

*Full Length Research Paper*

# West African Monsoon (WAM) and Atlantic Cold Tongue (ACT) onsets using Regional Climate Model

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This study focused on the occurrences and relationships of the Atlantic Cold Tongue (ACT) and the West African Monsoon (WAM) onsets in the Sudano-Sahelian region during 1983-2000. Two simulations (WAFSSTERA and WAFSSTREY) were used to investigate the sensitivity of the Regional Atmospheric Model (MAR) for two different datasets of sea surface temperature (ERA-40 and Reynolds SST datasets). The MAR reproduced the rainfall high values in the Guinean zone. A band of weak precipitation (<1 mm/day) was observed in the two simulations. WAFSSTERA gave more annual daily average rainfall (~1.6 mm/day) than WAFSSTREY. The simulated onset date index (ODI) of WAM showed more northward penetration of monsoon flow, particularly during flood years in the Gulf of Guinea. These results also indicated that the monsoon onset did not depend on the SST dataset used to force the model. Finally, the negative difference between ODIs and ACT onset date could be due to warm SST. The yearly evolution of the southern heat transport was also in agreement with this difference. It showed yearly succession of negative and positive anomalies around 7.5°N (mean rainy equilibrium position over West Africa). Such study could be useful for the forecast of WAM onset and ACT scenario relationships.

**Key words:** Atlantic cold tongue, onset date index, regional climate model, sea surface temperature, West African Monsoon.

## INTRODUCTION

In the last twenty years, many studies have been conducted to understand how the surface conditions could influence the West African Monsoon (WAM) precipitations. These works, which particularly focused on Sahel region, identified the global Sea Surface Temperature (SST) (Lamb, 1978; Folland et al., 1986; Janicot et al., 1996), the continental surface conditions

(Nicholson, 1989; Cook, 1994) and the large-scale circulation as potential factors (Nicholson, 2013). For example, Lamb (1978) found that the deficit on the Sahelian rainfall is associated with negative and positive tropical Atlantic SST anomalies at north and south of 10°N, respectively. Using a General Circulation Model (GCM), Rowell et al. (1992) highlighted the influence of

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regional and global SST variations on Sahel rainfall regime at various timescale.

Moreover, Vizu and Cook (2001) found that the impact of SST anomalies on West African rainfall was more important in the Gulf of Guinea than in the north of the tropical Atlantic. They also investigated the sensitivity of some Regional Climate Models (RCM) to regional SST anomalies and found that RCMs simulated better the main processes that influence the inter-annual variability of the monsoon during the June and July periods.

Giannini et al. (2003) suggested that rainfall variability in the Sahel results from the oceanic forcing on the monsoon, which in return is amplified by the land-atmosphere interactions. The recent trends of Sahel drought were attributed to high SST values, which favours the establishment of a deep convection zone over the ocean and weakens the convergence of the monsoon flux over the continent.

The rainfall regime over West Africa, which is located at lower latitudes during March-September, is mainly driven by the monsoon cycle. The WAM is characterized by a pronounced seasonal wind shift due to thermodynamic contrasts between the Sahara and the equatorial Atlantic Ocean (Eltahir and Gong, 1996). Thereafter, the southwestern flow, established between the Atlantic and Saharan heat low, brings moisture into the continent. The WAM is characterized by a strong spatial and temporal variability due to convective rain, modulation of the seasonal cycle and synoptic disturbance activities. During the onset phase in March to June, the rain band extends from the Guinea coast to the north. The rain period in July to September starts with an abrupt shift of the core of the rain band roughly from 5°N to 10°N (Sultan and Janicot, 2000). This period is followed by a gradual southward retreat of the rain belt during September to November.

The Gulf of Guinea is characterized by a warm SST up to 28°C in April when the rain band is located along the coast. In July, the cooling of this oceanic region occurs because of many processes such as coastal and equatorial upwellings, which result in the formation of the Atlantic Cold Tongue (ACT) (Caniaux et al., 2011). These cold waters influence the monsoon system by favouring a SST gradient through the equator and strengthening the breeze effects to the continent by increasing the thermal gradient. Thus, the pressure gradient is reinforced and drives the convection zone northward (Okumura and Xie, 2004).

Chang et al. (2008) showed that drought over West Africa occurs for a reduced thermo-haline circulation in the northern Atlantic during Ice Age. The subsurface flux of the northern Brazil current reverses when this circulation weakens. It induces a reduction of the ocean upper layers stratification and a warming of the southern equatorial Atlantic SST that consequently reduces the wind of the monsoon flux and the precipitations over West Africa. Messenger et al (2003) also studied the West

African rainfall sensitivity to the increase of SST in the Gulf of Guinea by using the regional climate model, so-called Regional Atmospheric Model (MAR). They showed that SST is the main factor that influences the precipitations beyond 12°N at both seasonal and inter-annual timescales. A warmer SST is associated with a decrease in the magnitude of the African Easterly Jet and an increase in northward water content transport.

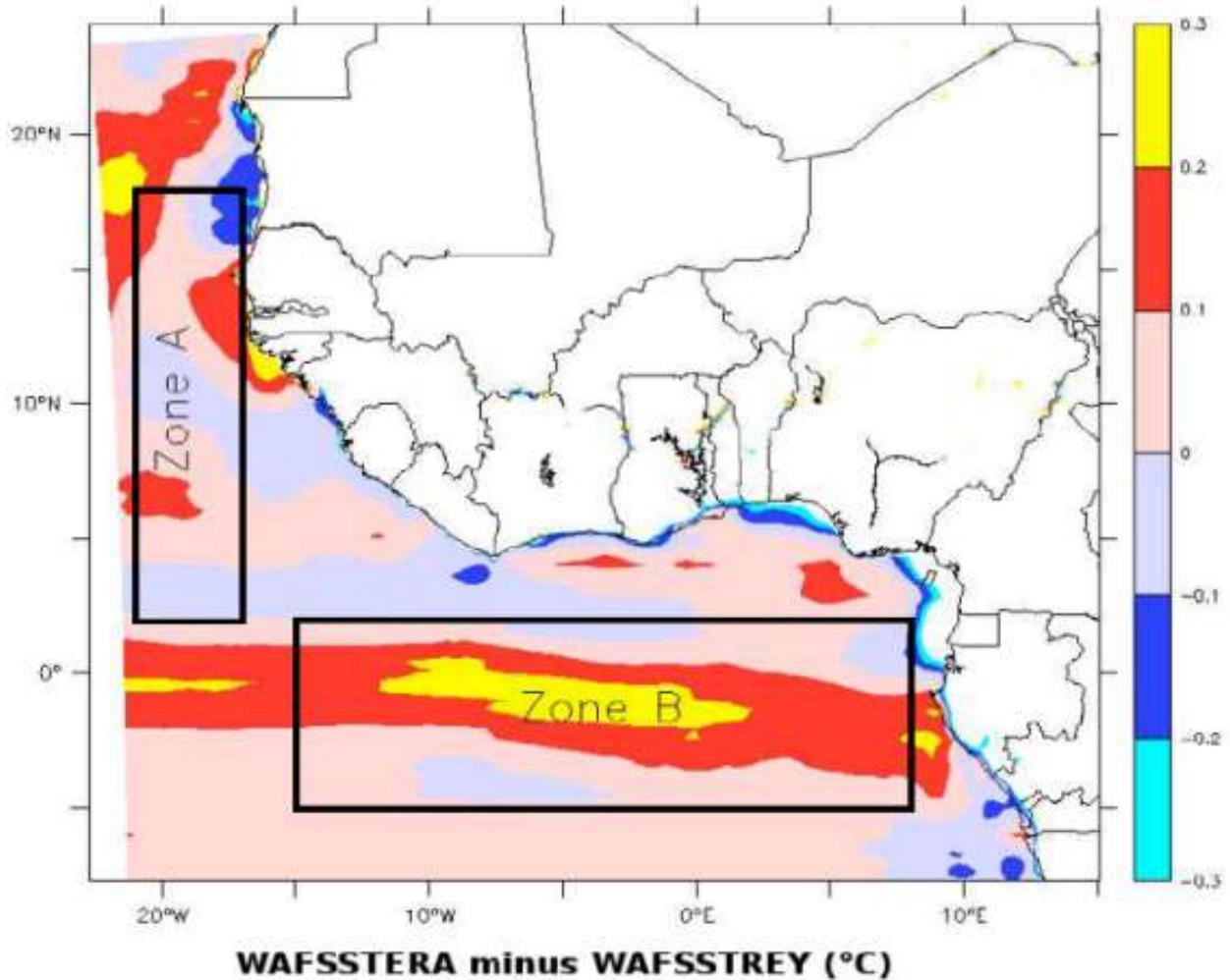
In this study, two SST datasets (that is, the Olv1SST data of Reynolds and Smith (1994) and the 2DVarSST of ERA-40 reanalysis) are used as surface forcing in two MAR simulations. These SST data have not been assimilated with the same methods. For example, SST data of Reynolds and Smith (1994) use the optimum interpolation method (Olv1), while those of the European Centre for Medium-range Weather Forecast (ECMWF) reanalyses (ERA-40) are analyzed variables produced by using the variational assimilation technique (2D-VAR). The 2DVarSST is believed by the producers to be more accurate since biases in Olv1SST are corrected in 2DVarSST version. Despite these facts, climate variables simulated by regional models are likely to contain biases when using Reynolds SST (Olv1SST) and ERA-40 SST (2DVarSST). Thus, the effects on rainfall, surface fluxes and WAM regime should be analysed to justify the use of any of these datasets.

The topic of the installation of WAM and SST in the tropical Atlantic has been studied by several authors (Fontaine and Louvet, 2006; Brandt et al., 2011; Diaconescu et al., 2015; Vellinga et al., 2013). Diaconescu et al. (2015) focused their evaluation on various daily precipitation statistics and on the monsoon onset and retreat over the Sahel region. They used daily precipitations of the AFRICA-CORDEX experiment from two ERA-Interim driven Canadian RCMs. Vellinga et al. (2013) explored onset predictability in an ensemble prediction system (GloSea4) and compare it to onset in two reanalysis datasets. They found that many models show the influence of the tropical Atlantic on the onset predictability. The work of Brandt et al. (2011) evaluated the correlation between the WAM and ACT by using data of observation. The present study aims to simulate the WAM onsets in the Sudano-Sahelian region using MAR and to analyse their relationships with occurrences of the ACT. This work could be useful in improving study of the relationship between WAM onset and ACT and, in addition, investigates the influence of the ERA-40 2DVarSST and Reynolds Olv1SST on regional climate simulations in the northern area of the Gulf of Guinea with MAR.

## METHODOLOGY AND DATA

### Model description

The Regional Atmospheric Model (MAR) model is used at the Catholic University of Louvain and at the Institute of Astronomy and Geophysics, Georges Lemaître (Belgium). It is also used at the



**Figure 1.** Mean regional SST differences between ERA40 and Reynolds from 1984 to 2000. Colour interval is 0.1°C. Zones A and B represent the areas of the two SST data set comparisons.

Laboratory of Atmospheric Physics and Fluid Mechanics of the University of Felix Houphouët-Boigny of Cocody (Côte d'Ivoire) (Yoroba et al., 2011; Kouassi et al., 2012). A short description of the MAR model is given here with emphasis on its adaptation for West African climate simulations. For a complete description of the model, Gallée and Schayes (1994) and Vanvyve et al. (2008) researches should be considered. The MAR is a hydrostatic primitive equation model in which the vertical coordinate is the normalized pressure; one where no approximation is made in the mass conservation equation. The hydrological cycle of MAR (Gallée, 1995) includes a cloud microphysical model, with conservation equations for cloud droplet, raindrop and cloud ice crystals and also snowflake concentrations. The parameterization of the warm cloud microphysics is based on the work of Kessler (1969). The solar radiation scheme is taken from Fouquart and Bonnel (1980), while the long-wave radiation scheme follows a wide-band formulation of the radiative transfer equation (Morcrette, 1984). Cloud properties are taken into account in the solar and infrared radiation schemes by computing the liquid water path in each model layer from the concentration of cloud droplets and ice crystals. The MAR is adapted to tropical regions by including the convective adjustment scheme of Bechtold et al. (2001). In addition, the atmospheric part of the model is coupled to the Soil Vegetation

Atmosphere Transfers model (SVAT) (De Ridder and Gallée, 1998), which contains one vegetation layer and seven unevenly spaced soil layers, with a finer resolution near the surface. The surface forcing is made with the SST derived from ERA-40 reanalysis data or from the prescribed SST of Reynolds (Reynolds and Smith, 1994).

#### Data

The simulation is made over the domain 23°W - 15°E and 7.2°S - 24°N, including the southern tropical Atlantic and the West Africa (Figure 1). The model output variables have a finer resolution with 40 km horizontal grid and 40 vertical levels. In this study, the continental soil moisture and soil temperature as well as atmospheric variables are initialised and forced at MAR lateral boundaries using the ERA-40 reanalysis data, which has a 1° × 1° horizontal grid, 60 vertical levels and 6-h time step. SSTs are forced from ECMWF and Reynolds dataset.

Two simulation experiments are performed using, 2DvarSST of ERA-40 (hereafter WAFSSTERA) and Olv1SST of Reynolds (hereafter WAFSSTREY), respectively. These simulation experiments extend from 1st of January, 1983 to 31st of December

**Table 1.** Experimental set-up.

Simulations	Domain	Size grid	SST forcing	Period
WAFSSTERA	23°W---15°E; 7.2°S---24°N	40 km	2DvarSST	1983 - 2000
WAFSSTREY	23°W---15°E; 7.2°S---24°N	40 km	Olv1SST	1983 - 2000

Model characteristics	Used options
Basic equations	3D, hydrostatic at grid point, compressible
Vertical coordinates	Normalized pressure $\sigma$
Type of projection	Stéréographic oblic
Horizontal grid	40 × 40 km
Lateral-boundary conditions	Dynamic relaxation with a 7-point grid buffer at the edge of the domain
Boundary conditions at the top of the model	Absorption layer
Boundary conditions at the soil surface	U, V = 0; T = Ts(t); q = qs(t)
Advection numerical-scheme	Leapfrog-scheme
Eddy scheme	E- $\epsilon$ closure scheme, order 1.5
Radiative transfer-scheme	ECMW
Diagram of cloud-microphysics	Explicit representation of the hydrological cycle
Convection scheme	Mass flow scheme of Bechtold (2001)
Surface scheme	SISVAT

2000 (Table 1), with a spin-up in 1983 for the SVAT model. ERA-40 SST data are originally reported on a  $1^\circ \times 1^\circ$  horizontal grid every 6 h. Olv1SST data of Reynolds are provided on a  $1^\circ \times 1^\circ$  horizontal grid and a daily resolution. The two SST datasets are both spatial and temporal, interpolated to a horizontal grid-with  $\sim 40$  km grid mesh and to a 6 h temporal resolution.

Climatic Research Unit (CRU) data are used to evaluate the model skill in simulating the rainfall trend and distribution. Monthly rainfall data from CRU (New et al., 2000) are registered by different rain gauge stations across the world and have been statistically interpolated on a  $0.5^\circ \times 0.5^\circ$  regular grid. Pocard and Janicot (2004) showed that this database is suitable for the rainfall variability studies over West Africa. The updated data extended from 1901 to 2007, but only the period 1983 to 2000 is used in this study, as previously stated.

The Louvet (2008) and Brandt et al. (2011) onset datasets are used as reference in the MAR onset evaluation. With regard to the above-mentioned dates, the Climate Prediction Centre Merged Analysis of Precipitation (CMAP) data (Louvet, 2008) and the Global Precipitation Climatology Project (GPCP) dataset (Brandt et al., 2011) were used to find the onset dates. They represent the first pentad of a 20-days period registering positive difference between mean GPCP precipitations at north of  $7.5^\circ\text{N}$  and south of this latitude. ACT data, used in Brandt et al. (2011), are extracted from Caniaux et al. (2011). ACT onset date is defined as the period when the surface area, defined by SST below  $25^\circ\text{C}$  inside the domain  $30^\circ\text{W} - 12^\circ\text{E}$  and  $5^\circ\text{S} - 5^\circ\text{N}$ , exceeds the empirically fixed threshold of  $0.4 \times 10^6 \text{ km}^2$ .

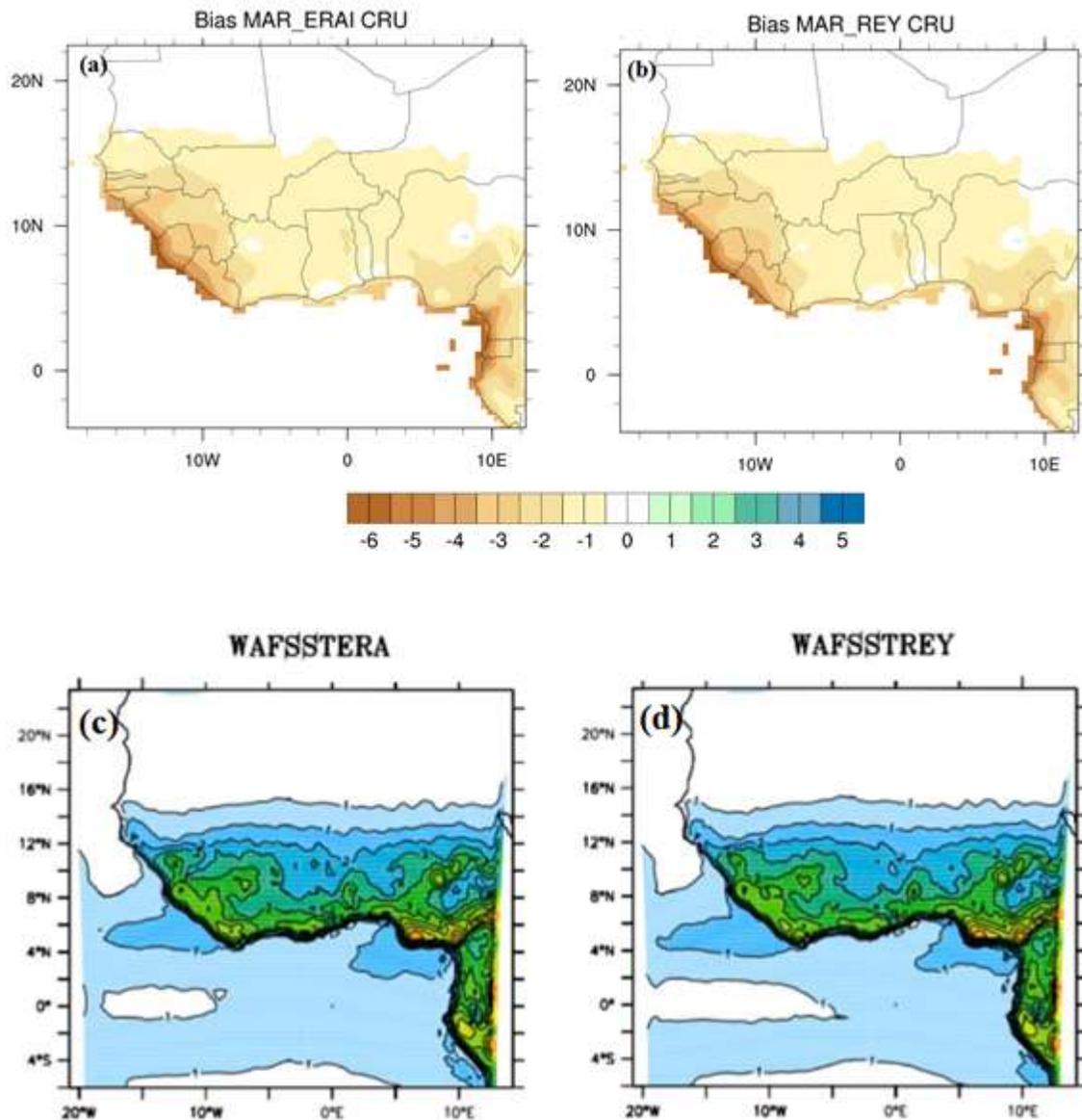
## RESULTS AND DISCUSSION

### Rainfall daily trend and annual variability as depicted by the two SST forcing data

#### *A rainfall daily trend from the two simulation experiment*

Figure 1 presents the mean (1984 to 2004) SST

difference between the WAFSSTERA and WAFSSTREY simulations. It shows warmer WAFSSTERA than WAFSSTREY in almost the entire domain. The most notable differences ( $\sim 0.3^\circ\text{C}$ ) are observed in both the Senegalese (for zone A) and the equatorial (for zone B) upwelling regions, which are subjected to strong variations of SST in the north of Gulf of Guinea (Servain et al., 1985). Along the coastal regions, WAFSSTREY seems to be warmer than WAFSSTERA, which may be related to biases introduced by the optimal interpolation method near the coast. Figures 2a and b illustrates the biases between the mean (1984 to 2004) of the daily precipitation fields of the two MAR simulations (WAFSSTERRA in Figure 2a and WAFSSTREY in Figure 2b) and the CRU dataset. These biases calculations have been performed from WAFSSTERA (Figure 2a) and WAFSSTREY (Figure 2b) and allow characterization of the difference between the simulation results compared to CRU dataset. Figures 2c and d show the mean (1984 to 2000) of the daily precipitation fields simulated by the two MAR simulations. The MAR model reproduces the high values of the rainfall over the Guinean zone as well as over the Fouta-Djalou and Cameroon mountains, even if the values are slightly underestimated. These simulated lower values could be due to the spatial resolution of the MAR model (Table 1) that smoothens the topography. The biases reach  $\sim 1$  mm/day in the rest of the West Africa area where the altitude is low. The isohyet of 1 mm/day is localized at  $15^\circ\text{N}$  for the two simulations, while it is positioned at  $17^\circ\text{N}$  in CRU data (not shown). Ramel (2005) and Yoroba (2012) observed that this anomaly in the isohyets position is not related to SST forcing but rather to the physics of the model. The area covered by the isohyet 1 mm/day is reduced over the Cameroon



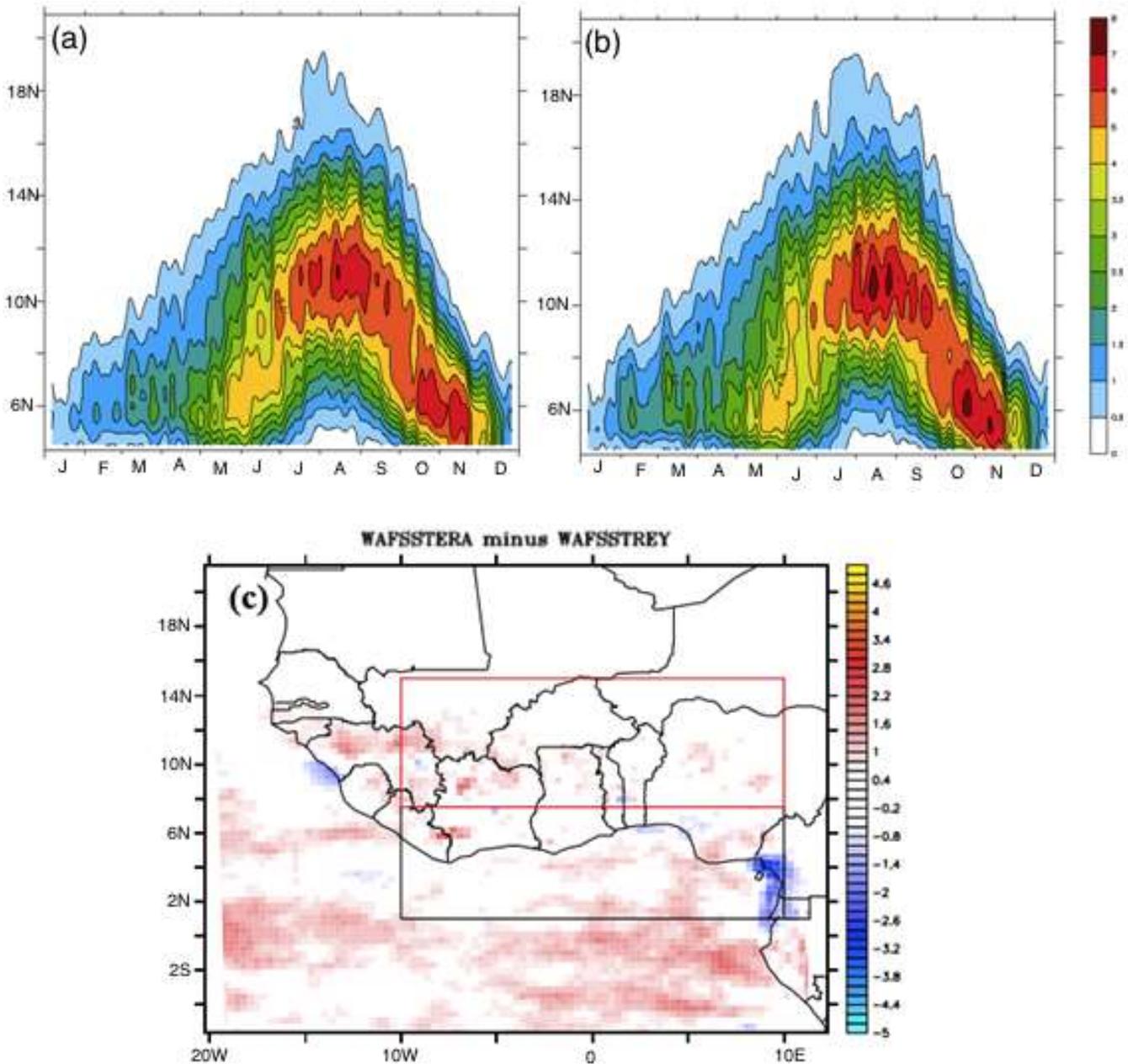
**Figure 2.** Biases between the mean spatial structures of the daily precipitation fields simulated for (a) WAFSSTERA and (b) WAFSSTREY during 1984 - 2000 period. Mean spatial structure of the daily precipitation for (c) WAFSSTERA and (d) WAFSSTREY during 1984 - 2000 period.

Mountains and is larger over the Fouta-Djallon in the WAFSSTERA simulation. Both WAFSSTERA and WAFSSTREY simulations show also some differences over the tropical Atlantic Ocean. A band of weak precipitation ( $< 1$  mm/day) extends from  $20^{\circ}\text{W}$  to  $4^{\circ}\text{W}$  at the equator with WAFSSTREY, while it is confined between  $18^{\circ}\text{W}$  and  $10^{\circ}\text{W}$  with WAFSSTERA. These two zones are included in the equatorial band, which (i) displays a high difference between the two SST dataset (Figure 1) and (ii) corresponds to the ACT area (Caniaux et al., 2011). The cold Olv1SST cools the air above the ocean surface and makes it too dense for rising and

condensing into water vapour. Thereafter, raindrops and clouds cannot form and air is free of clouds for a sufficient long period of time that allows confinement of the rain eastward, near the coast of Gabon.

#### **Rainfall annual variability from the two simulation experiments**

Figure 3 shows the time-longitude diagrams of the simulated mean annual rainfall averaged between  $10^{\circ}\text{W}$  and  $10^{\circ}\text{E}$  for WAFSSTERA (Figure 3a) and WAFSSTREY



**Figure 3.** Time-latitude diagrams of the mean rainfall of each day (mm/day) simulated by MAR, during the 1984-2000 period for (a) WAFSSTERA and (b) WAFSSTREY (averaged from 10°W to 10°E). (c) Spatial difference of the annual daily mean rainfall (mm/day) between both WAFSSTERA and WAFSSTREY simulations.

(Figure 3b) (Moufouma, 2003; Gallée et al., 2004). It also displays the mean annual difference from the two simulated rainfall over the 17-years data (Figure 3c).

The latitudinal migration of the Inter-Tropical Convergence Zone (ITCZ) is well represented in both simulations through the annual evolution of the daily rainfall (Figure 3a and b). The monsoon onset begins in mid-March (in Guinean zone) with a succession of more or less intense precipitations (~2 mm/day). At that point,

the rainfall band displaces irregularly northward, from the West African coast (5°N - 8°N) to the Sahel region (10°N - 14°N), between May and August, with an amount reaching 9 mm/day. Such latitudinal migration is consistent with the “monsoon jump” discussed by Sultan and Janicot (2000). The southward displacement of the precipitations from September to November seems to follow a more regular movement.

The mean annual difference of the simulated rainfall

over the 17-year data (Figure 3c) shows more rainfall with WAFSSTERA than WAFSSTREY in almost the entire Gulf of Guinea and over the continent. This difference reaches +1.6 mm/day in the equatorial Atlantic band (2°S---2°N) and +1.3 mm/day westward at 12°N on the continent. The rainfall difference that appeared between these two simulations could be due to the high values of the 2DvarSST compared to those of the Olv1SST (Figure 1), which leads to more moisture in the lower atmospheric layers, above the ocean.

However, negative bias (~-1.6 mm/day), which suggests more rainfall with WAFSSTREY than WAFSSTERA is observed near the Congo River discharge and at the Senegalese upwelling zone. These differences are probably also due to those between the two SST datasets. Indeed, Olv1SST is a combination of observed values, while 2DvarSST is from reanalyses. Thus, WAFSSTERA could not capture all the physical phenomena that concurred to the SST variability around the Senegalese coast and the Congo River, as for example, the sea surface salinity and river runoff (Da-Allada et al., 2013).

### **Onset of West African Monsoon and Atlantic Cold Tongue**

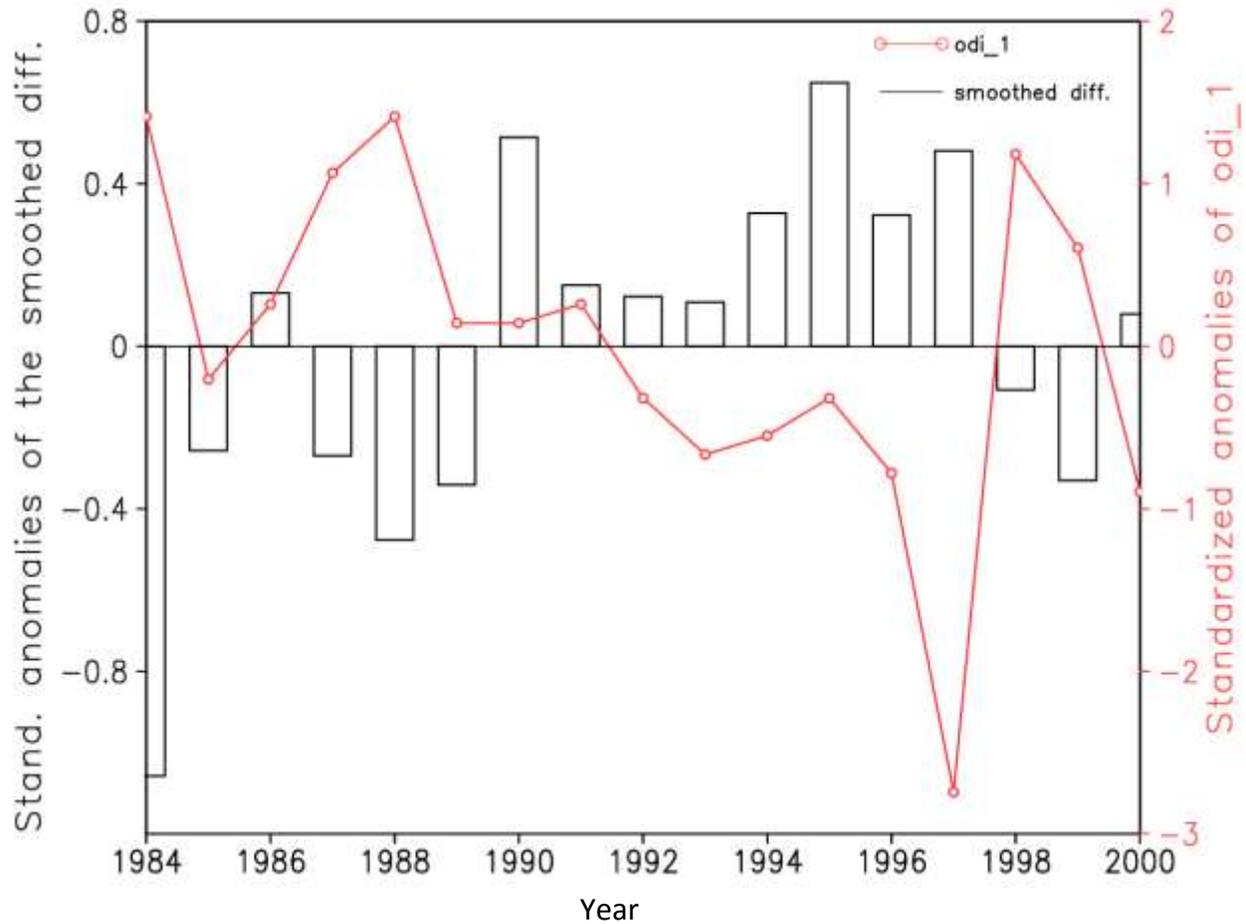
Here, relationships between the onsets of the West African Monsoon (WAM) and the Atlantic Cold Tongue (ACT) are analyzed using the two MAR simulation experiments. Brandt et al. (2011) observed that the inter-annual variability of SST in the ACT, during the boreal summer, is closely related to the rainfall variability in the countries facing the Gulf of Guinea and, particularly to WAM onset (Fontaine and Louvet, 2006).

### ***Use of the onset date index (ODI)***

The index used to describe the WAM onset is calculated in three steps, following the method of Fontaine and Louvet (2006), Vellinga et al. (2013) and Diaconescu et al. (2015). The mean daily rainfall both at the south (1°N - 7.5°N) and at the north (7.5°N - 15°N) of 7.5°N (Figure 3c) is computed first. This latitude (~7.5°N) corresponds to the mean latitudinal position between the two equilibrium latitudes (5°N and 10°N), associated with the major rainy season of the Guinean and Sudano-Sahelian zones. The second step consists in computing the difference between the northern (7.5°N - 15°N) mean daily rainfall and that at the southern (1°N - 7.5°N). This difference illustrates the latitudinal migration of the ITCZ and the contrast between the ocean and the continent. Thereafter, a moving average is applied to these last values to eliminate the fluctuations less than 10 days. When the rain band migrates northward (respectively southward), this smoothed difference increases

(respectively decreases). Considering this definition, the onset date of the rainy season - noted Onset Date Index (ODI) - is considered as the Julian date after which the smoothed difference has successive positive values for at least 20 days. This threshold is used to ensure the detection of a strong and robust climate signal, as for example, the "monsoon jump" (Sultan et al., 2003) and not the beginning of the first Sudano-Sahelian rains. One notes that the smoothed difference is related to rainfall, while ODI is about Julian days. The ODIs are noted ODI\_1 for WAFSSTERA simulation and ODI\_2 for the WAFSSTREY simulation. The difference, only described, between ODI\_1 and ODI\_2 is close to 1 day, except in 1994 and 1999, where the difference is 4 days. As these two ODIs are quite similar, Figure 4 shows the inter-annual evolutions of the ODI\_1 standardized departure from the mean date and evolutions of the standardized daily departure, during July, from the smoothed difference between the northern (7.5°N - 15°N) and the southern (1°N - 7.5°N) mean daily rainfall. A more northward (respectively southward) penetration of the monsoon flow, that means positive (respectively negative) values of the smoothed difference, occurs when ODI\_1 is observed early (respectively later). That means negative (respectively positive) values of ODI\_1, leading to a correlation coefficient of -0.68 at 95% significant level. For instance, during 1984 and 1987, which are known to be flooding years on the coast of the Gulf of Guinea (Kouadio et al., 2003; Kouadio et al., 2007, 2011), the ODI\_1 values are above normal. These years are related to a late northward migration of the ITCZ, an abnormal deceleration of the trade-winds and a warming of the Gulf of Guinea SST, from July to September (Kouadio et al., 2003, 2007).

In Figure 5, ODI\_1 and ODI\_2 indices are now compared with the referenced ODIs, calculated by Louvet (2008) (noted ODI\_REF1) and Brandt et al. (2011) (noted ODI\_REF2). A positive (respectively negative) difference between the simulated ODIs and references indicates that the onset date of the WAM in the simulation occurs after (respectively before) the references. ODI\_1 and ODI\_2 have similar differences with respect to the two references. It implies that the monsoon onset does not depend on the SST dataset, used to force the model. This panel also shows a mean bias error of 5-day lag for the difference between each simulated ODI and the two references. The mean difference with ODI\_REF2 is higher (~5 days) than that with ODI\_REF1 (~ 0 day). That could probably be due to the different precipitation datasets used to calculate the two reference ODIs. When considering the difference with ODI\_REF1 (differences with ODI\_REF1 and ODI\_REF2 have the same evolution), the great differences are 14 days from 1987 to 1988, 10 days in 1991, 21 days in 1997 and, finally, 12 days in 1998, which represent El-Niño periods. For example in 1991, the ODI\_REF1 WAM onset date is 7<sup>th</sup> of July, while it is 26<sup>th</sup> of June for ODI\_1 and 27<sup>th</sup> of



**Figure 4.** Inter-annual evolutions both of the ODI\_1 standardized departure from the mean date (curve) and of the standardized daily departure, during July, from the smoothed difference between the northern (7.5°N - 15°N) and the southern (1°N - 7.5°N)-(histogram).

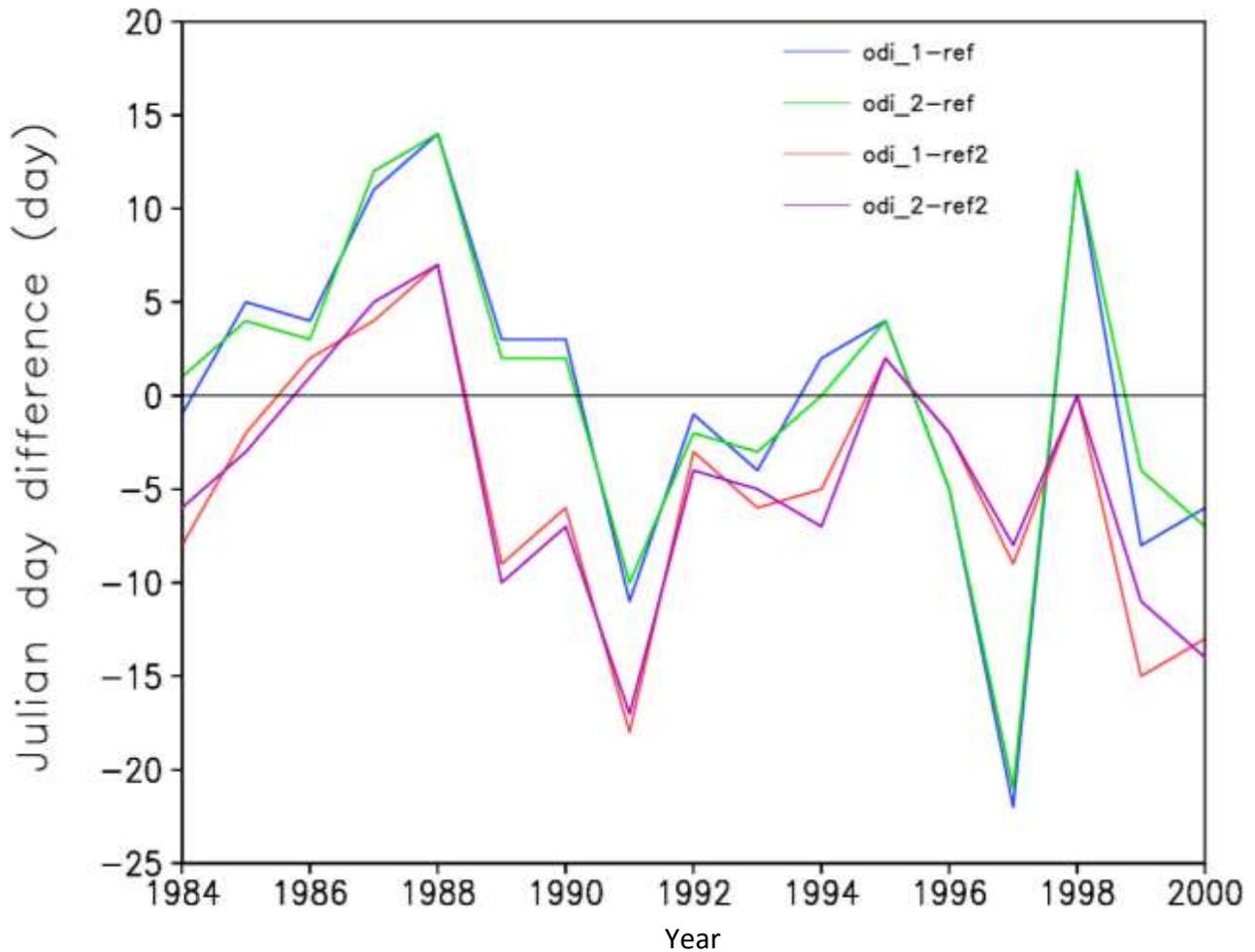
June for ODI\_2.

**Influence of ACT on ODI**

The ACT onset dates are defined as the dates for which the SST in the equatorial Atlantic falls below 26°C within 20°W - 8°E and 5°N - 5°S. This temperature threshold has also been used to define the equatorial and coastal upwellings in the Gulf of Guinea (Arfi et al., 1991; Ali et al., 2011). In this study, both the interpolated daily 2DvarSST and Olv1SST are spatially averaged using all SST values, including the upwelling temperature threshold. Then, the ACT onset date is determined by applying a 10-day moving average on these series as in the ODI cases. The ACT onset dates are noted ACT\_1, ACT\_2 and ACT\_REF2, respectively, for WAFSSTERA, WAFSSTREY and the referenced ACT onset date of Brandt et al. (2011) on 30°W - 12°E and 5°N - 5°S window.

Figure 6a shows the inter-annual evolution of ODI\_1

and ACT\_1 (as ODI\_1 and ODI\_2 give a similar result) and between ODI\_REF2 and ACT\_REF2 (Figure 6b) (Brandt et al., 2011). Evolutions of ODI\_1 and ACT\_1 are in agreement, since their correlation coefficient is about 0.69 and statistically significant at the 95% confidence level. For ODI\_REF2 and ACT\_REF2 in 1984 to 2000 period, the correlation reaches a no-significant coefficient of about 0.26. However, Brandt et al. (2011) found a significant value at 97% confidence level for the correlation between these two last variables in 1974 to 2004 periods. The weak ODI\_REF2-/ACT\_REF2 correlation, found in the calculations, could be due to (i) the reduced period, compared to that in Brandt et al. (2011) and (ii) the years that follow El-Niño periods, as for instance, 1986 to 1987 and 1991. Another remark is that high negative values up to 7 days are found in 1987 (~-7 days), in 1995 (~-14 days) and in 1996 (-21 days) for MAR simulations. One could suppose that for these years the onset of the WAM occurs before the ACT onset. Generally, the WAM onset begins after the ACT onset, as shown by the positive differences. However, if one



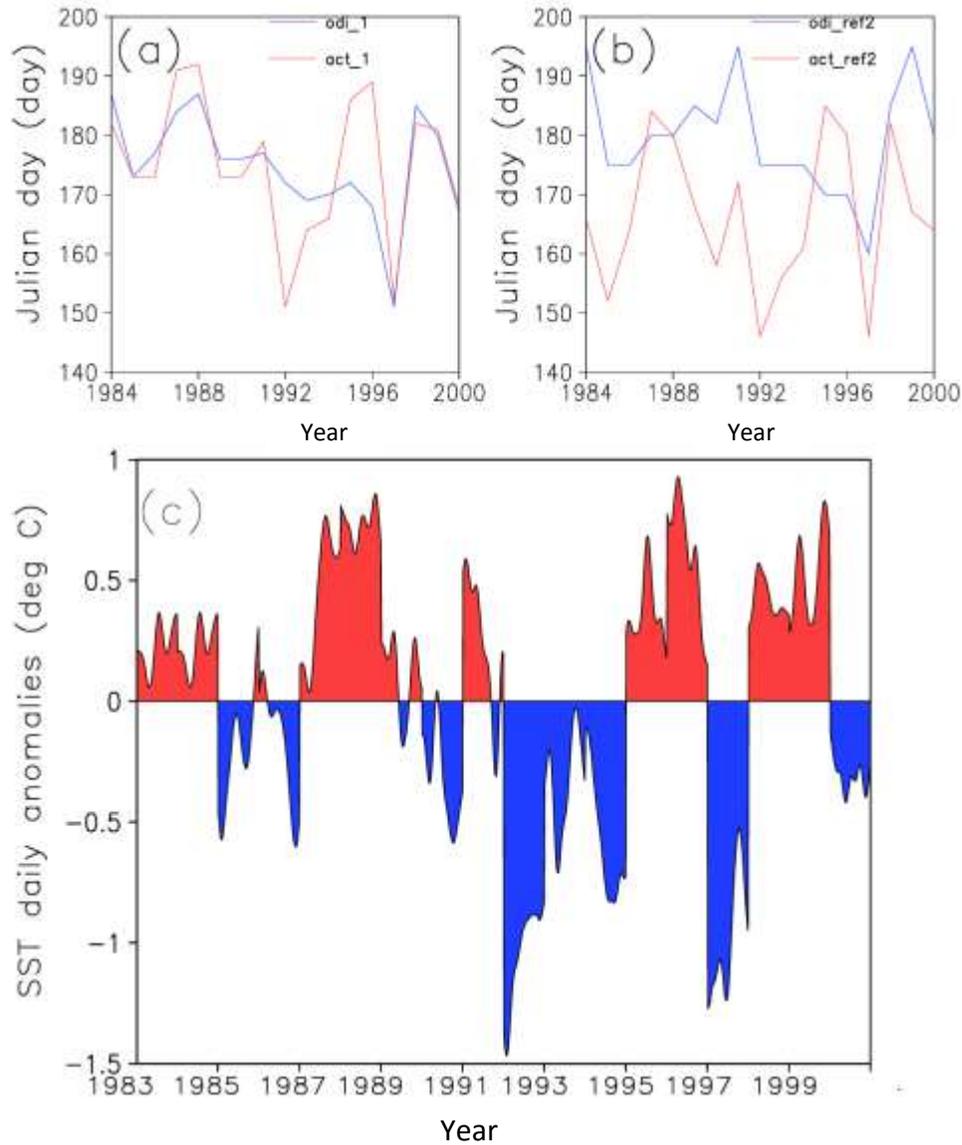
**Figure 5.** Difference between onset date indexes simulated by using 2DvarSST (noted ODI\_1) and Olv1SST (noted ODI\_2) and the referenced onset date indexes of Louvet (2008) (noted ODI\_REF1) and Brandt et al. (2011) (noted ODI\_REF2).

considers the daily departure of WAFSSTERA in the region B (Figure 6c), SST remained warm in these three years. Such oceanic conditions did not favour the rapid establishment of the upwelling (a SST below 26°C) and, then, lead to a negative difference between ODI\_1 and ACT\_1.

**Oceanic-atmospheric conditions during particular years**

Here outlines the oceanic-atmospheric conditions occurring during 1984, 1987, 1995 and 1996. During 1987, 1995 and 1996, high negative differences up to 7 days are observed between the WAM onset and ACT onset because of a high SST value. The year 1984 has late onset and documents, as well as 1987, a wide flood on the coast of the Gulf of Guinea (Kouadio et al., 2003), while 1987, 1995 and 1996 have early onsets. Since the monsoon onset does not depend on the SST and,

consequently, on dataset used to force the model, WAFSSTERA is then used to simulate the meridional heat transport (MHT). For each of the above-mentioned years, the MHT simulated by the model is longitudinally averaged between 10°W and 10°E and, then, vertically averaged in the monsoon column from 1000 to 850 hPa. The MHT is calculated using the simulated southern wind multiplied by the specific humidity. Then, anomalies have been found between each daily value of MHT and the mean year. Positive (respectively negative) anomalies could mean excess (respectively deficit) of a northward (respectively southward) moisture transport, depending on the sign of the southern wind. The time-latitude diagrams of MHT anomaly (Figure 7) show opposite patterns during the year. When positive MHT anomalies are found in the north, negative values are noticed in the south. For example, during March-October period, positive anomalies pattern of MHT extends southward from 7.5°N up to 2°S. Positive anomalies are observed in boreal summer, while negative ones extend northward

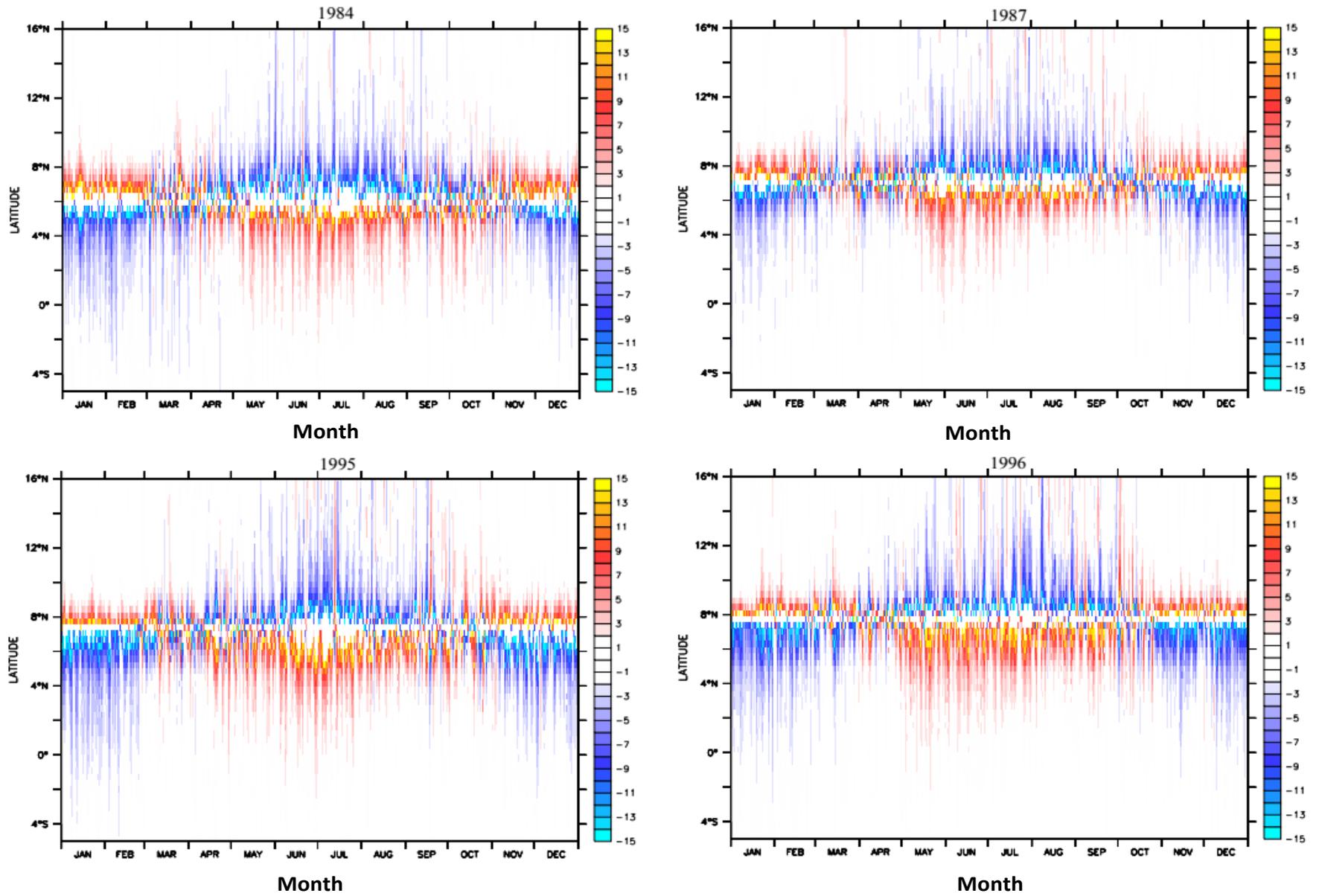


**Figure 6.** Inter-annual evolution of (a) the difference between Julian days of ODI\_1 and ACT\_1 (blue colour), and (b) between ODI\_REF2 and ACT\_REF2 (red colour); (c) Daily anomalies of WAFSSTERA within region B (shown in Figure 1), calculated as the mean departure from the 1983 - 2000 daily trend.

from 7.5°N to 15°N. Such contrast occurs in boreal summer when much moisture is carried from the ocean to the continent. In November - February period, no moisture is transported from the ocean, but positive anomalies are observed between 7.5°N and 9°N. This band could represent the limit of the humid zone in West Africa and, then, the influence of this forest region on the climate.

The panels also show the yearly succession of negative and positive anomalies around a mean limit that reaches approximately 7.5°N. This limit corresponds to the mean position between the two equilibrium latitudes (5°N and 10°N), associated with the major rainy season

of the Guinean and Sudano-Sahelian zones previously mentioned. For example, in 1984, 1987, 1995 and 1996, the limit between positive and negative anomalies of MHT is located at 6°N, 7°N, 7.5°N and 8°N, respectively. These different positions could signify different ways of the monsoon penetration into the continent. When the limit is southward of 7.5°N, the moisture does not reach the north, but is concentrated at the south. That is the case, for instance, of 1984 and 1987, during which, in boreal summer, the West African littoral experienced floods. When the limit reached or has been above 7.5°N, both north and south of West Africa experienced heavy rainfall, since the moisture covered these two zones.



**Figure 7.** Time-latitude diagrams of the simulated Meridional heat transport (MHT) anomalies, performed between each daily value of MHT and the mean year. MHT is first averaged between 10°W and 10°E in the monsoon column from 1000 to 850 hPa during (a) 1984 (b) 1987 (c) 1995 and (d) 1996 before anomalies are performed.

## Conclusions

This work (i) simulates the relationship between occurrences of the Atlantic Cold Tongue (ACT) and the West African Monsoon (WAM) onsets in the Sudano-Sahelian region using the Regional Atmospheric Model (MAR) and (ii) investigates the influence of both ERA-40 (2DvarSST) and Reynolds (Olv1SST) SST data on regional climate simulations over the Gulf of Guinea. To undertake this work, two simulations (WAFSSTERA and WAFSSTREY) experiments are performed using, respectively, SSTs forcing from ERA (2DvarSST) and Reynolds (Olv1SST), during the 1983-2000 period in the domain 23°W - 15°E and 7.2°S - 15°N. The independent experiments begin on 1st of January, 1983 and end on 31st of December, 2000. The year 1983 is considered as a spin-up time for the Soil Vegetation Atmosphere Transfers model coupled with MAR.

The simulation results show that the MAR model reproduces high values of the rainfall over the Guinean zone even if the amount is slightly underestimated when compared to the CRU data. In the tropical ocean, the MAR model shows a band of weak precipitation (< 1 mm/day) extending from 20°W to 4°W at the equator with WAFSSTREY; while another one is confined between 18°W and 10°W with WAFSSTERA simulation. These two zones are included in the equatorial band and correspond to the ACT area. WAFSSTERA gives more annual daily average rainfall (~1.6 mm/day) than WAFSSTREY in the Gulf of Guinea. In contrast, more rainfalls in WAFSSTREY simulation (~1.6 mm/day) are observed near the Congo River discharge and at the Senegalese upwelling zone. To analyze the relationships between onsets of the WAM and ACT, an onset date index (ODI) is computed, firstly by calculating a 10-day moving average of the difference between the mean daily rainfall at the north and south of 7.5°N. ODI is considered as the Julian date index after which the smoothed index has successive positive values for at least 20 days. This is realized to ensure the detection of a strong and robust climate signal like the “monsoon jump”, and not the beginning of the first Sudano-Sahelian rains. The ODIs are noted as ODI\_1 and ODI\_2 for the results, which were obtained, respectively, with WAFSSTERA and WAFSSTREY. A more northward penetration of the monsoon flow is noticed when ODI\_1 or ODI\_2 is observed early for particular years. This is the case of 1987, which is known to be a flood year at the coast of the Gulf of Guinea. This year is related to late northward migration of the ITCZ, an abnormal deceleration of the trade-winds and, finally, a warming of SST in the Gulf of Guinea from July to September.

The comparison between simulated ODIs (ODI\_1 and ODI\_2) and those calculated by Louvet (2008) and Brandt et al. (2011) are quite similar and means that the monsoon onset does not depend on the SST and, consequently, on dataset used to force the model.

Moreover, the difference between the ODIs and ACT

onset date shows high negative values up to 7 days in 1987 (-7 days), in 1995 (-14 days) and in 1996 (-21 days). This is due to high SST values in these three years. Such oceanic conditions did not favour the rapid establishment of the upwelling and, then, leads to a negative difference. It is also corroborated by the yearly evolution of the meridional heat transport (MHT) anomaly in 1984, 1987, 1995 and 1996. The band (7.5°N - 9°N) of positive MHT anomalies could represent the limit of the humid zone in West Africa, and then, the influence of this forest region on the climate. Moreover, the yearly succession of negative and positive anomalies are observed around a mean limit 7.5°N, which corresponds to the mean position between the two equilibrium latitudes (5°N and 10°N), associated with the Guinean and Sudano-Sahelian major rainy seasons. It has been found that, when the limit is southward of 7.5°N, the moisture does not reach the north, but is concentrated at the south, as for instance in 1984 and in 1987. When the limit reached or has been above 7.5°N, both north and south of West Africa experienced heavy rainfall since the moisture covered these two zones.

The present simulation opens opportunities to study the influence of different SST datasets on climate simulations and analyze the relationship between occurrences of the ACT and WAM onsets. That could be a valuable tool to study different scenarios of ACT and the mechanisms from which it could influence the climate change impacts on the West African countries.

## CONFLICT OF INTERESTS

The authors have not declared any conflict of interests.

## REFERENCES

- Ali KE, Kouadio KY, Zahiri E-P, Aman A, Assamoi AP, Bourlès B (2011). Influence of the Gulf of Guinea coastal and equatorial upwellings on the precipitations along its northern coasts during the boreal summer period. *Asian J. Appl. Sci.* 4(3):271-285.
- Arfi R, Pezennec O, Cissoko S, Mensah M (1991). Variations spatiale et temporelle de la résurgence ivoiro-ghanéenne. In *Pêcheries Ouest-Africaine: Variabilité, Instabilité et Changement*, P. Cury and C. Roy, Eds., Orstom, Paris, France, pp. 162-172.
- Bechtold P, Bazile E, Guichard F, Mascart P, Richard E (2001). A mass flux convection scheme for regional and global models. *Quart. J. Roy. Meteor. Soc.* 127:869-886.
- Brandt P, Caniaux G, Bourlès B, Lazar A, Dengler M, Funk A, Hormann V, Giordani H, Marin F (2011). Equatorial upper-ocean dynamics and their interaction with the West African monsoon. *Atmos. Sci. Lett.* 12:24-30.
- Caniaux G, Giordani H, Redelsperger J-L, Guichard F, Key E, Wade M (2011). Coupling between the Atlantic cold tongue and the West African monsoon in boreal spring and summer. *J. Geophys. Res.* 116(C04003).
- Chang R, Zhang R, Hazeleger W, Wen C, Wan X, Ji L, Haarsma RJ, Breugem W-P, Seidel H (2008). Oceanic link between abrupt changes in the North Atlantic Ocean and the African monsoon. *Nat. Geosci.* 1:444-448.
- Cook KH (1994). Mechanism by which surface drying perturbs tropical precipitation fields. *J. Climate* 7:400-413.

- Da-Allada CY, Alory G, Du Penhoat Y, Kestenare E, Durand F, Hounkonnou NM (2013). Seasonal mixed-layer salinity balance in the tropical Atlantic Ocean: Mean state and seasonal cycle. *J. Geophys. Res. Oceans* 118:332-345.
- De Ridder K, Gallée H (1998). Land surface induced regional climate change in Southern Israel. *J. Appl. Meteor.* 37:1470-1485.
- Diaconescu EP, Gachon P, Scinocca J, Laprise R (2015). Evaluation of daily precipitation statistics and monsoon onset/retreat over western Sahel in multiple data sets. *Clim. Dyn.* 45:1325-1354.
- Eltahir EAB, Gong C (1996). Dynamics of the wet and dry years in West Africa. *J. Clim.* 9:1030-1042.
- Folland CK, Palmer TN, Parker DE (1986). Sahel rainfall and worldwide sea temperatures. *Nature* 320:602-607.
- Fontaine B, Louvet S (2006). Sudan-Sahel rainfall onset: Definition of an objective index, types of years, and experimental hindcasts. *J. Geophys. Res.* 111:(D20103).
- Fouquart Y, Bonnel B (1980). Computation of solar heating of the Earth's atmosphere: A new parameterization. *Beitr. Phys. Atmos.* 53:35-62.
- Gallée H (1995). Simulation of the meso-cyclonic activity in the Ross Sea, Antarctica. *Monthly Weather Rev.* 123:2051-2069.
- Gallée H, Schayes G (1994). Development of the three-dimensional meso- $\gamma$  primitive equations model: katabatic wind simulation in the area of terra nova bay, Antarctica. *Monthly Weather Rev.* 122:671-685.
- Gallée H, Moufouma OW, Bechtold P, Brasseur O, Dupays I, Marbaix P, Messenger C, Ramel R, Lebel T (2004). A high resolution simulation of a West African rainy season using a regional climate model. *J. Geophys. Res.* 109(D05 108).
- Giannini A, Saravanan R, Chang P (2003). Oceanic forcing of Sahel rainfall on inter-annual to inter-decadal time scales. *Sci.* 302:1027-1030.
- Janicot S, Moron V, Fontaine B (1996). Sahel drought and ENSO dynamics. *Geophys. Res. Lett.* 23:515-518.
- Kessler E (1969). On the distribution and continuity of water substances in atmospheric circulation. *Meteor. Monogr. Am. Meteor. Soc.* 10(32):84.
- Kouadio KY, Ali KE, Zahiri E-P, Assamoi AP (2007). Etude de la prédictibilité de la pluviométrie en Côte d'Ivoire Durant la période de juillet à septembre. *Rev. Ivoir. Sci. Technol.* 10(5):117-134.
- Kouadio KY, Aman A, Ochou DA, Ali KE, Assamoi AP (2011). Rainfall variability patterns in West Africa: Case of Côte d'Ivoire and Ghana. *J. Environ. Sci. Eng.* 5:1229-1238.
- Kouadio KY, Ochou DA, Servain J (2003). Tropical Atlantic and the rainfall variability in Côte d'Ivoire. *Geophys. Res. Lett.* 30(5).
- Kouassi B, Diawara A, Kouadio KY, Schayes G, Yoroba F, Kouassi AA, Zahiri E-P, Assamoi P (2012). Numerical Study of a West African Squall Line Using a Regional Climate Model. *Atmos. Clim. Sci.* 2:14-22.
- Lamb P (1978). Large-scale tropical Atlantic surface circulation patterns associated with sub-Saharan anomalies. *Tellus* 30:240-251.
- Louvet S (2008). Modulations intra-saisonnnières de la mousson d'Afrique de l'ouest et impacts sur les vecteurs du paludisme à Ndiap (Sénégal): Diagnostics et prévisibilité. Thèse de l'Université de Bourgogne.
- Messenger C, Gallée H, Brasseur O (2004). Precipitation sensitivity to regional SST in a regional climate simulation during the West African monsoon for two dry years. *Clim. Dyn.* 22:249-266.
- Morcrette JJ (1984). Sur la paramétrisation du rayonnement dans les modèles de la circulation générale atmosphérique. Ph.D. thesis, Université des sciences et techniques de Lille. P 373.
- Moufouma-Okia W (2003). Modélisation Climatique de l'Afrique de l'Ouest avec le Modèle Regional MAR. Ph.D. Thesis, Institut National Polytechnique de Grenoble, Grenoble.
- New M, Hulme M, Jones P (2000). Representing twentieth century space-time climate variability. Part II: development of a 1901-1996 monthly grids of terrestrial surface climate. *J. Clim.* 13:2217-2238.
- Nicholson SE (1989). African drought: characteristics, casual theories and global connections. Understanding climate change. *Geophys. Monogr.* 52:79-100.
- Nicholson SE (2013). The West African Sahel: A review of recent studies on the rainfall regime and its inter-annual variability. *ISRN Meteorol.* 2013:32 <http://dx.doi.org/10.1155/2013/453521>.
- Okumura Y, Xie S-P (2004). Interaction of the Atlantic equatorial cold tongue and the African monsoon. *J. Clim.* 17:3589-3602.
- Poccard I, Janicot S (2004). Comparison of rainfall structures between NCEP/NCAR reanalyses and observed data over tropical Africa. *Clim. Dyn.* 16:897-915.
- Ramel R (2005). Surface processes impacts on West African climate. Ph.D. Thesis, Université Joseph Fourier de Grenoble, France, P 149.
- Reynolds R, Smith TM (1994). Improved global sea surface temperature analyses using optimum interpolation. *J. Clim.* 7:929-948.
- Rowell DP, Folland CK, Maskell K, Owen JA, Ward MN (1992). Modelling the influence of global sea surface temperatures on the variability and predictability of seasonal Sahel rainfall. *Geophys. Res. Lett.* 19:905-908.
- Servain J, Picaut J, Busalachi A (1985). Inter-annual and seasonal variability of the tropical Atlantic Ocean depicted by sixteen years of sea surface temperature and wind stress. *Elsevier Oceanogr. Series* 40:211-237.
- Sultan B, Janicot S (2000). Abrupt shift of the ITCZ over West Africa and intra-seasonal variability. *Geophys. Res. Lett.* 27(20):3353-3356.
- Vanvyve E, Hall N, Messenger C, Leroux S, Van Ypersele JP (2008). Internal variability in a regional climate model over West Africa. *Clim. Dyn.* 30:191-202.
- Sultan B, Janicot S, Diedhiou A (2003). The West African monsoon dynamics. Part I: Documentation of intraseasonal variability. *J. Clim.* 16:3407-3427.
- Vellinga M, Arribas A, Graham R (2013). Seasonal forecasts for regional onset of the West African monsoon. *Climate Dyn.* 40:3047-3070.
- Vizy EK, Cook KH (2001). Mechanisms by which Gulf of Guinea and eastern North Atlantic Sea surface temperature anomalies can influence African rainfall. *J. Climate* 14:795-821.
- Yoroba F (2012). Schémas convectifs dans le modèle climatique régional MAR et vraisemblance des champs de précipitations ouest-africains. PhDThesis, Université Félix Houphouët-Boigny de Cocody-Abidjan (Côte d'Ivoire), P 170.
- Yoroba F, Diawara A, Kouadio KY, Schayes G, Assamoi AP, Kouassi KB, Kouassi AA, Toualy E (2011). Analysis of the West African rainfall using a Regional Climate Model. *Int. J. Environ. Sci.* 1(6):1339-1349.