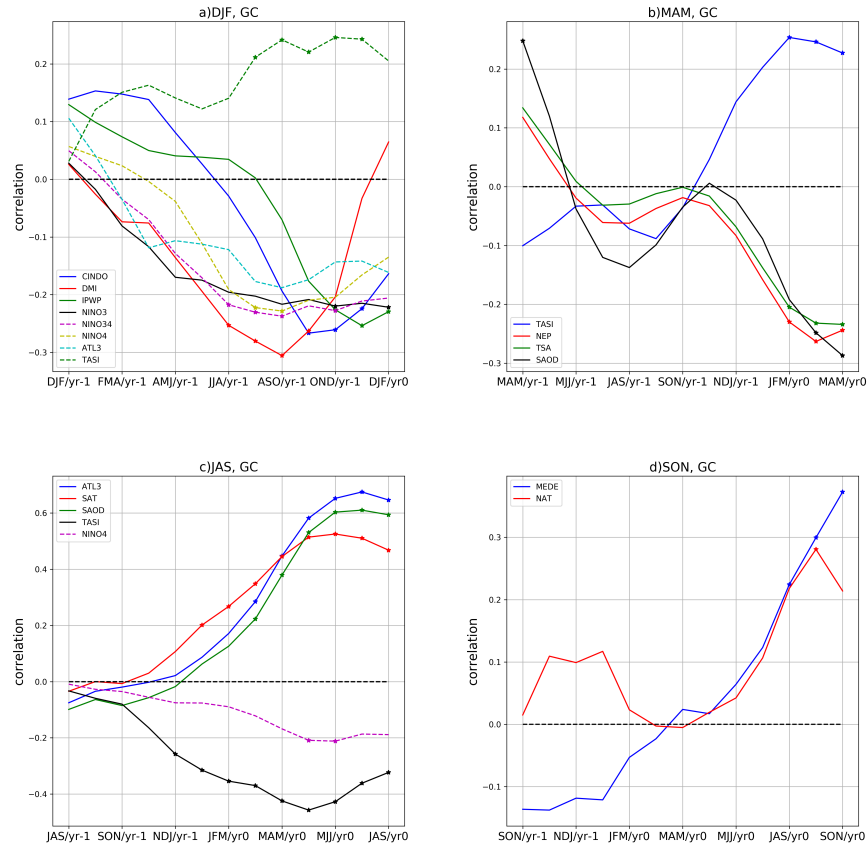


Figure A.1: Lagged correlation between each seasonal Guinean Coast rainfall index and each consecutive 3-months averaged SST indices, prior to the season considered: (a) DJF, (b) MAM, (c) JJAS, (d) ON.

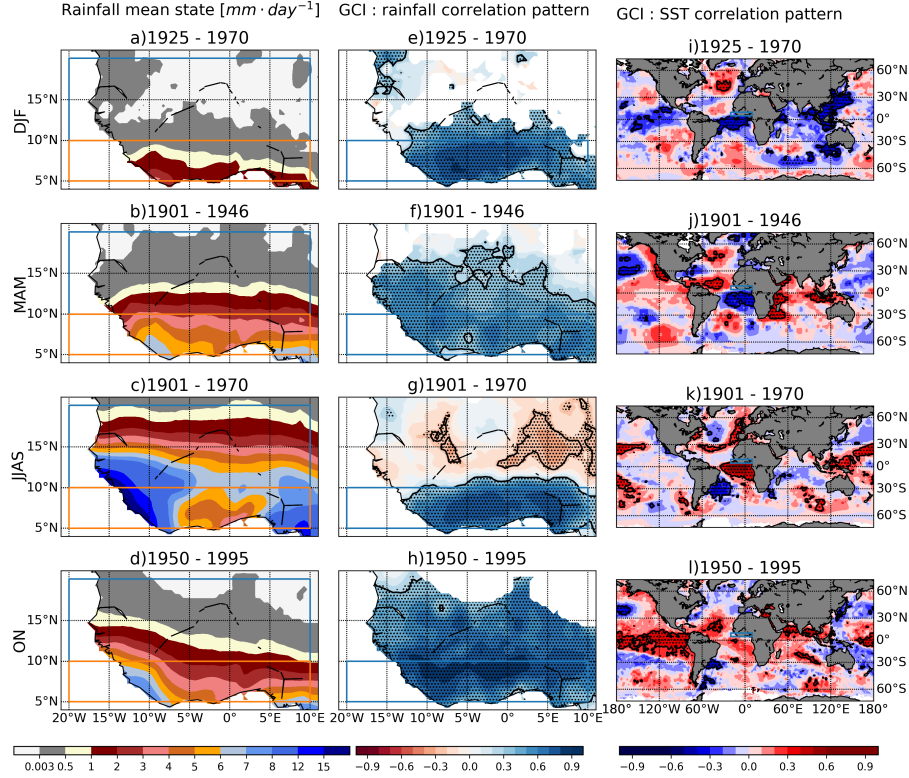


Figure A.2: (a,b,c,d) Seasonal means of rainfall over West Africa. (e,f,g,h) Correlation between the high frequency Guinean Coast rainfall index and the seasonal rainfall anomaly at each grid point. (i,j,k,l) Correlation between the high frequency Guinean Coast rainfall index and SST anomalies for each season. Contour lines and dots in the correlation patterns represent significant values at 95% confidence level according to an *F*-test with respectively a standard degree of freedom and the effective degree of freedom. For each season, analyses are performed over periods of the most robust SST-rainfall relationship. For DJF, the IPWP-GCR relationship is stationary over 1925 - 1970. For MAM, the tropical south Atlantic and the GCR connection is stable over 1901 - 1946. For JJAS, the link between ATL3 and GCR is stable over 1901-2016. However, the main change in the ATL3 influence on the Sahelian rainfall occurs after 1970s. For ON, the link between MEDE and the GCR is stationary after 1950s.

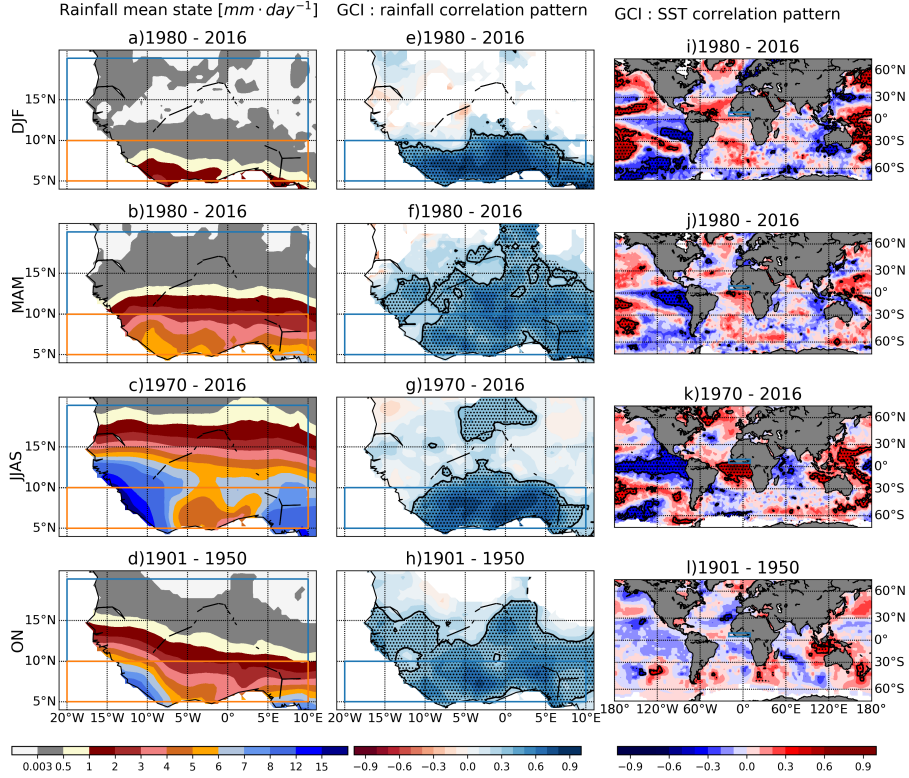


Figure A.3: (a,b,c,d) Seasonal means of rainfall over West Africa. (e,f,g,h) Correlation between the high frequency Guinean Coast rainfall index and the seasonal rainfall anomaly at each grid point. (i,j,k,l) Correlation between the high frequency Guinean Coast rainfall index and SST anomalies for each season. Contour lines and dots in the correlation patterns represent significant values at 95% confidence level according to an *F*-test with respectively a standard degree of freedom and the effective degree of freedom. For each season, analyses are performed over periods of weakening of the robust connection obtained in each season.

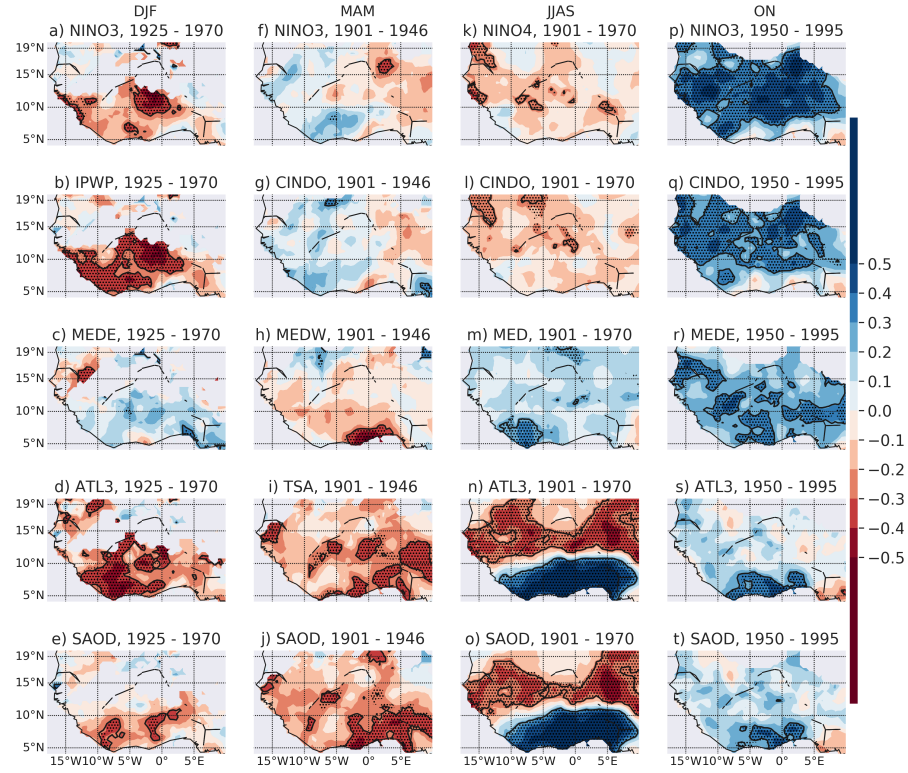


Figure A.4: Rainfall patterns associated with the main SST modes of variability for each season in periods of stationary linkages: (a-e) DJF, (f-j) MAM, (k-o) JJAS, (p-t) ON. Those patterns are obtained by correlating the interannual SST indices with rainfall anomalies in West Africa. Contour lines and dots indicate significant correlations at 95% confidence level according to an F-test with a standard and effective degree of freedom respectively.

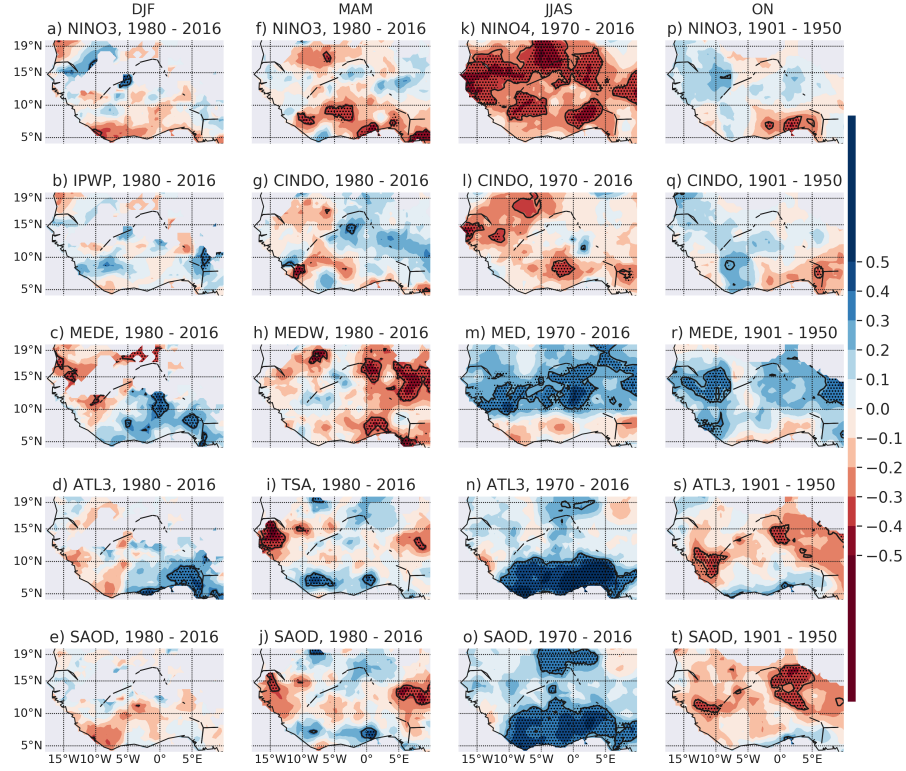


Figure A.5: Rainfall patterns associated with the main SST modes of variability for each season in periods of weakening of the most robust linkages: (a-e) DJF, (f-j) MAM, (k-o) JJAS, (p-t) ON. Those patterns are obtained by correlating the interannual SST indices with rainfall anomalies in West Africa. Contour lines and dots indicate significant correlations at 95% confidence level according to an F-test with a standard and effective degree of freedom respectively.

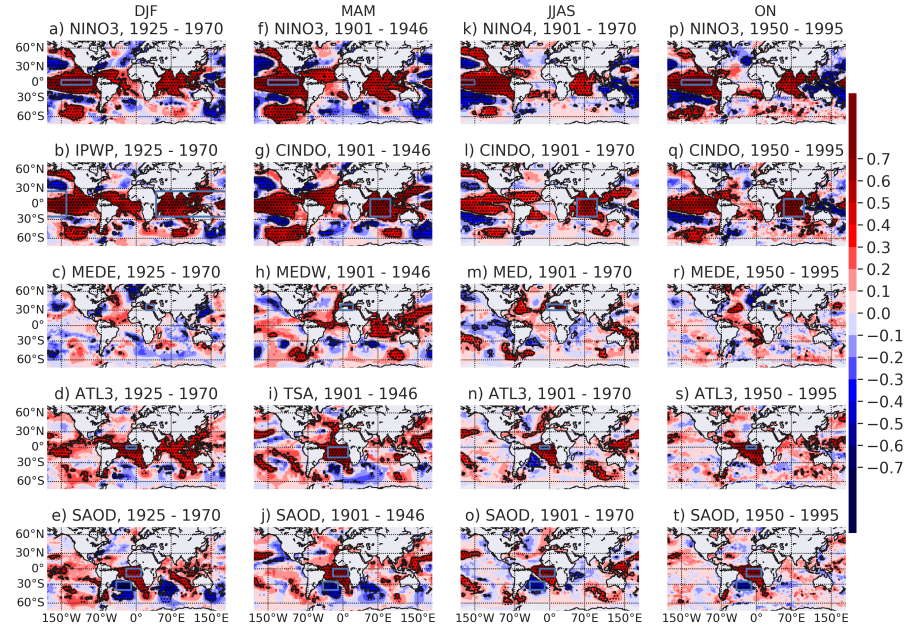


Figure A.6: SST patterns associated with the main SST modes of variability for each season in periods of stationary linkages: (a-e) DJF, (f-j) MAM, (k-o) JJAS, (p-t) ON . Those patterns are obtained by correlating the interannual SST indices with global SST anomalies of the same seasons. Contour lines and dots indicate significant correlations at 95% confidence level according to an F-test with a standard and effective degree of freedom respectively.

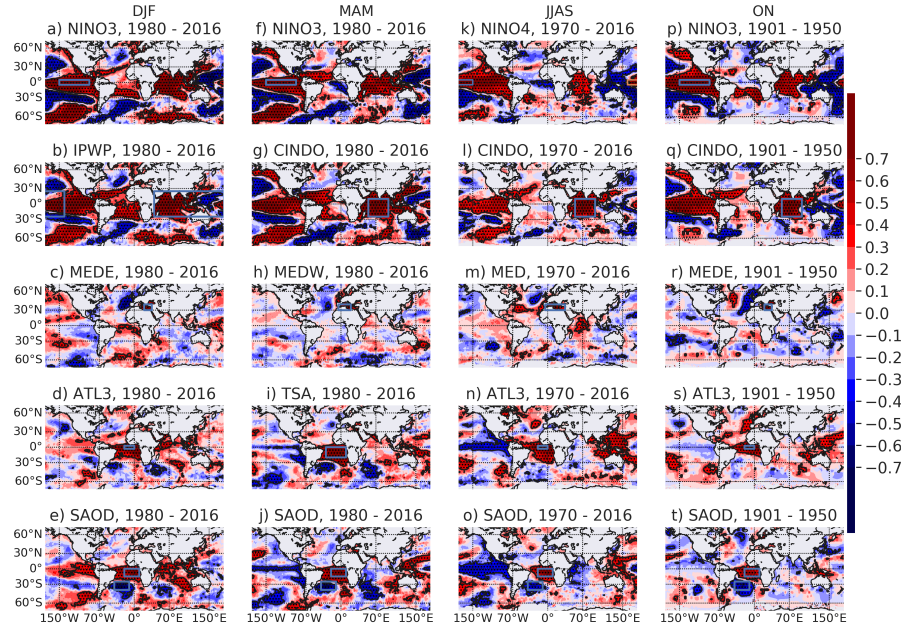


Figure A.7: SST patterns associated with the main SST modes of variability for each season in periods of weakening of the most robust linkages: (a-e) DJF, (f-j) MAM, (k-o) JJAS, (p-t) ON. Those patterns are obtained by correlating the interannual SST indices with global SST anomalies of the same season. Contour lines and dots indicate significant correlations at 95% confidence level according to an F-test with a standard and effective degree of freedom respectively.

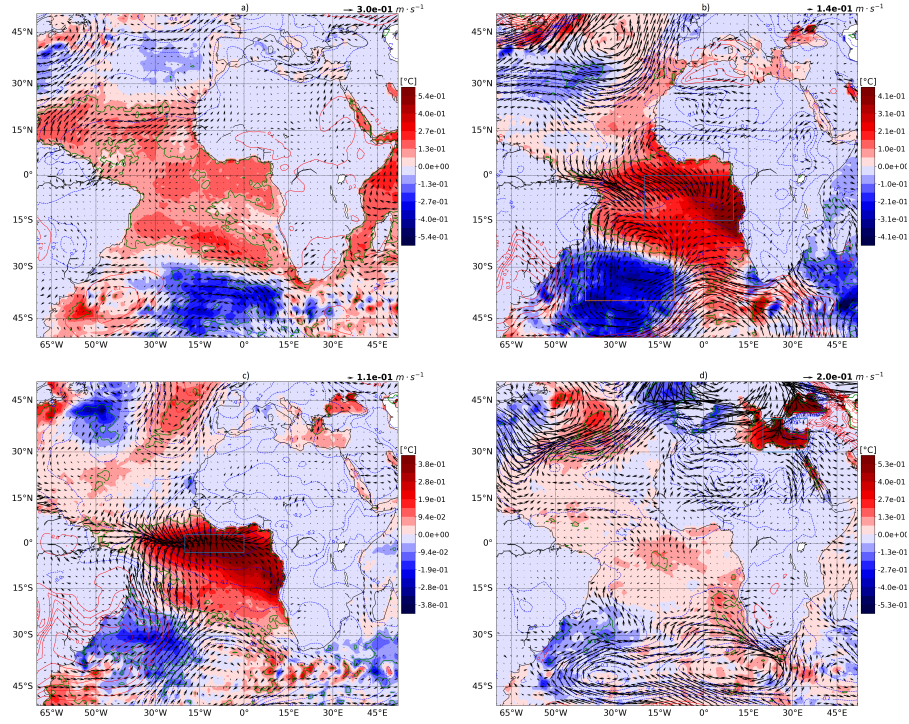


Figure A.8: Regression of sea surface temperature (colors, in °C), sea level pressure (blue and red contours, in hPa) and surface winds (arrows) anomalies onto (a) DJF IPWP, (b) MAM SAOD, (c) JJAS ATL3, (d) ON MEDE standardized indices over 1925 – 1970, 1901 – 1946, 1901 – 2010, 1950– 1995 periods respectively. Green contours denote significant SSTAs at 95% confidence level (Student-test).

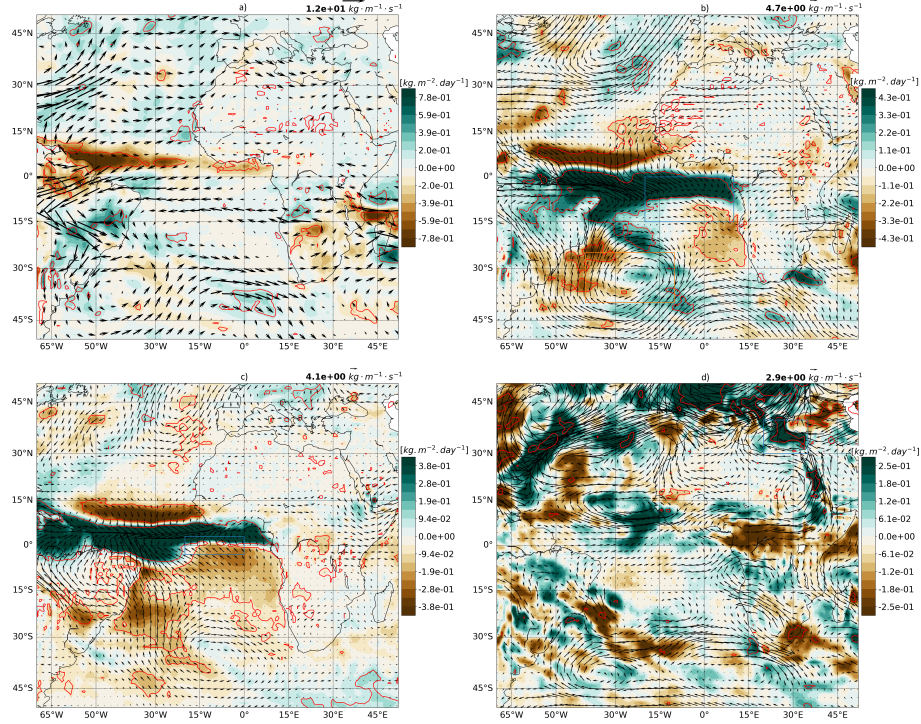


Figure A.9: Regression of vertically integrated moisture flux convergence (colors) and vertically integrated moisture flux (arrows) anomalies onto (a) DJF IPWP, (b) MAM SAOD, (c) JJAS ATL3 and (d) ON MEDE standardized SSTA indices. The regressions are evaluated for the 1925 - 1970, 1901 - 1946, 1901 - 2010, 1950 - 1995 periods respectively. Red contours denote significant VIMFC at 95% confidence level (Student-test).

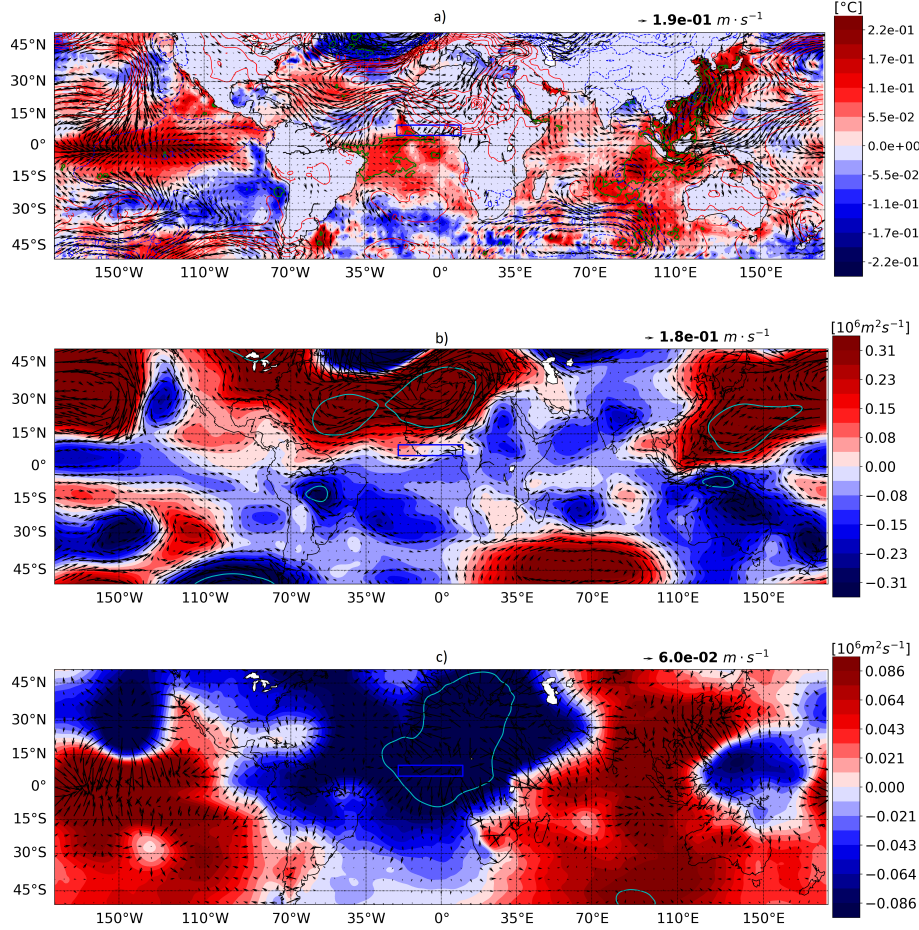


Figure A.10: DJF: regression of (a) sea surface temperature anomalies (colors, in $^{\circ}\text{C}$), sea level pressure anomalies (blue and red contours, in hPa), surface wind anomalies (arrows); (b) 850 hPa streamfunction anomalies (colors) and anomalies of non-divergent component of the wind (arrows); (c) 850 hPa anomalies of potential velocity (colors) and divergent component of the wind (arrows) onto the standardized high-pass Guinean Coast rainfall index. The rainfall index is multiplied by -1 in order to catch an anomalous warming of the tropical basins in relation with the rainfall reduction in Guinea Coast. Green contours in (a), and cyan contours in (b) and (c) indicate significant shaded values at 95% confidence level, according to a Student test. The regressions are computed for the 1925 - 1970 period. The blue box indicates the area used to compute the rainfall index of the Guinean Coast.

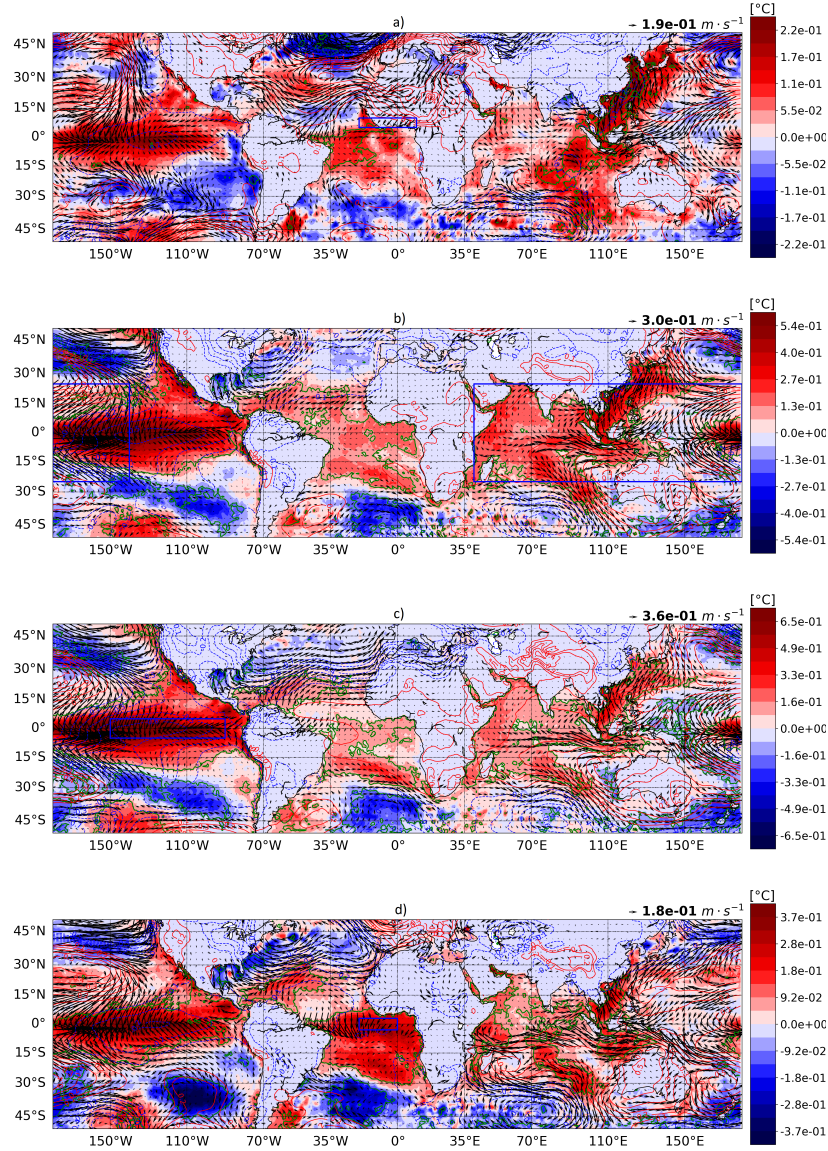


Figure A.11: DJF: regression of the sea surface temperature (colors, in $^{\circ}\text{C}$), sea level pressure (blue and red contours, in hPa) and surface wind (arrows) anomalies onto the standardized high-pass -1 x GCR (a), IPWP (b), NINO3 (c) and ATL3 (d) indices. The rainfall index in (a) is multiplied by -1 in order to catch an anomalous warming of the tropical basins in relation with the rainfall reduction in Guinea Coast. Green contours indicate significant shaded values at 95% confidence level, according to a Student test. The regressions are computed for the 1925 - 1970 period. The blue box indicates the area used to compute the index.

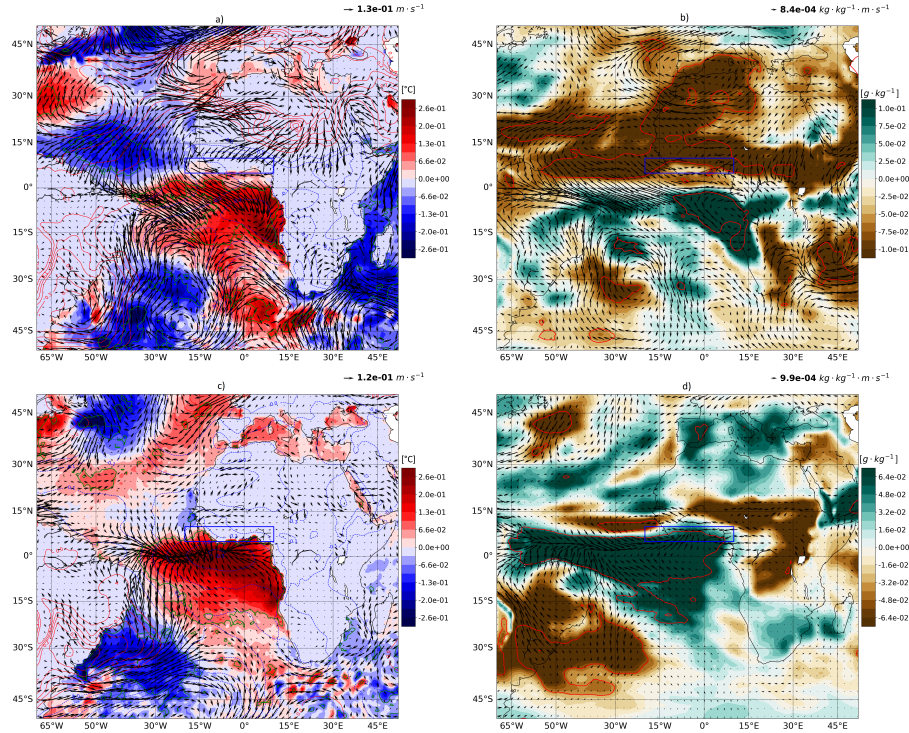


Figure A.12: Regression of: sea surface temperature (colors, in °C), sea level pressure (blue and red contours, in hPa) and surface winds (arrows) anomalies onto the Guinean Coast rainfall index during MAM (a) and JJAS (c). Regression of 850 hPa specific humidity (colors, in g · kg⁻¹) and moisture flux (vectors) anomalies onto the Guinean Coast rainfall index during MAM (b) and JJAS (d). The rainfall index used in MAM is multiplied by -1 in order to catch positive phases of the South Atlantic Ocean Dipole. Green contours in (a) and (c) indicate significant SSTAs at 95% confidence level, according to a Student test. Red contours in (b) and (d) indicate significant anomalies of specific humidity at 95% confidence level, according to a Student test. The MAM (JJAS) regressions are computed for the 1901 - 1946 (1901 - 2010) period. The blue boxes indicate the area used to compute the GCR index.

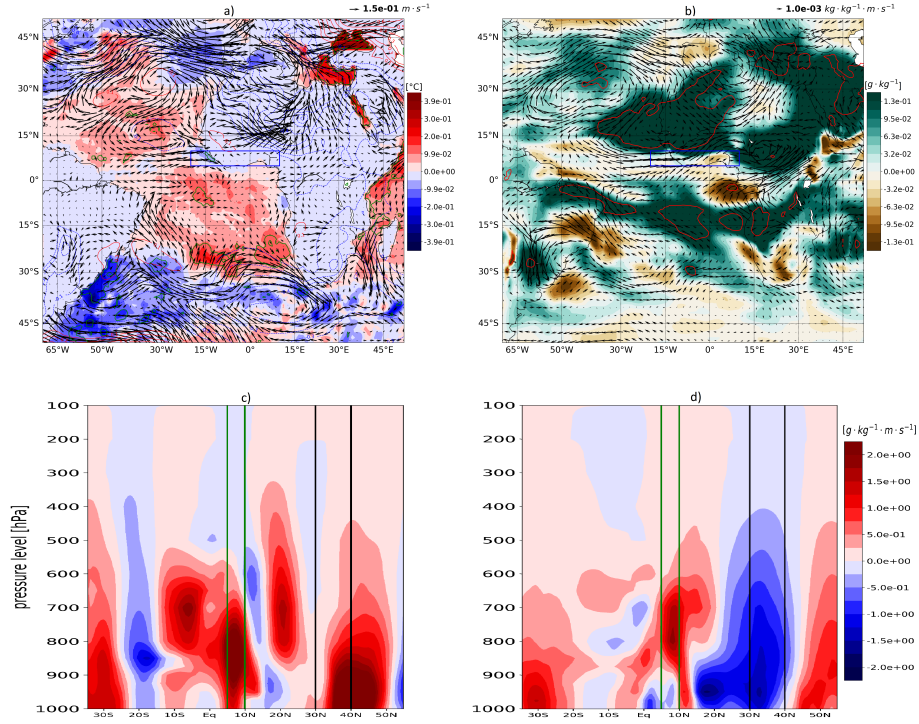


Figure A.13: ON: regression of (a) sea surface temperature (colors, in $^{\circ}\text{C}$), sea level pressure (blue and red contours, in hPa) and surface winds (arrows) anomalies and (b) 850 hPa specific humidity (colors, in $\text{g} \cdot \text{kg}^{-1}$) and moisture flux (vectors) anomalies onto the Guinean Coast rainfall index. Cross section of moisture flux anomalies (colors, in $\text{g} \cdot \text{kg}^{-1} \cdot \text{m} \cdot \text{s}^{-1}$) over the $20^{\circ}\text{W} - 0^{\circ}\text{E}$ longitudinal band: zonal component (c) and meridional component (d). The moisture flux anomalies are regressed onto the standardized ON Guinean Coast rainfall index. The green (black) vertical lines in (c) and (d) represent the latitudinal band of the Guinea Coast (Mediterranean Sea). Green contours in (a) indicate significant SSTAs at 95% confidence level, according to a Student test. Red contours in (b) indicate significant anomalies of specific humidity at 95% confidence level, according to a Student test. The regressions are computed for the 1950 - 1995 period. The blue boxes indicate the area used to compute the GCR index.

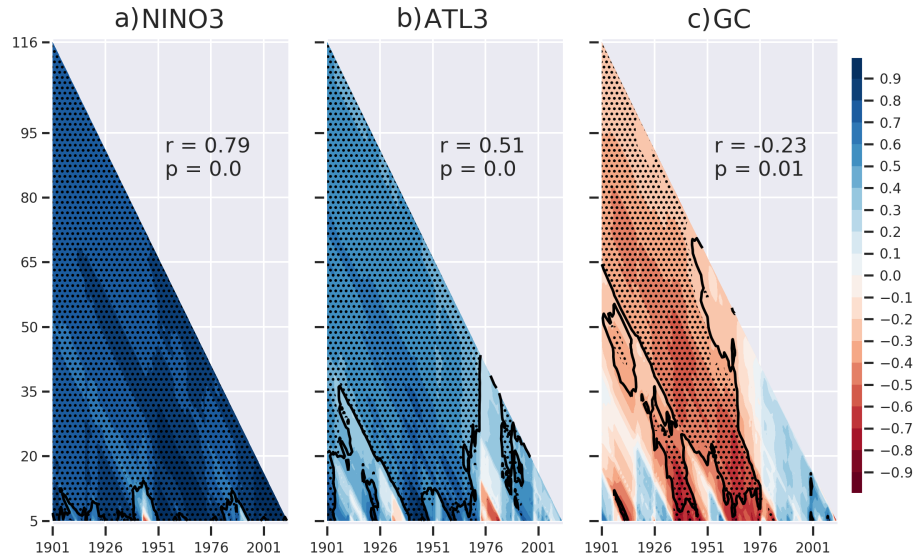


Figure A.14: Running correlation between the interannual component of the IPWP SST index and indices of NINO3, ATL3, and GCR modes of variability for different starting years (horizontal axis), different window lengths (vertical axis) during DJF. Contour lines and dots represent significant correlation values at 95% confidence level based on a F-test with a standard and effective degree of freedom respectively. Displayed r and p values correspond to the correlation value of the full series and the associated Student-test p_{value} respectively.

B

Supplement to Chapter 3

B.1 Supplementary figures

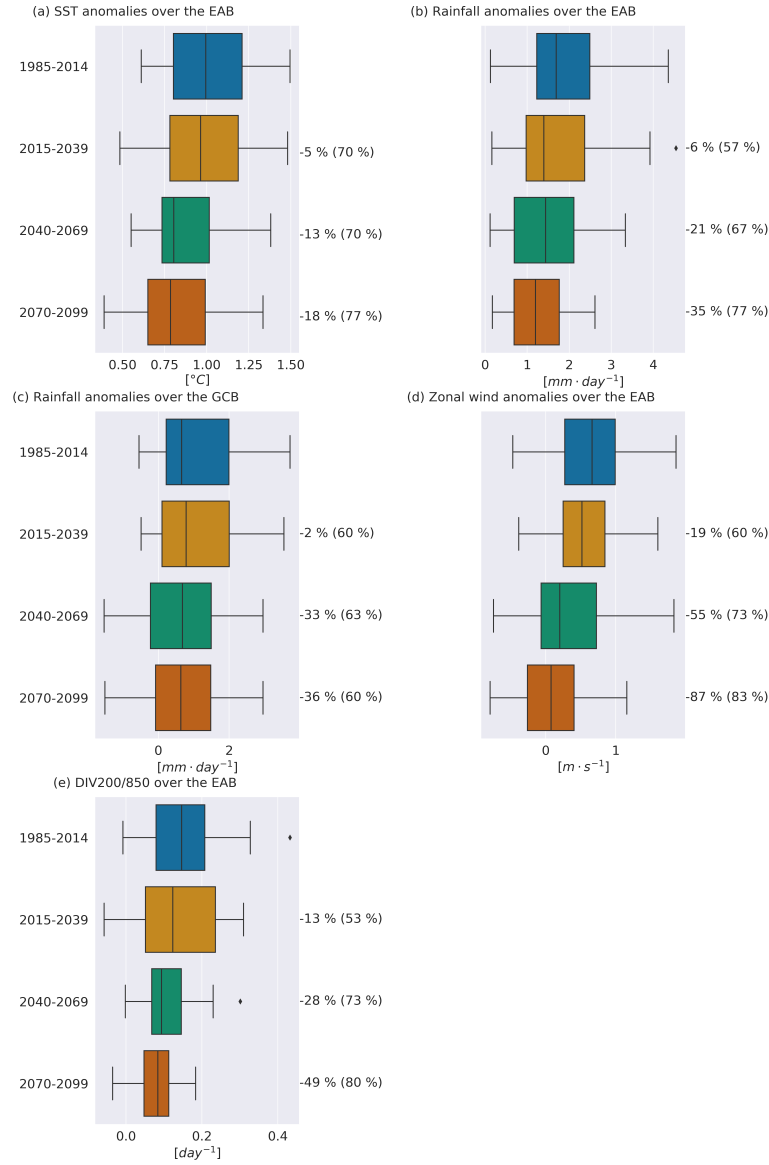


Figure B.1: Boxplots of the JAS (a) SST, (b) rainfall, (d) 850 hPa zonal wind and (e) horizontal wind divergence difference between 200 and 850 hPa (DIV200/850) anomalies averaged over the equatorial Atlantic box. Boxplot of the rainfall anomalies averaged over the Guinea Coast (c). For a given period, the anomalies for each variable are computed as composite positive minus negative AEM years for each model. Four different periods are considered: 1985–2014, 2015–2039, 2040–2069 and 2070–2099. Each boxplot represents the distribution across 30 different models.



Figure B.2: SST indices of the Atlantic Niño: JAS mean of monthly SST anomalies averaged over the Atlantic Niño area, for the 1985-2014 period (green curves). The linear (blue curves) and quadratic (orange curves) trends are superimposed on each panel. SST outputs from CMIP6 historical simulations (30 GCMs) and the ERA5 reanalysis are considered.

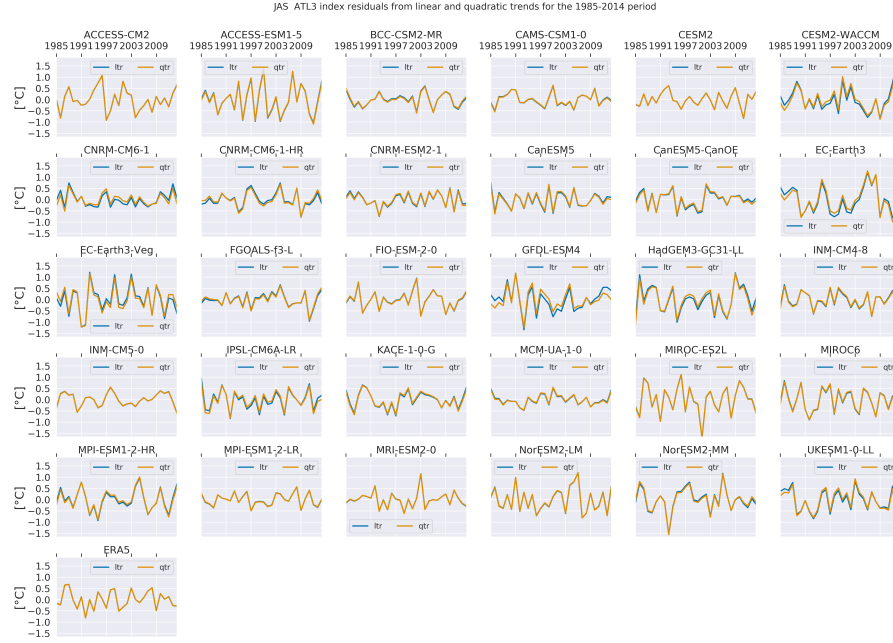


Figure B.3: Residuals of the detrended JAS ATL3 SST indices after removing the linear trend (blue curves) and the quadratic trend (orange curves). The displayed 1985-2014 time series are from 30 CMIP6 models and ERA5.

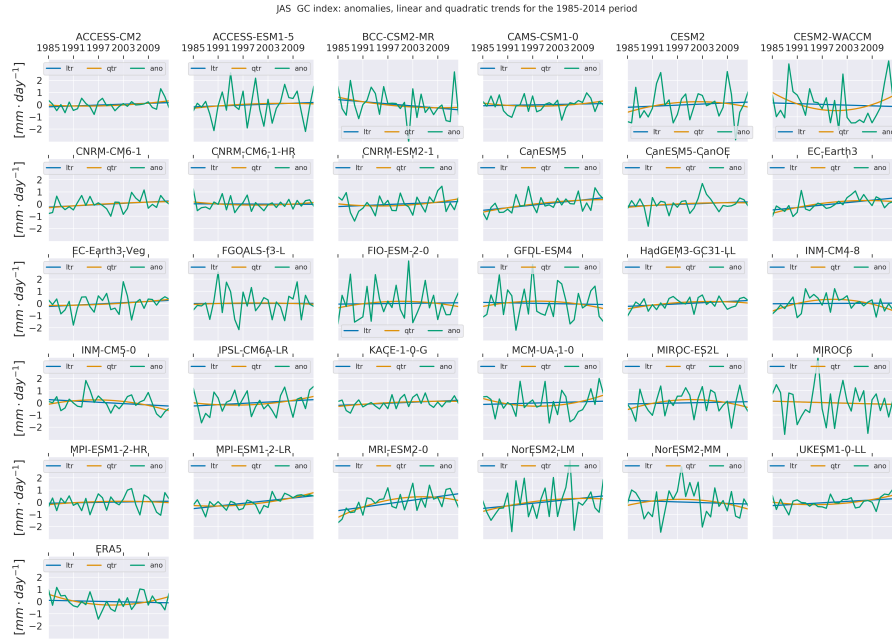


Figure B.4: JAS mean of monthly rainfall anomalies averaged over the Guinea Coast, for the 1985-2014 period (green curves). The linear (blue curves) and quadratic (orange curves) trends are superimposed on each panel. Rainfall outputs from CMIP6 historical simulations (30 GCMs) and the ERA5 reanalysis are considered.

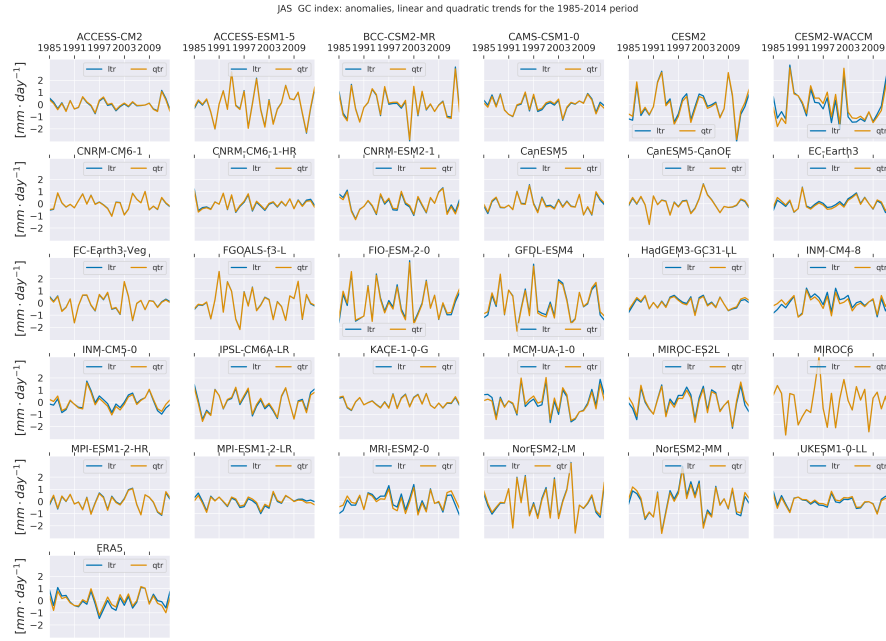


Figure B.5: Residuals of the detrended JAS Guinea Coast rainfall indices after removing the linear trend (blue curves) and the quadratic trend (orange curves). The displayed 1985-2014 time series are from 30 CMIP6 models and ERA5.

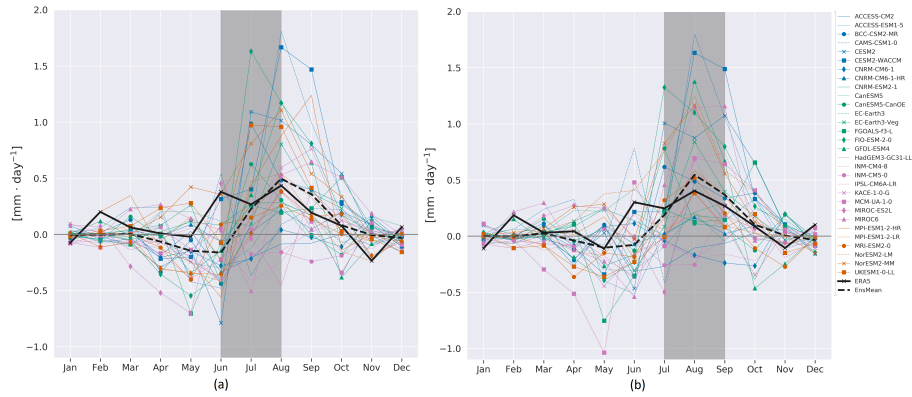
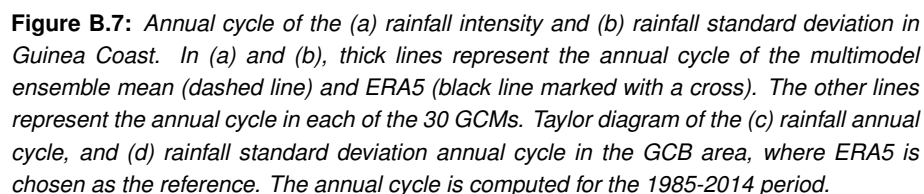


Figure B.6: Monthly rainfall anomalies of Guinea Coast regressed onto the JJA (a) and JAS (b) standardized ATL3 SST index over the 1985-2014 period. Outputs from 30 CMIP6 historical simulations and ERA5 are analyzed. Gray vertical bands indicate the SST season considered in each case.



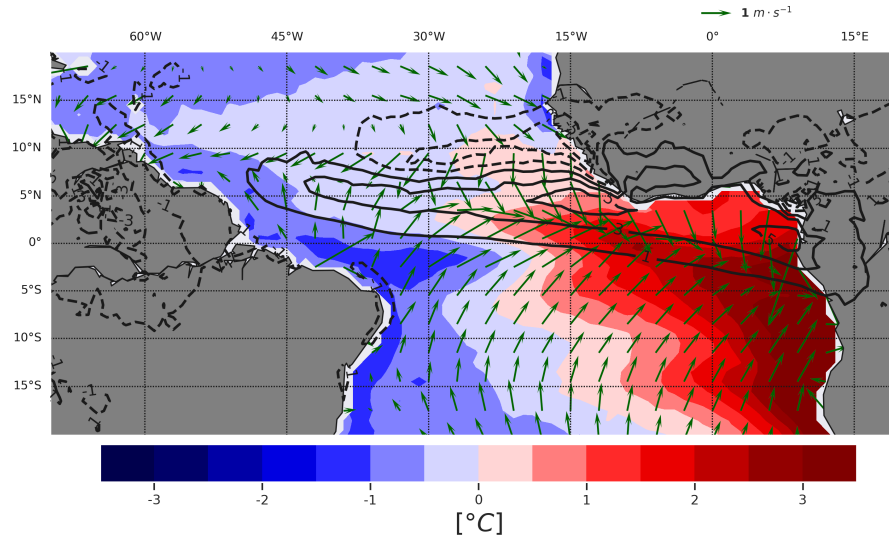


Figure B.8: Mean biases (relative to ERA5) of the ensemble mean of 22 GCMs for the JAS SST (in colors), rainfall (in contours) and 10 m horizontal wind (arrows) over 1985-2014. Among the 30 GCMs used in this study, these 22 models correspond to the ones that have the 10 m horizontal wind components for both the historical and SSP5-8.5 simulations.

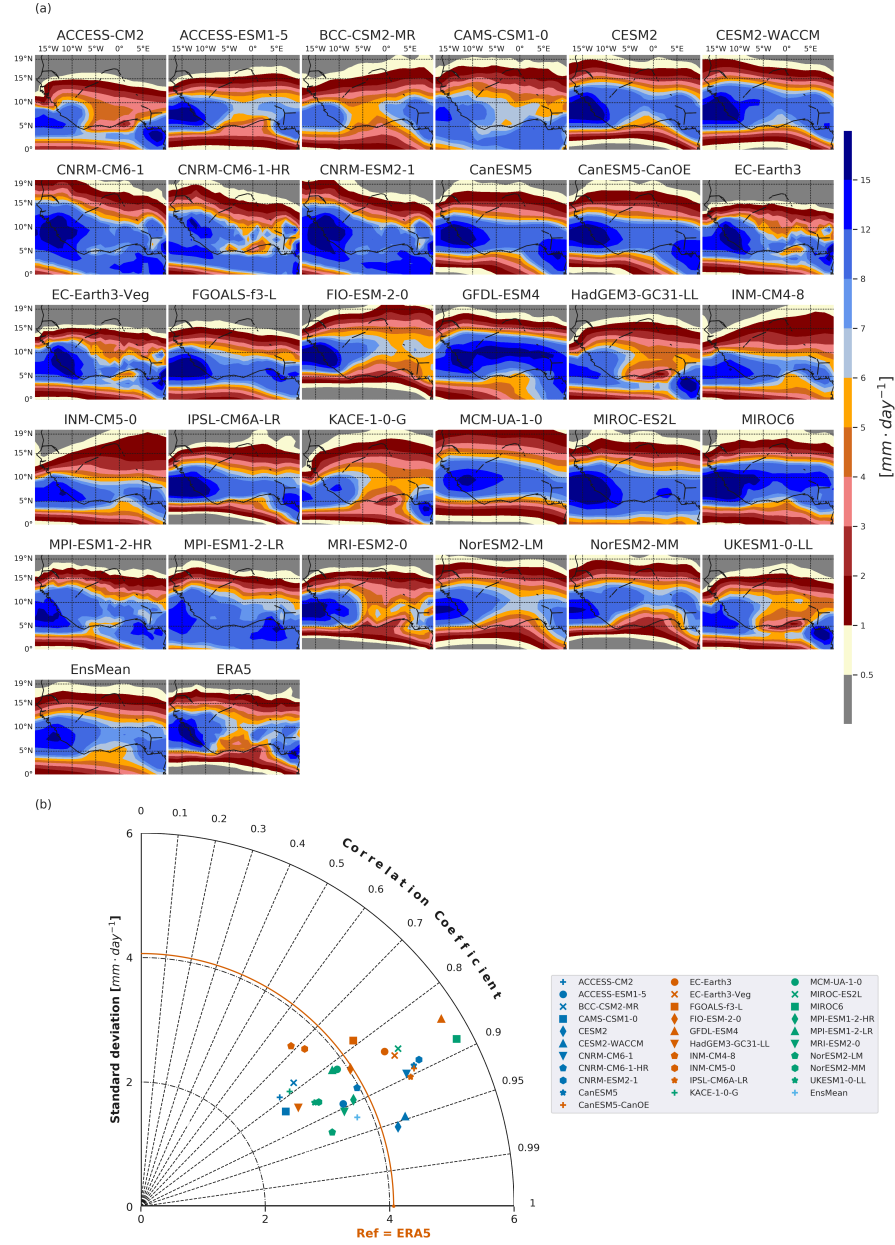


Figure B.9: (a) 1985-2014 JAS rainfall mean in 30 GCMs and in ERA5. (b) Taylor diagram of the JAS rainfall mean over West Africa (4°N - 20°N), where the mean JAS rainfall spatial distributions in the models are compared to the one in ERA5.

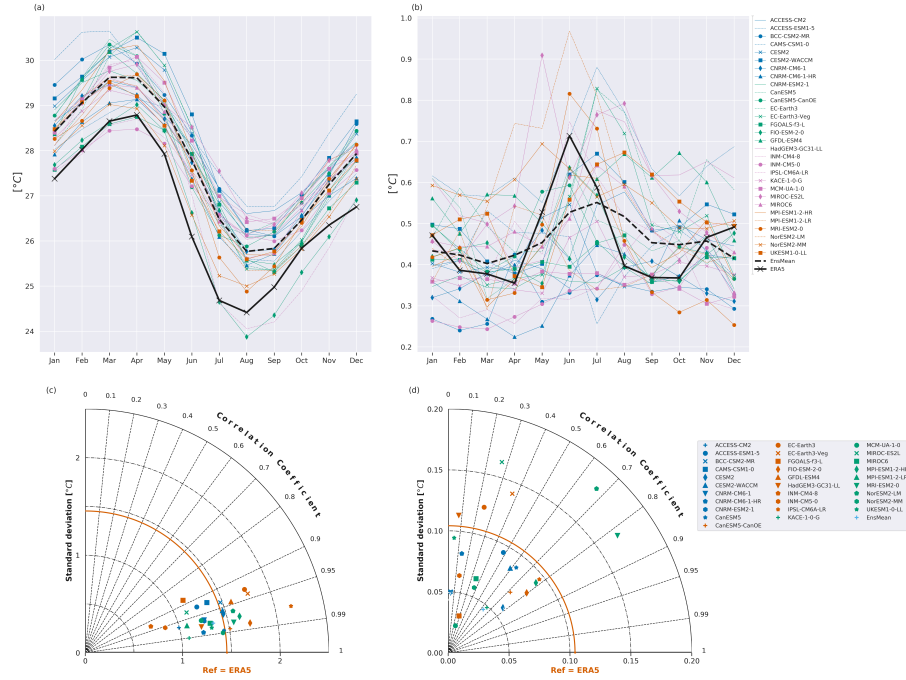


Figure B.10: Annual cycle of the (a) sea surface temperature and (b) sea surface temperature standard deviation in the ATL3 area. In (a) and (b), thick lines represent the annual cycle of the multimodel ensemble mean (dashed line) and ERA5 (black line marked with a cross). The other lines represent the annual cycle in each of the 30 GCMs. Taylor diagram of the (c) SST annual cycle and (d) SST standard deviation annual cycle in the ATL3 area, where ERA5 is chosen as the reference. The annual cycle is computed for the 1985–2014 period.

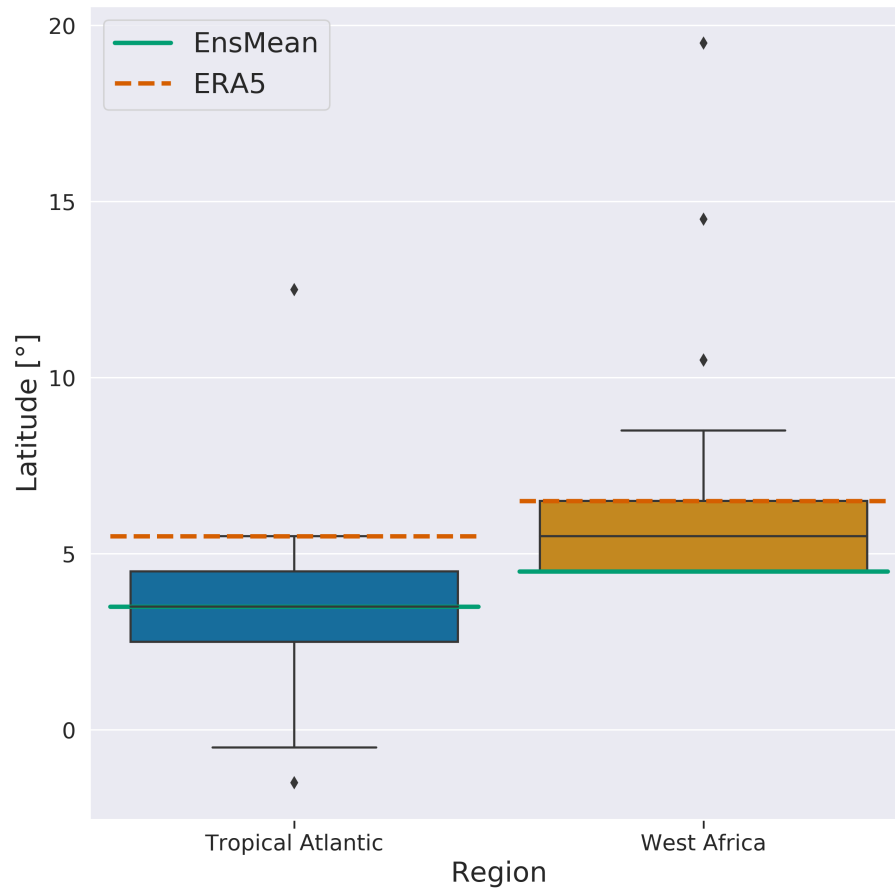


Figure B.11: Position in latitude of the maximum JAS rainfall associated with the JAS ATL3 SST index in the tropical Atlantic (zonal average over 70°W and 10°E , blue boxplot) and West Africa (zonal average between 20°W and 10°E , dark gold boxplot). The boxplots represent the distribution of the positions computed in each of the 30 models. The black line inside each box represents the median value of the models, and the outliers are represented by the diamond-shaped symbols. The other horizontal lines represent the mean positions obtained in the multi-model ensemble mean (green line) and ERA5 (dashed orange line).

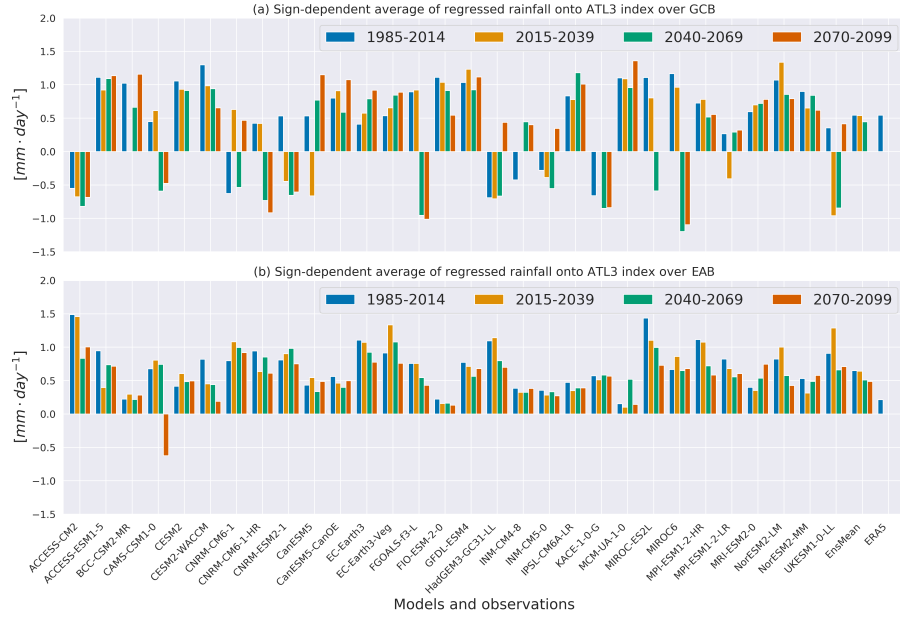


Figure B.12: Sign-dependent average of the JAS rainfall regressed coefficients onto the standardized JAS ATL3 SST index over the Guinea Coast box (a) and the equatorial Atlantic box (b) for the 30 GCMs over the present-day, near-term, mid-term and long-term periods. The sign-dependent average value in ERA5 is computed for the 1985-2014 period.

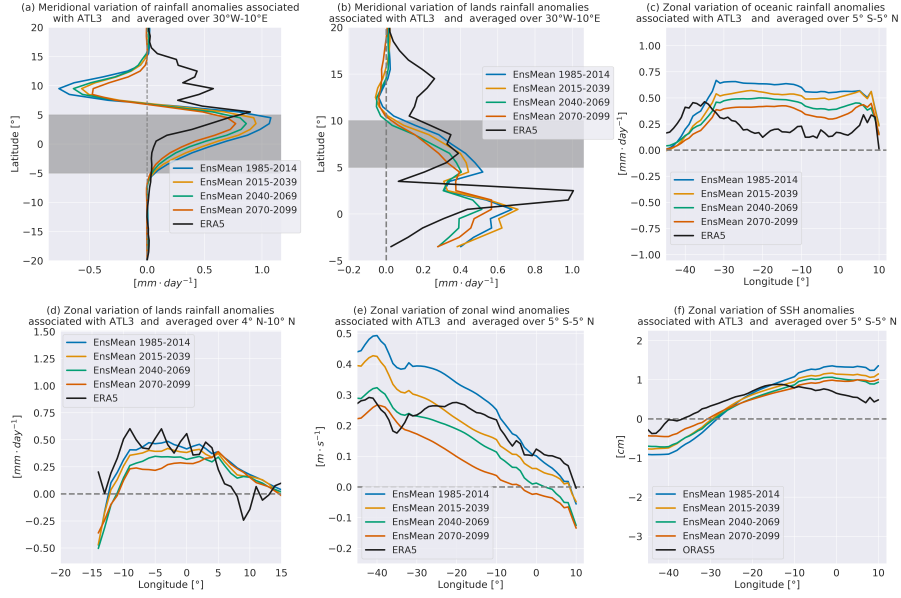


Figure B.13: Rainfall anomalies associated with the standardized JAS ATL3 SST index and averaged over 30°W and 10°E for oceanic areas (a) and land areas (b). Gray bands in (a) and (b) represent the equatorial Atlantic and Guinea Coast regions, respectively. Zonal variation of the JAS rainfall anomalies associated with the standardized JAS ATL3 SST index and averaged over the equatorial Atlantic (c) and Guinea Coast regions (d). Zonal variation of the JAS 850 hPa zonal wind (e) and sea surface height anomalies (f) associated with the standardized JAS ATL3 SST index and averaged over the equatorial Atlantic. In each panel, the solid black line represents ERA5 or ORAS5 for 1985-2014, and the other solid lines represent the ensemble mean of the 30 GCMs for the present and future periods.

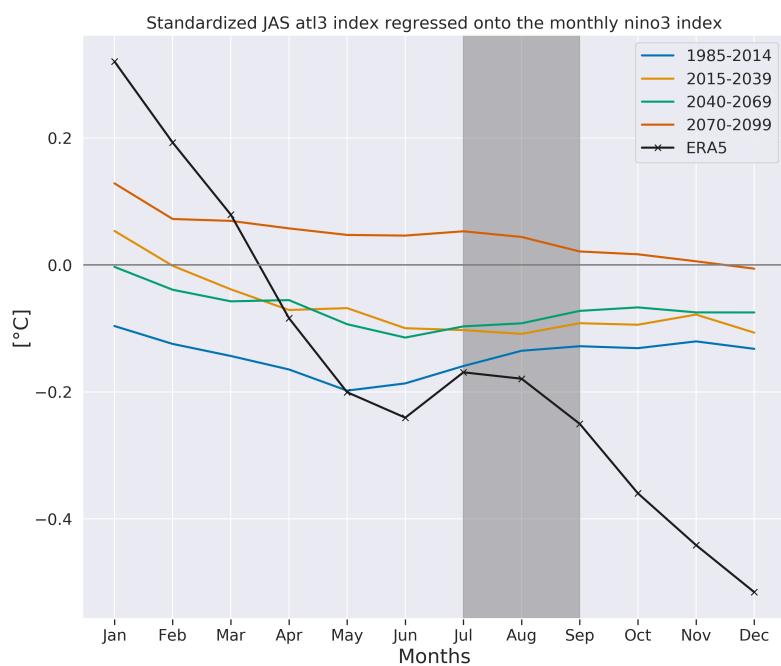


Figure B.14: Monthly stratified Niño3 index regressed onto the standardized JAS ATL3 SST index for different periods. The 1985-2014 period is considered for ERA5 (black curve). The other curves correspond to the ensemble mean response of 30 CMIP6 models over four different periods: 1985-2014, 2015-2039, 2040-2069 and 2070-2099.

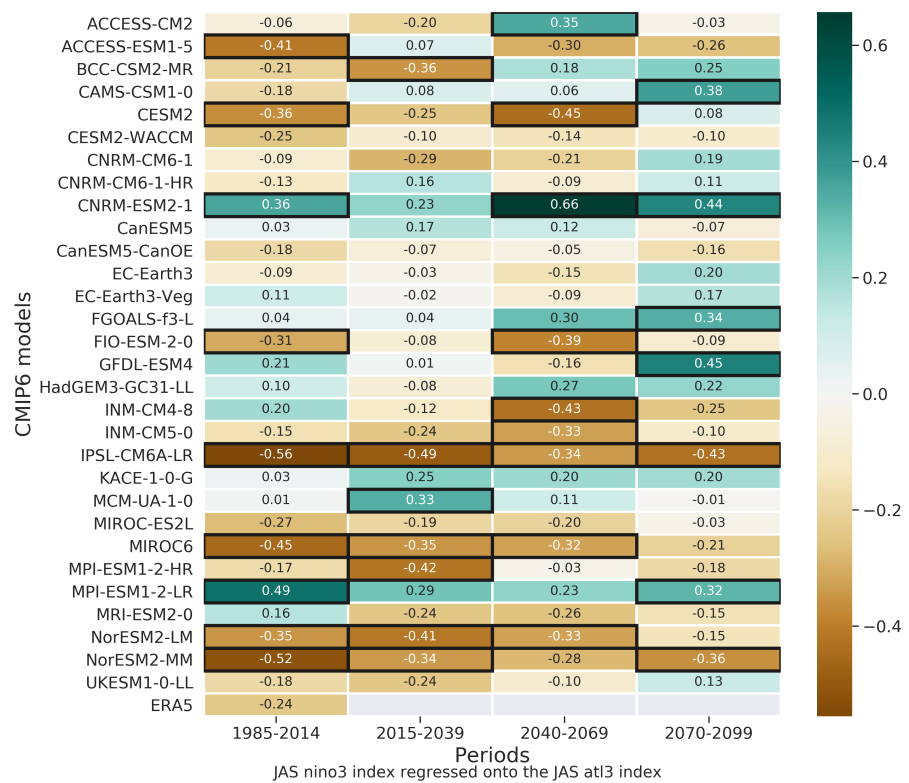


Figure B.15: JAS Niño3 index correlation with the JAS ATL3 SST index over four different periods. 30 CMIP6 models and the reanalysis ERA5 are analyzed. Significant regression coefficients at 90% confidence level (Student test) are highlighted with a black box.

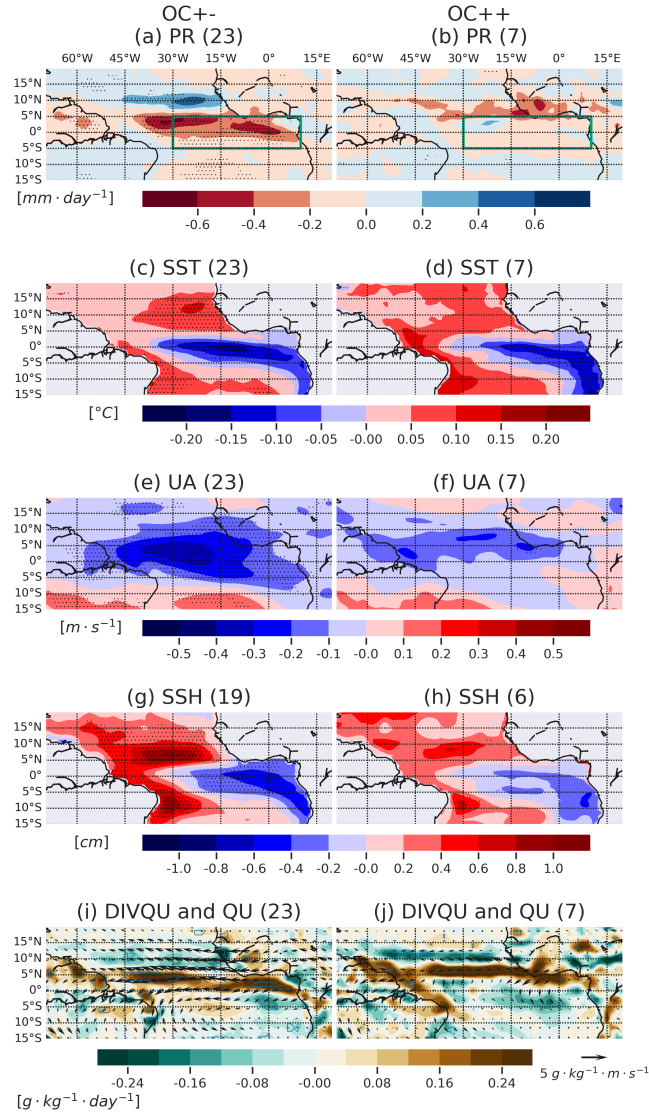


Figure B.16: Long-term changes of the JAS rainfall (a-b), SST (c-d), 850 hPa zonal wind (e-f), sea surface height (g-h), moisture flux (vectors) and moisture flux divergence (in colors) (i-j) regression patterns associated with the standardized JAS ATL3 SST index, relative to the present-day climate (2070-2099 minus 1985-2014). Stippling in (a)-(h) and contours in (i)-(j) indicate areas where the mean change (in colors) is significant at 95 % level according to a two-sided Welch t-test and where at least two thirds of the models agree on the sign of the change. The number of models in each group is indicated in parentheses.

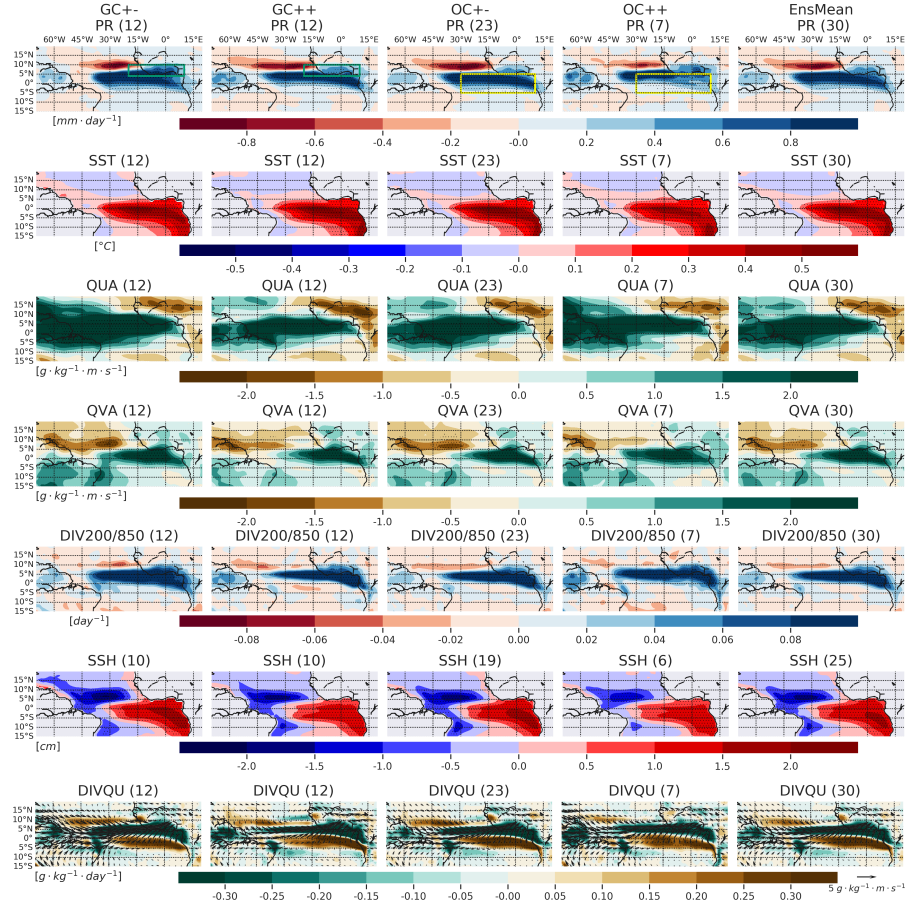


Figure B.17: 1985-2014 regression maps of the JAS rainfall, SST, 850 hPa zonal and meridional moisture flux, DIV200/850, SSH, 850 hPa moisture flux and moisture flux divergence anomalies onto the standardized JAS ATL3 SST index, for the GC+-, GC++, OC+-, OC++ and the 30 GCMs EnsMean groups. Stippling indicates areas where the regression coefficients are significant at 95 % confidence level for at least 50 % of the models in each group, and where more than 80 % of the models agree on the sign of the regression coefficient. The number of models in each group is indicated in parentheses.

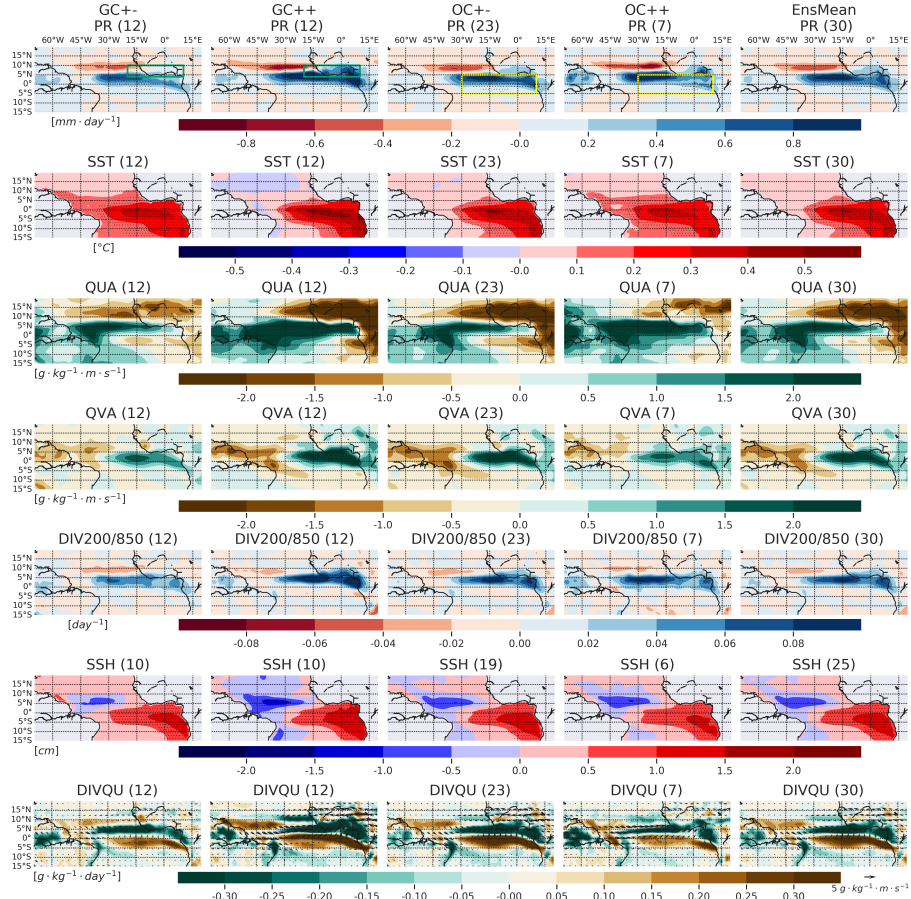


Figure B.18: 2070-2099 regression maps of the JAS rainfall, SST, 850 hPa zonal and meridional moisture flux, DIV200/850, SSH, 850 hPa moisture flux and moisture flux divergence anomalies onto the standardized JAS ATL3 SST index, for the GC+-, GC++, OC+-, OC++ and the 30 GCMs EnsMean groups. Stippling indicates areas where the regression coefficients are significant at 95 % confidence level for at least 50 % of the models in each group, and where more than 80 % of the models agree on the sign of the regression coefficient. The number of models in each group is indicated in parentheses.

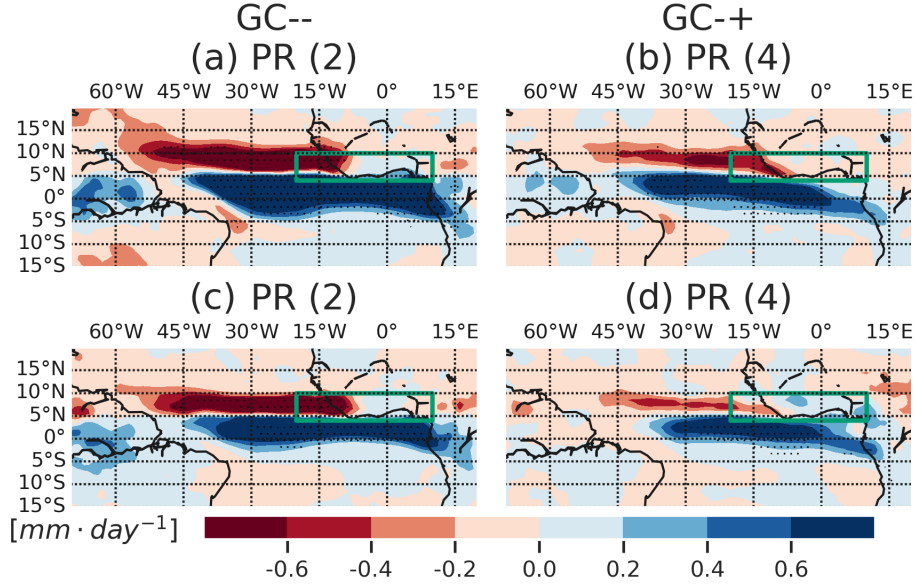


Figure B.19: 1985-2014 (a)-(b) and 2070-2099 (c)-(d) regression maps of the JAS rainfall anomalies onto the standardized JAS ATL3 SST index for the GC- (left column) and GC+ (right column) groups. Stippling indicates areas where the regression coefficients are significant at 95 % confidence level for at least 50 % of the models in each group, and where more than 80 % of the models agree on the sign of the regression coefficient. The number of models in each group is indicated in parentheses.

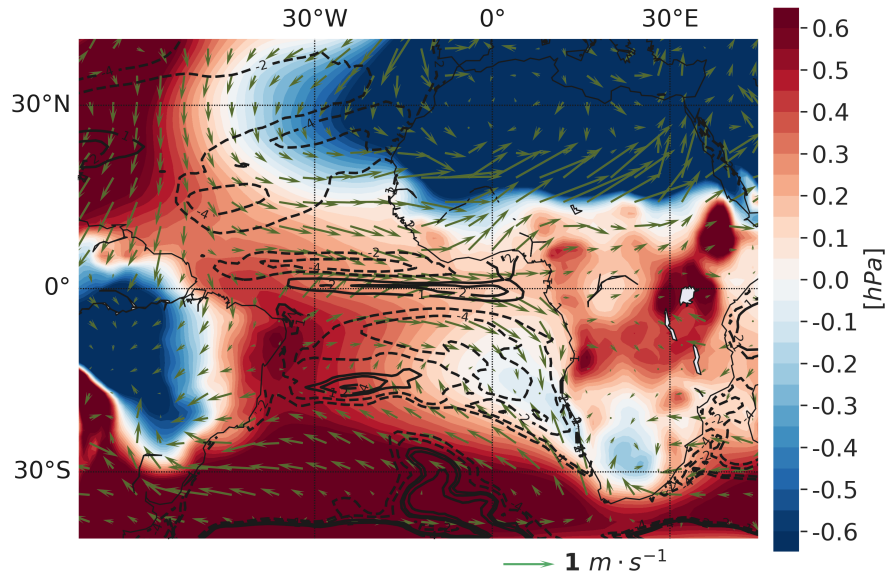


Figure B.20: Long-term mean state change (2070-2099 minus 1985-2014) of the sea level pressure (in colors), 10 m horizontal wind (arrows) and mixed layer depth (in contours in m). These changes are computed for the JAS season and averaged over 22 GCMs (Table 3.1).

B.2 Supplementary tables

Table B.1: Multi-model averages of the JAS Guinea Coast rainfall, EAB rainfall, EAB SST, EAB 850 hPa zonal wind, EAB DIV200/850 and EAB SSH anomalies related to the JAS EAM in the GC++, GC+- and Ensmean groups.

		GC++	GC+-	EnsMean	Unit
GCR	1985-2014	0.40	0.56	0.37	$\text{mm} \cdot \text{day}^{-1}$
	2070-2099	0.49	0.10	0.24	$\text{mm} \cdot \text{day}^{-1}$
	%of change	22.63	-83.01	-34.81	
OCR	1985-2014	0.55	0.67	0.62	$\text{mm} \cdot \text{day}^{-1}$
	2070-2099	0.42	0.33	0.40	$\text{mm} \cdot \text{day}^{-1}$
	%of change	-22.93	-50.74	-35.95	
SST	1985-2014	0.33	0.33	0.33	$^{\circ}\text{C}$
	2070-2099	0.28	0.26	0.27	$^{\circ}\text{C}$
	%of change	-15.22	-20.68	-18.50	
UA	1985-2014	0.22	0.31	0.24	$\text{m} \cdot \text{s}^{-1}$
	2070-2099	0.12	0.03	0.05	$\text{m} \cdot \text{s}^{-1}$
	%of change	-44.18	-89.74	-79.44	
DIV200/850	1985-2014	0.04	0.06	0.05	day^{-1}
	2070-2099	0.03	0.02	0.03	day^{-1}
	%of change	-32.88	-61.60	-49.81	
ZOS	1985-2014	0.75	1.02	0.87	cm
	2070-2099	0.68	0.73	0.66	cm
	%of change	-9.40	-28.46	-23.42	

Table B.2: *Models in the different categories for the 2015-2039 period.*

GC++	GC+-	GC-+	GC— —	OC+	OC-
CAMS-CSM1-0	ACCESS-ESM1-5	CNRM-CM6-1	ACCESS-CM2	BCC-CSM2-MR	ACCESS-CM2
CanESM5-CanOE	BCC-CSM2-MR	INM-CM4-8	HadGEM3-GC31-LL	CAMS-CSM1-0	ACCESS-ESM1-5
EC-Earth3	CESM2	KACE-1-0-G	INM-CM5-0	CESM2	CESM2-WACCM
EC-Earth3-Veg	CESM2-WACCM			CNRM-CM6-1	CNRM-CM6-1-HR
FGOALS-f3-L	CNRM-CM6-1-HR			CNRM-ESM2-1	CanESM5-CanOE
GFDL-ESM4	CNRM-ESM2-1			CanESM5	EC-Earth3
MPI-ESM1-2-HR	CanESM5			EC-Earth3-Veg	FGOALS-f3-L
MRI-ESM2-0	FIO-ESM-2-0			HadGEM3-GC31-LL	FIO-ESM-2-0
NorESM2-LM	IPSL-CM6A-LR			MIROC6	GFDL-ESM4
	MCM-UA-1-0			NorESM2-LM	INM-CM4-8
	MIROC-ES2L			UKESM1-0-LL	INM-CM5-0
	MIROC6				IPSL-CM6A-LR
	MPI-ESM1-2-LR				KACE-1-0-G
	NorESM2-MM				MCM-UA-1-0
	UKESM1-0-LL				MIROC-ES2L
					MPI-ESM1-2-HR
					MPI-ESM1-2-LR
					MRI-ESM2-0
					NorESM2-MM

Table B.3: *Models in the different categories for the 2040-2069 period.*

GC++	GC+-	GC-+	GC— —	OC+	OC-
CanESM5	ACCESS-ESM1-5	CNRM-CM6-1	ACCESS-CM2	CAMS-CSM1-0	ACCESS-CM2
EC-Earth3	BCC-CSM2-MR	HadGEM3-GC31-LL	INM-CM5-0	CESM2	ACCESS-ESM1-5
EC-Earth3-Veg	CAMS-CSM1-0	INM-CM4-8	KACE-1-0-G	CNRM-CM6-1	BCC-CSM2-MR
IPSL-CM6A-LR	CESM2			CNRM-ESM2-1	CESM2-WACCM
MPI-ESM1-2-LR	CESM2-WACCM			EC-Earth3-Veg	CNRM-CM6-1-HR
MRI-ESM2-0	CNRM-CM6-1-HR			KACE-1-0-G	CanESM5
	CNRM-ESM2-1			MCM-UA-1-0	CanESM5-CanOE
	CanESM5-CanOE			MRI-ESM2-0	EC-Earth3
	FGOALS-f3-L				FGOALS-f3-L
	FIO-ESM-2-0				FIO-ESM-2-0
	GFDL-ESM4				GFDL-ESM4
	MCM-UA-1-0				HadGEM3-GC31-LL
	MIROC-ES2L				INM-CM4-8
	MIROC6				INM-CM5-0
	MPI-ESM1-2-HR				IPSL-CM6A-LR
	NorESM2-LM				MIROC-ES2L
	NorESM2-MM				MIROC6
	UKESM1-0-LL				MPI-ESM1-2-HR
					MPI-ESM1-2-LR
					NorESM2-LM
					NorESM2-MM
					UKESM1-0-LL

Table B.4: Models in the different categories for the 2070-2099 period.

GC++	GC+-	GC-+	GC— —	OC+	OC-
ACCESS-ESM1-5	CAMS-CSM1-0	CNRM-CM6-1	ACCESS-CM2	BCC-CSM2-MR	ACCESS-CM2
BCC-CSM2-MR	CESM2	HadGEM3-GC31-LL	KACE-1-0-G	CESM2	ACCESS-ESM1-5
CanESM5	CESM2-WACCM	INM-CM4-8		CNRM-CM6-1	CAMS-CSM1-0
CanESM5-CanOE	CNRM-CM6-1-HR	INM-CM5-0		CanESM5	CESM2-WACCM
EC-Earth3	CNRM-ESM2-1			MIROC6	CNRM-CM6-1-HR
EC-Earth3-Veg	FGOALS-f3-L			MRI-ESM2-0	CNRM-ESM2-1
GFDL-ESM4	FIO-ESM-2-0			NorESM2-MM	CanESM5-CanOE
IPSL-CM6A-LR	MIROC-ES2L				EC-Earth3
MCM-UA-1-0	MIROC6				EC-Earth3-Veg
MPI-ESM1-2-LR	MPI-ESM1-2-HR				FGOALS-f3-L
MRI-ESM2-0	NorESM2-LM				FIO-ESM-2-0
UKESM1-0-LL	NorESM2-MM				GFDL-ESM4
					HadGEM3-GC31-LL
					INM-CM4-8
					INM-CM5-0
					IPSL-CM6A-LR
					KACE-1-0-G
					MCM-UA-1-0
					MIROC-ES2L
					MPI-ESM1-2-HR
					MPI-ESM1-2-LR
					NorESM2-LM
					UKESM1-0-LL

C

Supplement to Chapter 4

C.1 Additional figures

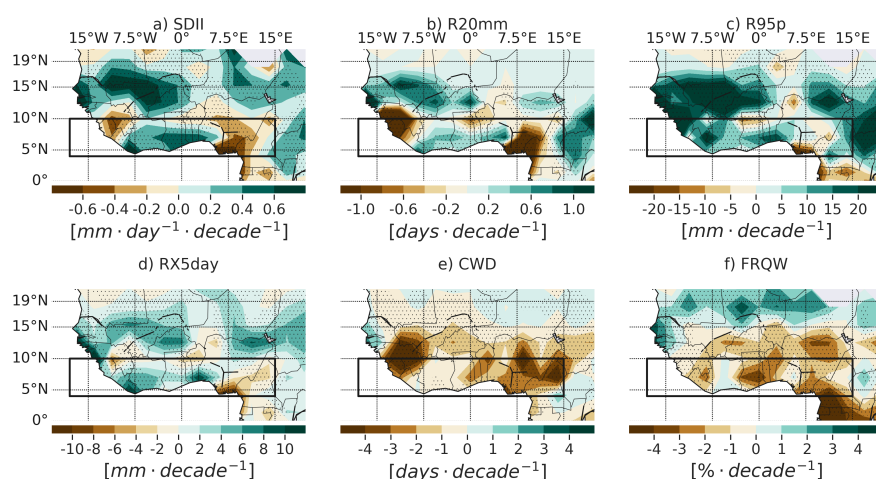


Figure C.1: *EnsMedian* distribution of the trends in JAS extreme rainfall indices over the 1995-2014 period from six observed rainfall datasets. The black rectangles indicate the Guinea Coast region. Stippling indicates grid points where two-third of the observations agree on the sign of the trend.

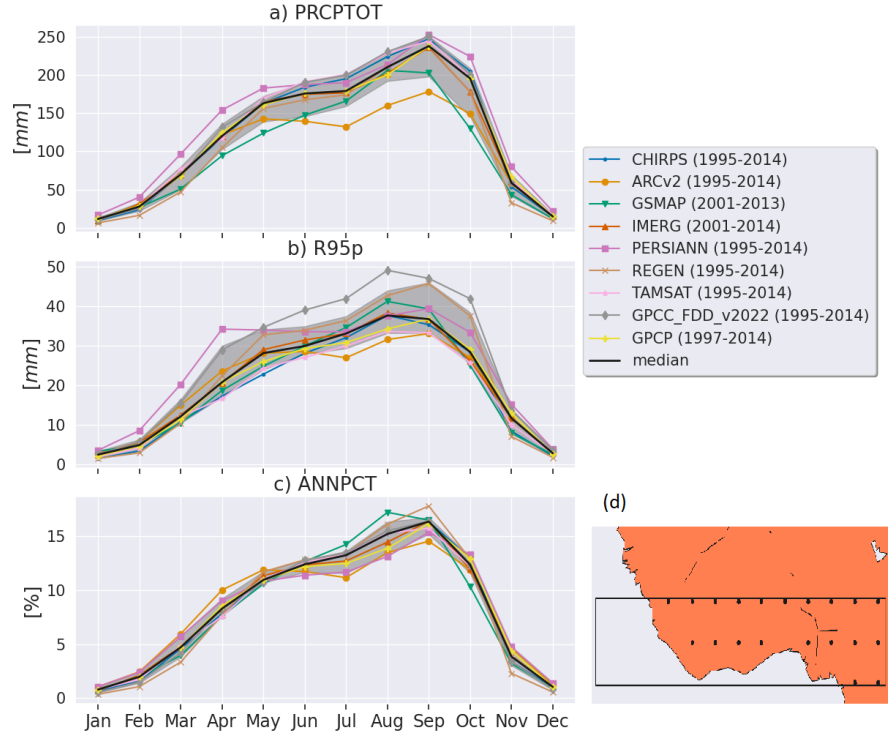


Figure C.2: Annual cycle of the (a) total wet day precipitation (PRCPTOT), (b) very wet days precipitation (sum of the daily rainfall over days when the rainfall exceeds the 95th percentile), and (c) contribution of the total monthly rainfall to the total annual rainfall, for nine different observational datasets. The period considered for each dataset is displayed in the legend. The ensemble median of the nine observations is indicated by the black curve. The grey shading shows the 10th to 90th percentile range of the observations. (d) West-Africa region, showing 20 grid points selected to compute the area-average of the three indices over Guinea Coast region (black rectangle).

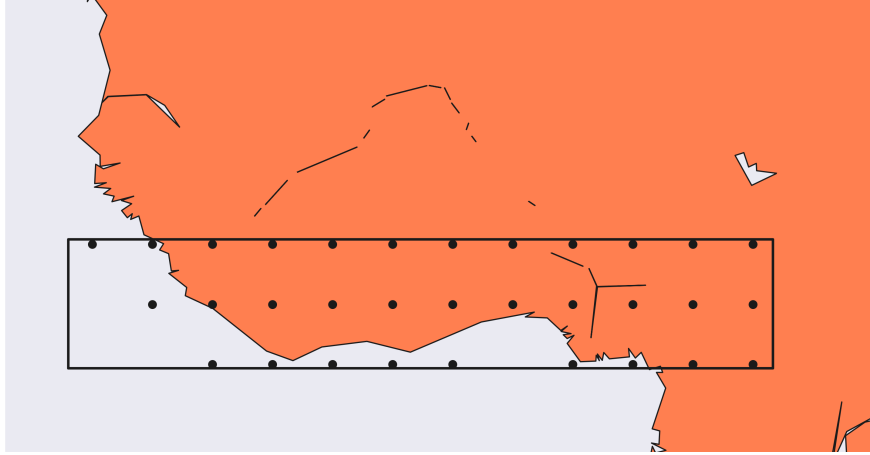


Figure C.3: Map of West Africa showing the Guinea Coast region (the black rectangle) and the 32 grid points used when considering an area-average over Guinea Coast.

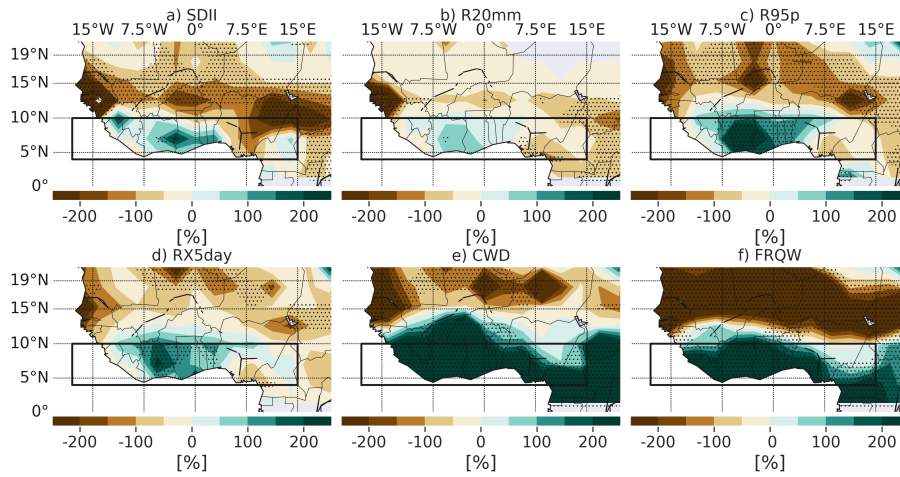


Figure C.4: Spatial distribution of the EnsMedian bias percentage relative to the interquartile range of the six observations for the JAS climatology of the extreme rainfall indices over the 1995-2014 period. In total, there are six observations $\times$ 24 models = 144 different biases. The stippling indicates regions where two-thirds of the biases agree on the sign of the EnsMedian of the 144 biases. The black rectangles indicate the Guinea Coast region.

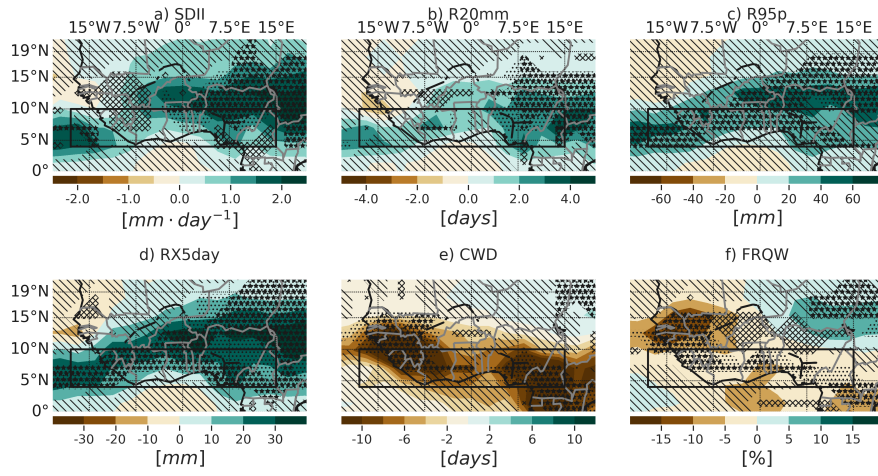


Figure C.5: Projected multi-model ensemble long-term median change in the JAS rainfall extreme indices over West Africa, relative to the present-day period (2080-2099 minus 1995-2014). Stippling indicates regions where the change robustly emerges from internal variability (at least 66% of the models show a change greater than the IAV and at least 80% of the models agree on the sign of change); hatching ($\backslash$) indicates regions where the change is unrobust (fewer than 66% of the models show change greater than the IAV); crossed lines (X) indicate conflicting signals where at least 66% of the models show change greater than the IAV, with less than 80% agreement on the sign of the change. Stars indicate an another approach used in Monerie et al. [2017], showing grid points where the forced response is stronger than the internal variability.

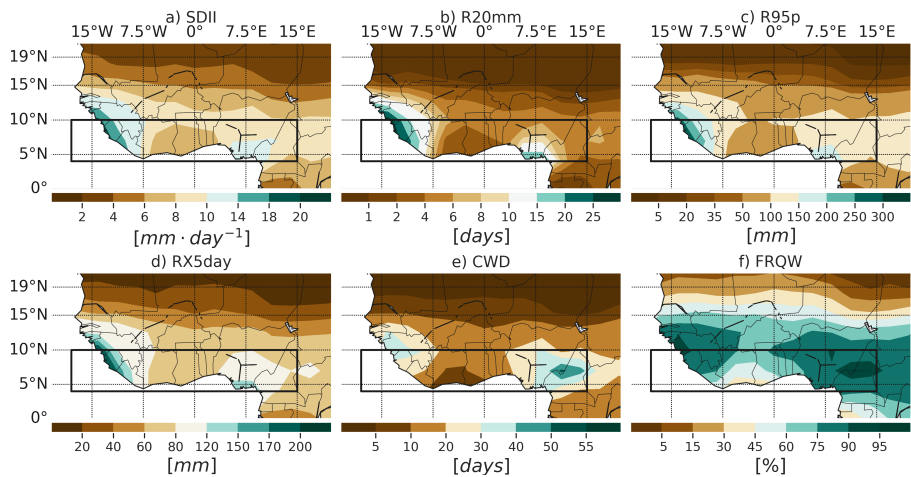


Figure C.6: Spatial distribution of the *EnsMedian* JAS extreme rainfall indices over the 1995-2014 period from 6 different rainfall dataset. The black rectangles indicate the Guinea Coast region.

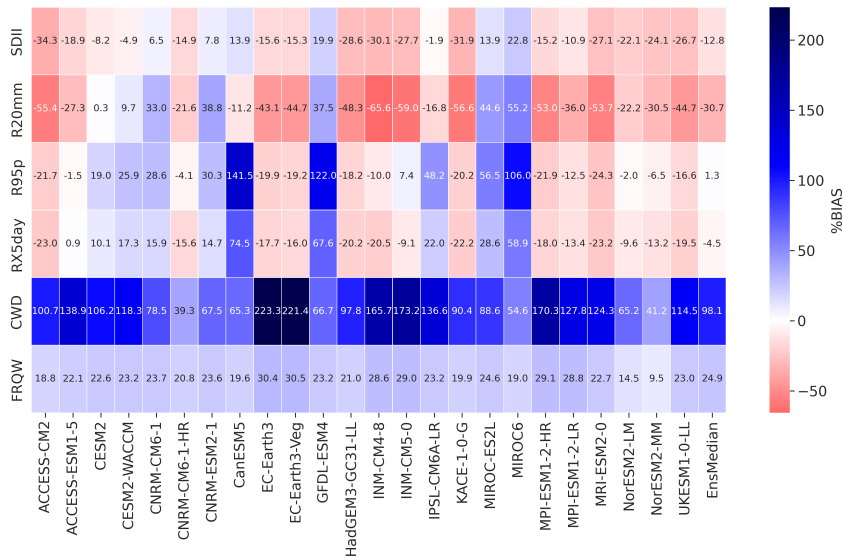


Figure C.7: Performance metrics of the individual GCM in representing six different extreme rainfall indices: *EnsMedian* of %BIAS across six different observations.

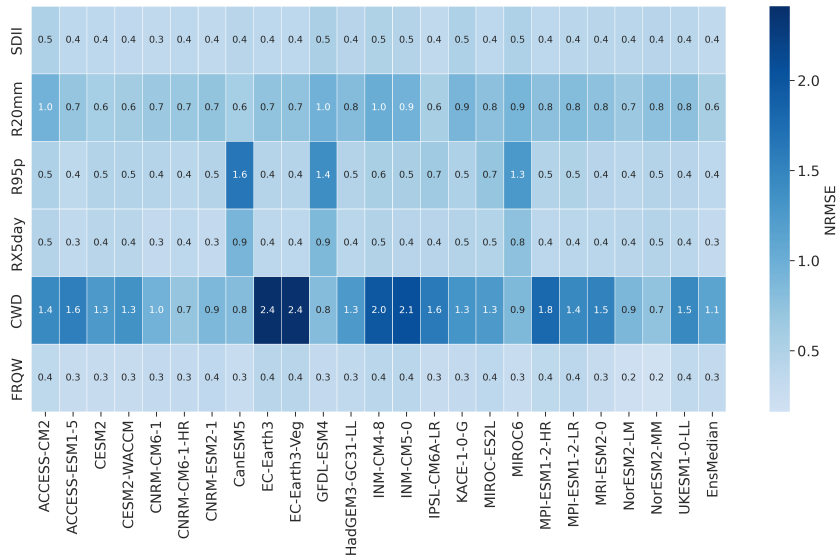


Figure C.8: Performance metrics of the individual GCM in representing six different extreme rainfall indices: EnsMedian of NRMSE across six different observations.

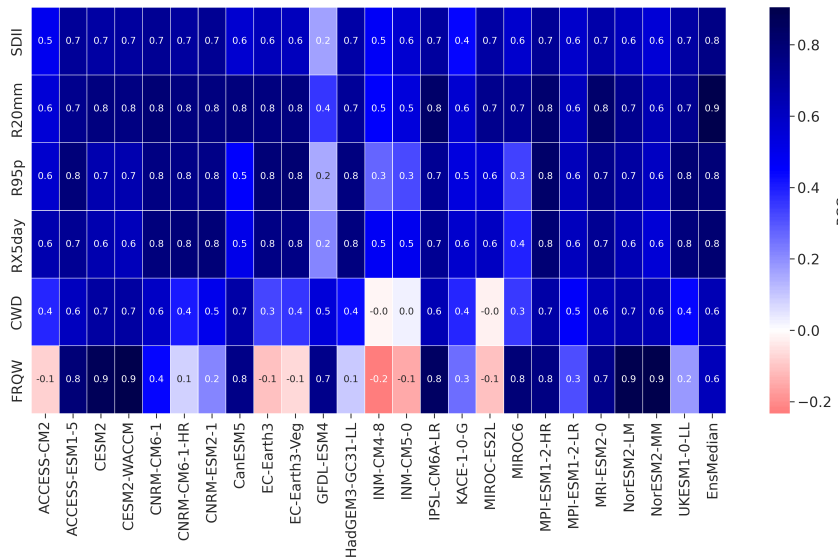


Figure C.9: Performance metrics of the individual GCM in representing six different extreme rainfall indices: EnsMedian of PCC across six different observations.

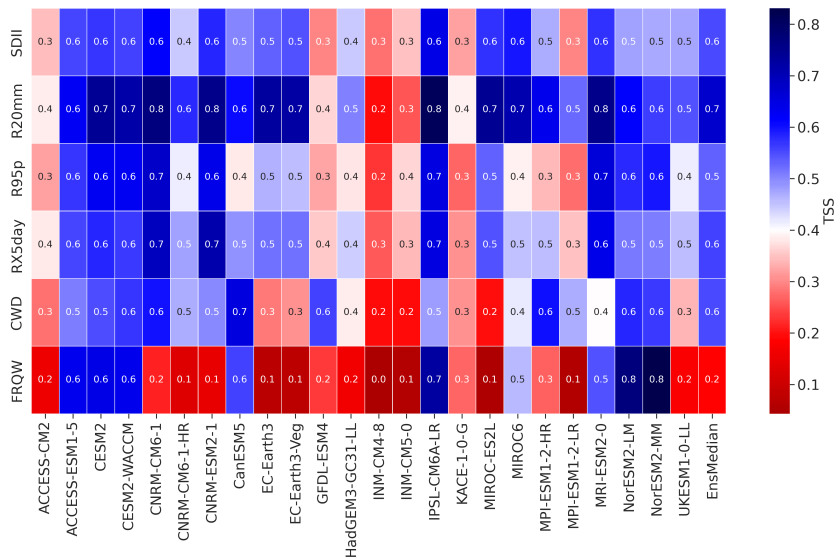


Figure C.10: Performance metrics of the individual GCM in representing six different extreme rainfall indices: EnsMedian of TSS across six different observations.

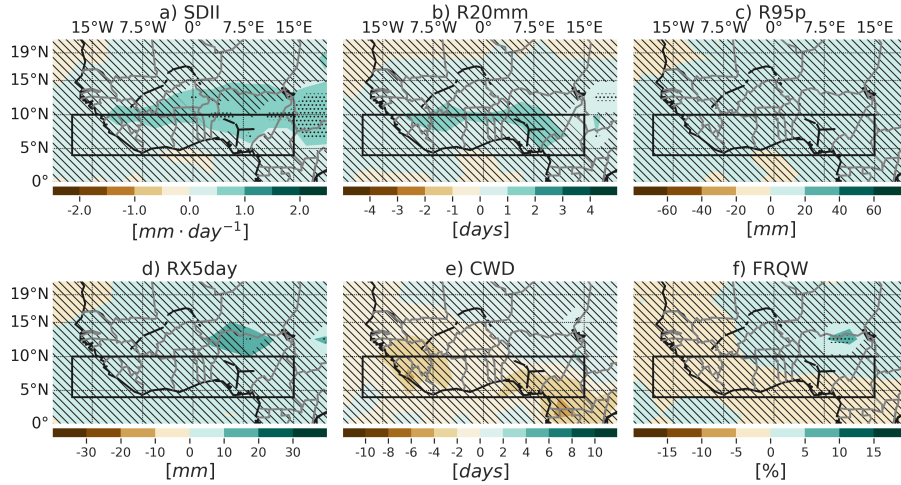


Figure C.11: Projected multi-model ensemble near-term median change in the JAS rainfall extreme indices over West Africa, relative to the present-day period (2021-2040 minus 1995-2014). Stippling indicates regions where the change robustly emerges from internal variability (at least 66% of the models show a change greater than the IAV and at least 80% of the models agree on the sign of the change); hatching (\) indicates regions where the change is unrobust (fewer than 66% of the models show change greater than the IAV); crossed lines (X) indicate conflicting signals where at least 66% of the models show change greater than the IAV, with less than 80% agreement on the sign of the change.

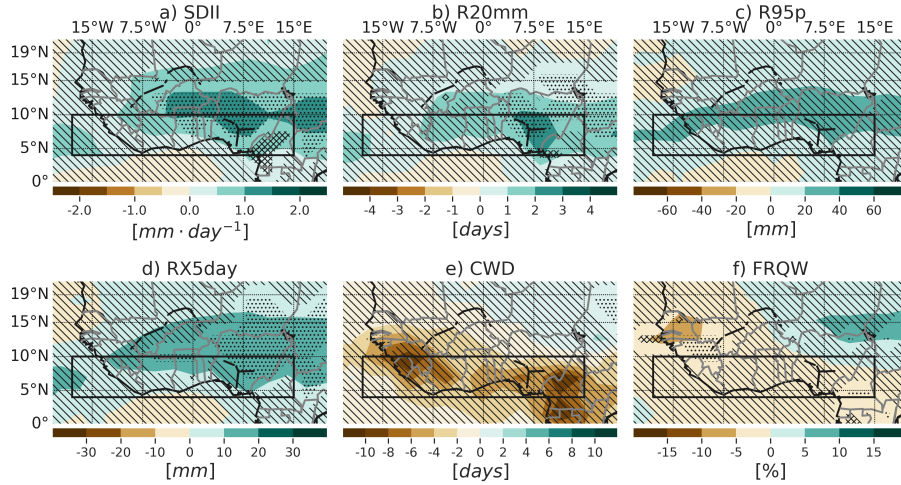


Figure C.12: Projected multi-model ensemble mid-term median change in the JAS rainfall extreme indices over West Africa, relative to the present-day period (2041-2060 minus 1995-2014). Stippling indicates regions where the change robustly emerges from internal variability (at least 66% of the models show a change greater than the IAV and at least 80% of the models agree on the sign of the change); hatching ($\backslash$) indicates regions where the change is unrobust (fewer than 66% of the models show change greater than the IAV); crossed lines (X) indicate conflicting signals where at least 66% of the models show change greater than the IAV, with less than 80% agreement on the sign of the change.

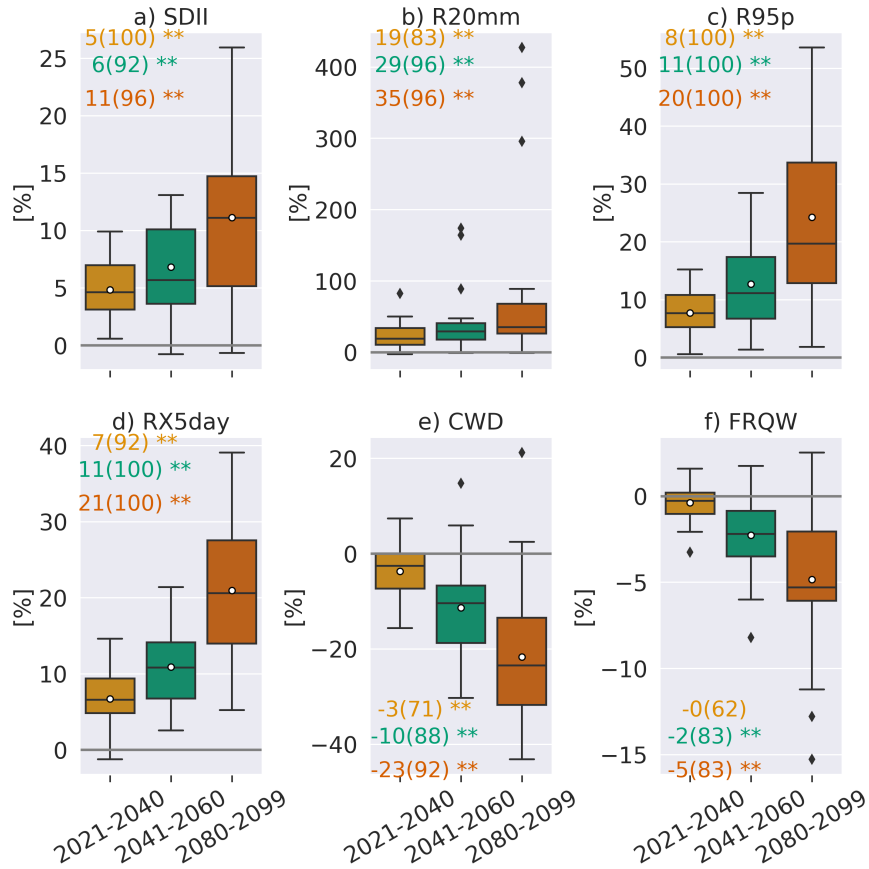


Figure C.13: Boxplot of the average of the near-term, mid-term and long-term percentage of changes (relative to the present-day period) in the mean of the rainfall extreme indices over the Guinea Coast. Each boxplot represents the distribution of 24 GCMs. Outliers are indicated by the black marks (losanges). The median (mean) value of each distribution is indicated with a black horizontal bar (a white circle). In each panel, the median value of the change percentages (and the percentage of agreement in parentheses) are indicated for each period, with the same color as the associated boxplot. The two stars indicate more than two-third agreement on the sign of the EnsMedian change among the 24 GCMs. The percentage of changes are computed at each grid point before performing the averages over Guinea Coast.

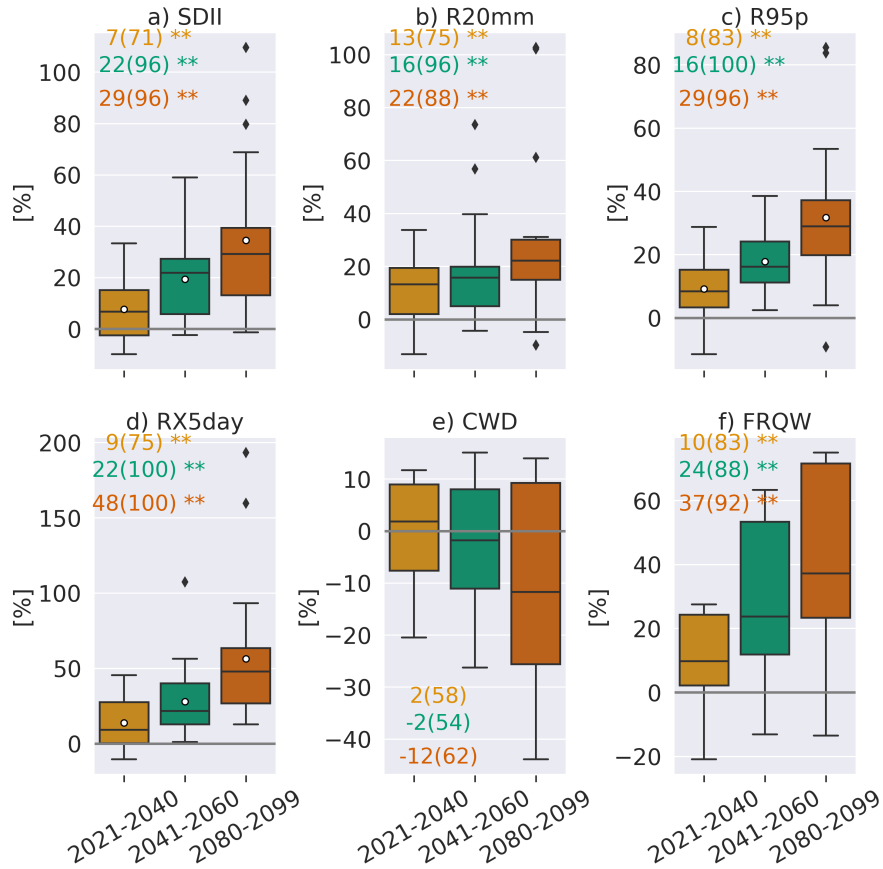


Figure C.14: Boxplot of the average of the near-term, mid-term and long-term percentage of changes (relative to the present-day period) in the standard deviation of the rainfall extreme indices over the Guinea Coast. Each boxplot represents the distribution of 24 GCMs. Outliers are indicated by the black marks (losanges). The median (mean) value of each distribution is indicated with a black horizontal bar (a white circle). In each panel, the median value of the change percentages (and the percentage of agreement in parentheses) are indicated for each period, with the same color as the associated boxplot. The two stars indicate more than two-third agreement on the sign of the EnsMedian change among the 24 GCMs. The percentage of changes are computed at each grid point before performing the averages over Guinea Coast.

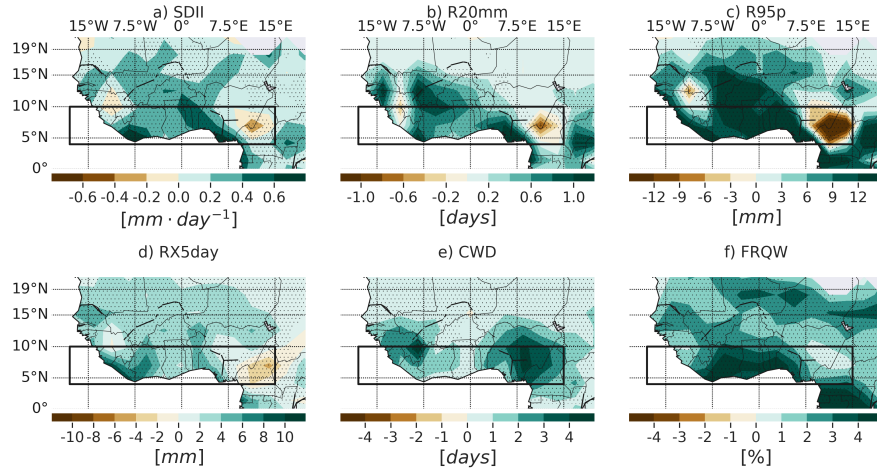


Figure C.15: *EnsMedian* the JAS observed rainfall extreme indices (from six different sources) regressed onto the standardized JAS AEM SST index (from HADISST dataset) over the 1995-2014 period. The stippling represents grid points where at least 4 data agree on the sign of the regression coefficient. The black rectangles indicate the Guinea Coast region.

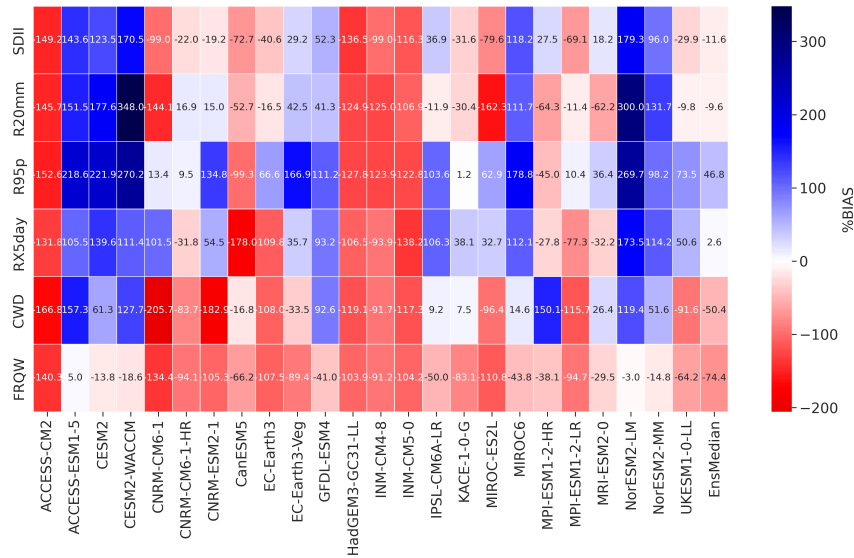


Figure C.16: Performance metric of the individual GCM and the 24 GCMs *EnsMedian* in representing six different extreme rainfall indices: median values of %BIAS across six different observations.

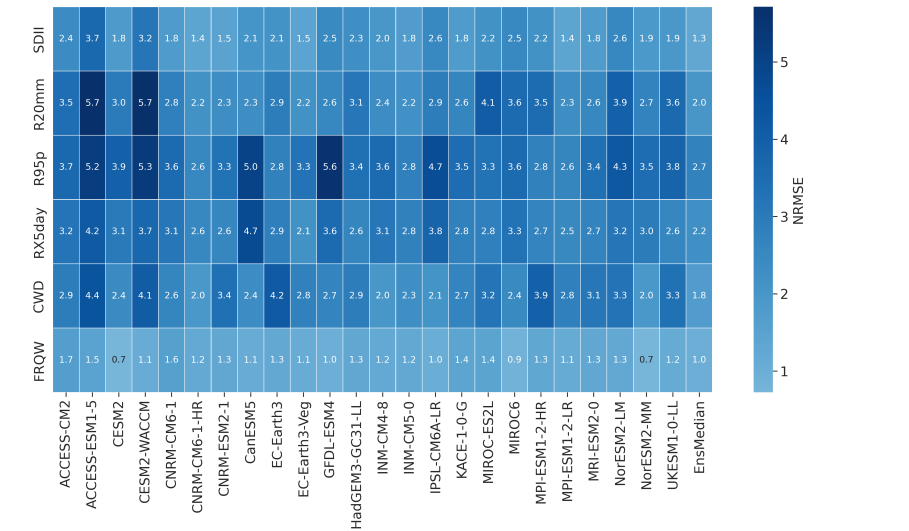


Figure C.17: Performance metric of the individual GCM and the 24 GCMs EnsMedian in representing six different extreme rainfall indices: median values of NRMSE across six different observations.

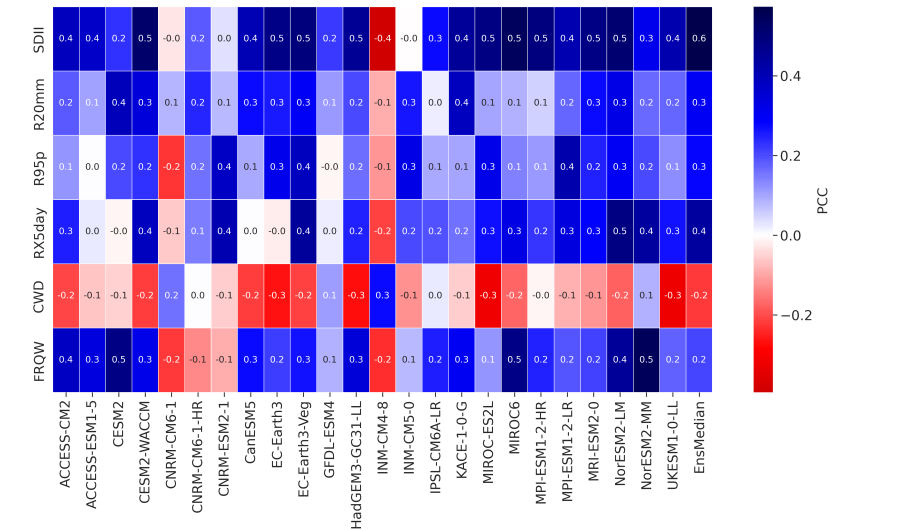


Figure C.18: Performance metric of the individual GCM and the 24 GCMs EnsMedian in representing six different extreme rainfall indices: median values of PCC across six different observations.

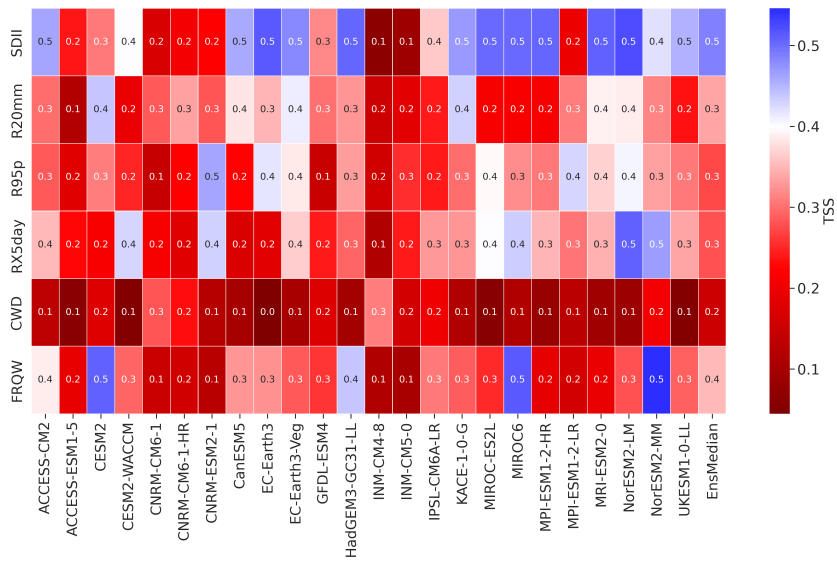


Figure C.19: Performance metric of the individual GCM and the 24 GCMs EnsMedian in representing six different extreme rainfall indices: median values of the TSS across six different observations.

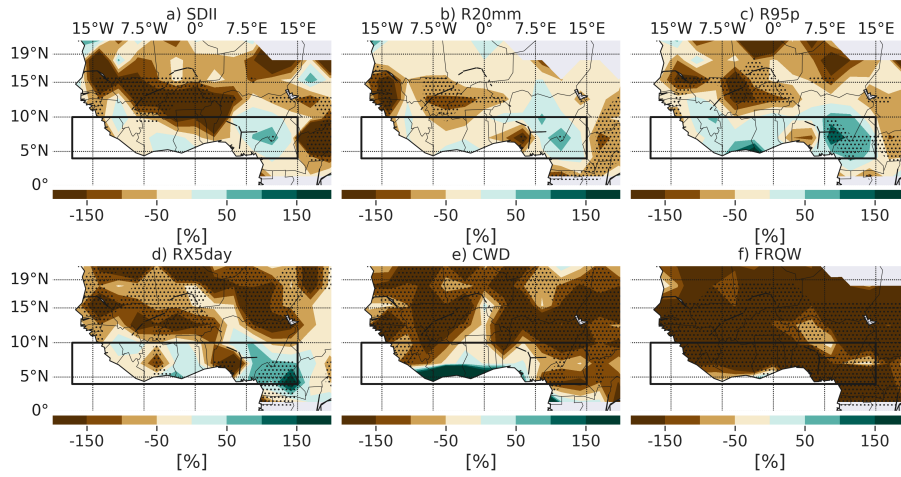


Figure C.20: Spatial distribution of the EnsMedian bias percentage relative to the interquartile range of the six observations patterns for the JAS regression pattern of the extreme rainfall indices related to one standard deviation of the Atlantic equatorial mode over 1995-2014. In total, there are six observations $\times$ 24 models = 144 different biases. The stippling indicates regions where two-third of the biases agree on the sign of the EnsMedian of the 144 biases. The black rectangles indicate the Guinea Coast region.

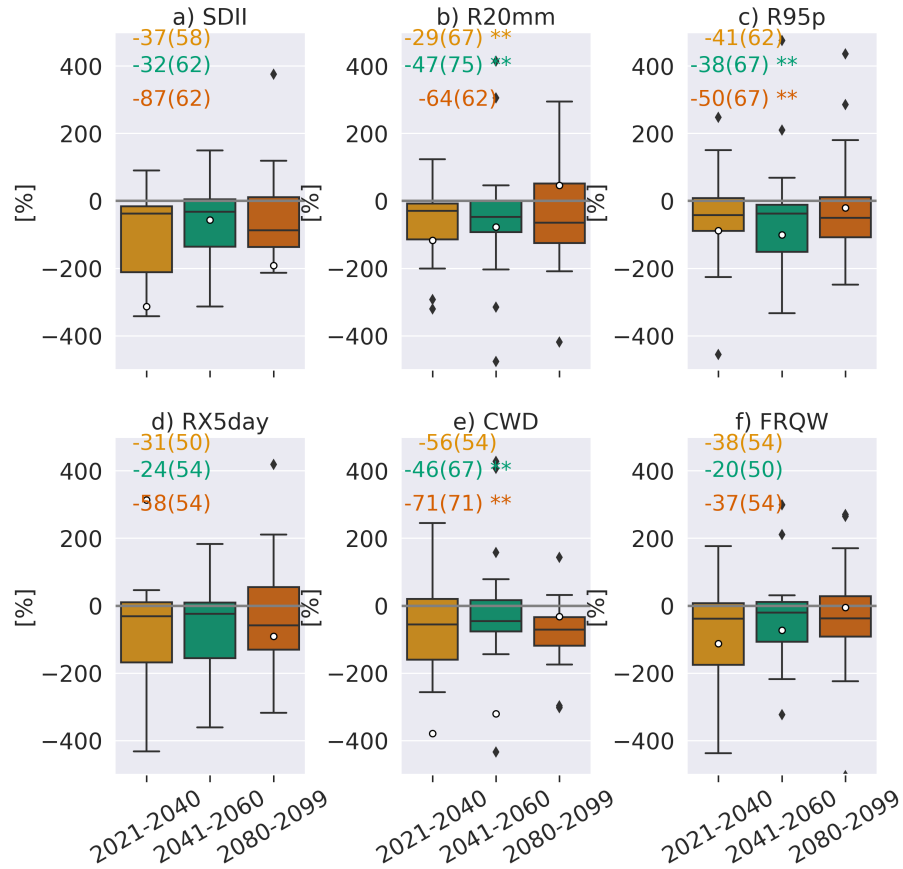


Figure C.21: Near-term, mid-term and long-term percentage of changes in the Guinea Coast area-average of the JAS extreme rainfall responses to one standard deviation of the JAS AEM index. Each boxplot represents the distribution of 24 GCMs. Outliers are indicated by the black marks (losanges). The median (mean) value of each distribution is indicated with a black horizontal bar (a white circle). In each panel, the median value of the change percentages (and the percentage of agreement in parentheses) are indicated for each period, with the same color as the associated boxplot. The two stars indicate more than two-third agreement on the sign of the EnsMedian change among the 24 GCMs. These changes are computed from the average of the regression coefficients over Guinea Coast for each period.

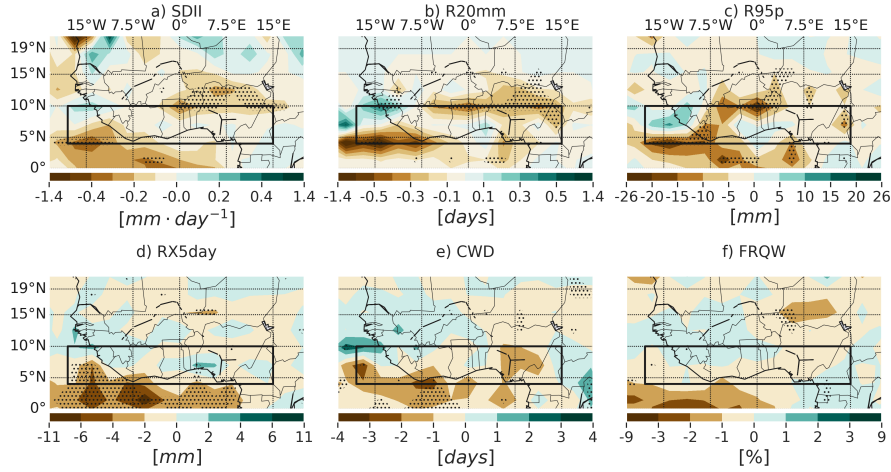


Figure C.22: Near-term changes in the regression patterns of the JAS extreme rainfall indices associated with the standardized JAS AEM SST index (2021-2040 minus 1995-2014). The stippling indicates grid points where two thirds of the models agree on the sign of the change.

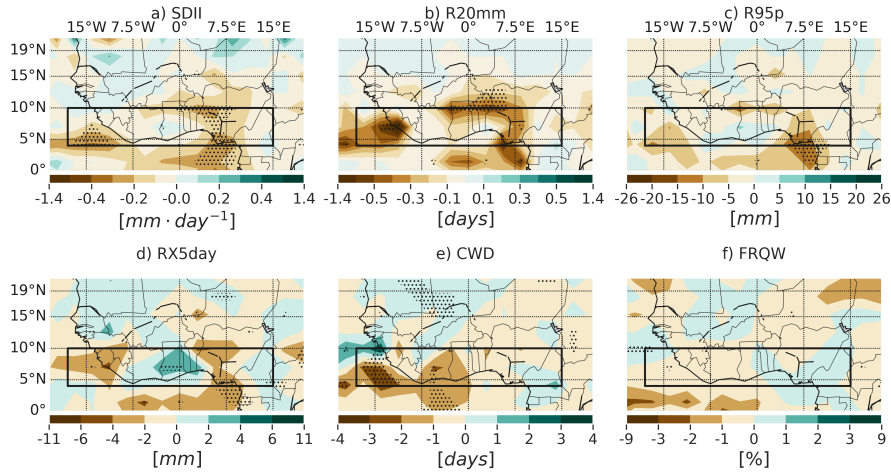


Figure C.23: Mid-term changes in the regression patterns of the JAS extreme rainfall indices associated with the standardized JAS AEM SST index (2041-2060 minus 1995-2014). The stippling indicates grid points where two thirds of the models agree on the sign of the change.

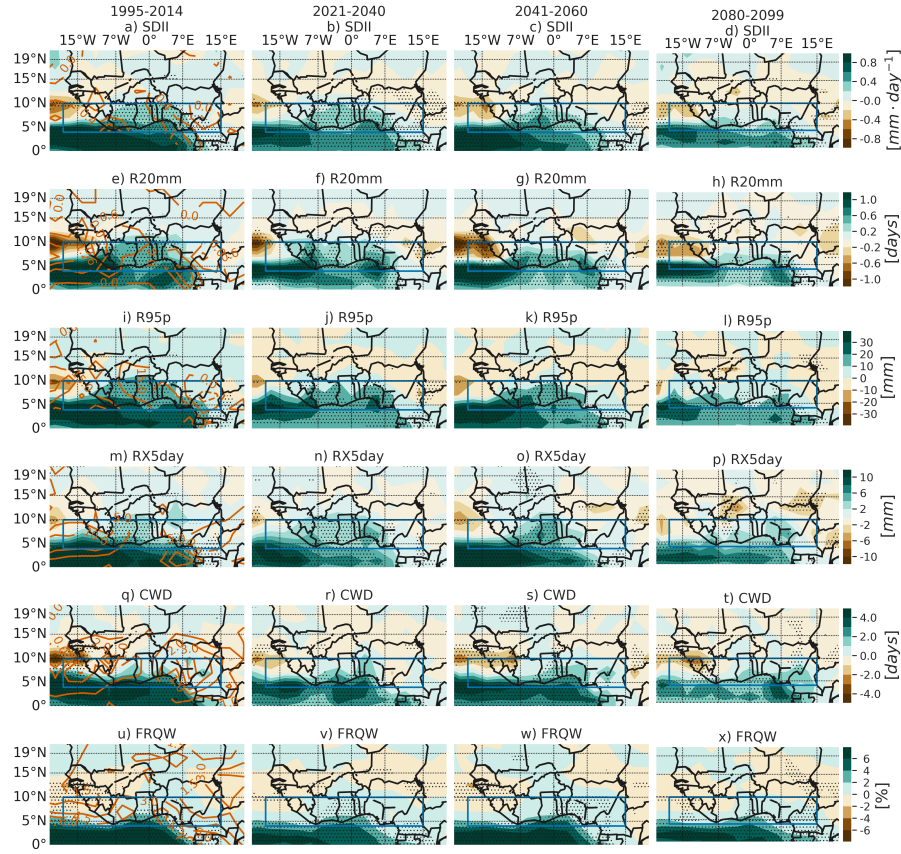


Figure C.24: Maps of the JAS rainfall extreme indices regressed onto the standardized JAS AEM SST index over the 1995-2014 period, 2021-2040, 2041-2060 and 2080-2099 periods. The stippling represents grid points where two thirds of the models agree on the sign of the EnsMedian regression coefficient across 24 GCMs. Contours in the 1995-2014 panels indicate the multi-observation median values of the regression coefficients from six different rainfall datasets. The black rectangles indicate the Guinea Coast region.

D

List of publications related to the thesis

First author

- Worou, K., Goosse, H., Fichet, T., Guichard, F., and Diakhate, M.: Interannual variability of rainfall in the Guinean Coast region and its links with sea surface temperature changes over the twentieth century for the different seasons. *Climate Dynamics* **55**, 449–470 (2020). <https://doi.org/10.1007/s00382-020-05276-5>. (Chapter 2)
- Worou, K., Goosse, H., Fichet, T., and Kucharski, F.: Weakened impact of the Atlantic Niño on the future equatorial Atlantic and Guinea Coast rainfall, *Earth System Dynamics*, **13**, 231–249, <https://doi.org/10.5194/esd-13-231-2022>, 2022. (Chapter 3)
- Worou, K., Fichet, T., and Goosse, H.: Future changes in the mean and variability of extreme rainfall indices over the Guinea Coast and role of the Atlantic equatorial mode, *Weather and Climate Dynamics Discuss.* [preprint], <https://doi.org/10.5194/wcd-2022-53>, in review, 2022 (Chapter 4)

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