Asian Journal of Applied Sciences 4 (3): 271-285, 2011
ISSN 1996-3343 / DOI: 10.3923/ajaps.2011.271.285

Influence of the Gulf of Guinea Coastal and Equatorial Upwellings on the Precipitations along its Northern Coasts during the Boreal Summer Period

[1]K.E. Ali, [1]K.Y. Kouadio, [1]E.-P. Zahiri, [1]A. Aman, [1]A.P. Assamoi and [2]B. Bourles
[1]LAPAMF, Universite de Cocody-Abidjan BP 231 Abidjan, Cote d'Ivoire
[2]IRD/LEGOS-CRHOB, Representation IRD, 08 BP 841 Cotonou, Republique du Benin

Corresponding Author: Angora Aman, Universite de Cocody-Abidjan 22 BP 582 Abidjan 22, LAPA-MF, UFR-SSMT, Cote d'Ivoire Tel: (225) 07 82 77 52 Fax: (225) 22 44 14 07

ABSTRACT
The Gulf of Guinea (GG) is an area where a seasonal upwelling takes place, along the equator and its northern coasts between Benin and Cote d'Ivoire. The coastal upwelling has a real impact on the local yet documented biological resources. However, climatic impact studies of this seasonal upwelling are paradoxically very rare and disseminated and this impact is still little known, especially on the potential part played by the upwelling onset on the regional precipitation in early boreal summer. This study shows that coastal precipitations of the July-September period are correlated by both the coastal and equatorial sea-surface temperatures (SSTS). This correlation results in a decrease or a rise of rainfall when the SSTs are abnormally cold or warm respectively. The coastal areas that are more subject to coastal and equatorial SSTs influence are located around the Cape Three Points, where the coastal upwelling exhibits the maximum of amplitude.

Key words: West Africa, guinea gulf, sea surface temperature, precipitation, coastal upwelling

INTRODUCTION
Lamb (1978), Hastenrath (1984), Lough (1986) and Wagner and da-Silva (1994) are among the first scientists who suggested that Sea-Surface Temperature (SST) anomalies in the tropical Atlantic are linked to the deficit in Sahelian rainfall in West Africa. More recent studies have confirmed these results (Janicot *et al.*, 1998; Fontaine *et al.*, 1999) demonstrating that, when SSTs are high in the Tropical Atlantic, Sahel precipitations decrease, whereas coastal precipitations increase. Lamb and Peppler (1992) also insisted on the key role played by SSTs in the rainfall variability of West Africa at interannual and decadal time-scale. Recently, Gu and Adler (2004), by using Tropical Rainfall Measuring Mission (TRMM) satellite data, demonstrated that the rainfall peak recorded in May along the coastal areas of the Gulf of Guinea (hereafter GG), was mainly modulated by the seasonal forcing of the ocean. Eltahir and Gong (1996) explained that the intensity of West African monsoon and the northward penetration of the rain band in boreal summer depend on the meridional gradient of the static and humid energy in the boundary layer between the ocean and the continent. That suggests a significant influence of the tropical Atlantic SSTs on the GG regional climate. More local studies confirmed the key role of the SSTs on the rainfall variability in West Africa. Opoku-Ankomah and Cordery (1994) showed a possible strong correlation between SSTs in the Tropical Atlantic and the rainfall in Ghana. In the same way,

Kouadio *et al.* (2002), whose study was based on statistical correlations between oceanic SSTs and precipitations, showed the impact of the tropical Atlantic SST on the precipitation in Cote d'Ivoire. SST thus appears as a key parameter for the modulation of the variability of precipitation. Previous theoretical, modelling and observational analyses suggested that intraseasonal SST variability provides important feedbacks onto atmospheric convection. For example, by using idealized column models, Sobel and Gildor (2003) and Maloney and Sobel (2004) suggest that atmosphere-ocean coupling can generate strong, local intraseasonal oscillation having a discharge/charge character. Thus, the SST variability is strongly linked with rainfall anomalies over the African continent. SSTs are also linked with the position and the latitudinal extension of the West Africa monsoon system (Vizy and Cook, 2001). Various diagnostic and sensitivity studies show that during the second half of the 20th century, the West African mean sources of predictability for the Sahelian rainy seasons are from the tropical eastern Pacific (negative correlation) and from meridional gradients of the SST anomaly in the tropical basins of the Atlantic and other oceans. In the case of the tropical Atlantic basin, correlations are positive (negative) with the North (Equatorial and South) basins (Folland *et al.*, 1986, 1991; Moron *et al.*, 1995; Rowell *et al.*, 1995; Janicot *et al.*, 1996; Fontaine *et al.*, 1998; Camberlin *et al.*, 2001). In the same way, the Guinean precipitations are mainly associated with the South and Equatorial Atlantic basins in boreal summer. Quite often a warmer basin leads in the disappearance of the August short dry season (Wagner and da-Silva, 1994). The second September-October rainy season is also mainly linked to the tropical Atlantic: warmer basin is associated with positive rainfall anomalies (Camberlin *et al.*, 2001).

In the northern part of the GG, the circulation is dominated by the eastward flowing Guinea Current, which is sometimes considered as the extension of the North Equatorial Counter Current in boreal summer. However, the circulation in the eastern part of the GG is complex, highly variable and not well documented (Bourles *et al.*, 2002). Presently, the processes that modulate the SSTs in this area and the relative role of oceanic processes and air-sea fluxes for the SST variability, are very complex and still remain not well understood (Foltz *et al.*, 2003; Bourras *et al.*, 2009; Marin *et al.*, 2009).

Another significant question is the effect of the coastal and equatorial upwellings on the West African monsoon. If the upwellings vary primarily seasonally (Gouriou and Reverdin, 1992), their existence and their amplitude are also variable from one year to another, partly according to the equatorial mode of variability that, along with the meridional mode, controls the climate and the SSTs in the tropical Atlantic (Servain *et al.*, 2000). Such SST variability is consequently likely to influence the onset and the intensity of the monsoon observed in West Africa (Fontaine *et al.*, 1999). For example, the seasonal SST cooling of the boreal summer established between Cote d'Ivoire and Benin tends to increase the meridional gradient of the SSTs in the GG. However, all these previous studies are either too focalized on reduced area or mostly devoted to Sahelian regions. Furthermore, they do not clearly address the relative role of coastal and equatorial upwellings on precipitations over the Northern coastal areas of the GG.

In this study, we both examine and try to determine how coastal and equatorial SSTs may influence the rainfall along the GG coastline in boreal summer (i.e., during the annual period of the oceanic major equatorial and coastal upwellings and African monsoon), in order to understand their potential impact on the rainfall variability during that period.

MATERIALS AND METHODS

Two types of data have been used in this study: monthly SST data from 1998 to 2007, retrieved from the TRMM Microwave Imager (hereafter TMI) (www.ssmi.com) on board the TRMM satellite and monthly Global Precipitation Climatology Center (GPCC) data during the same ten years.

Sea surface temperature and precipitation: Monthly averaged gridded TMI SST data were obtained from Remote Sensing Systems (http://www.ssmi.com) for the 1998-2007 periods. The data are provided on a 0.25°×0.25° grid and cover a global region extended from 40°S to 40°N. With a distinct advantage over the traditional Infrared (IR) products that require a cloud-free field of view, TMI can estimate SSTs in cloudy condition. It provides a continuous coverage of tropical SSTs (Wentz *et al*., 2000) and thus TMI SST products are better suited for studying SSTs in regions with strong atmospheric convection (Maloney *et al*., 2008).

Furthermore, the high spatial resolution of the TMI data provides a realistic product to assess the SST evolution, particularly in the case of a local study in a region marked by a strong spatio-temporal variability of the SST. Some studies have compared TMI products to other ones and have shown the quality of TMI data for regional and local analysis (e.g., Sengupta *et al*., 2001; Bhat *et al*., 2004). For our statistical study, we use the 1998-2007 period of TMI data, corresponding to a 10 years period during which they are available and the best possible..

In order to analyze the possible impact of the SST anomalies on the continental precipitations, we use the monthly precipitation fields from Global Precipitation Project (GPCC) data set (Rudolf *et al*., 1994) extracted on a 1°×1° regular grid, also for the 1998-2007 periods. This data set is based on quality-controlled data from 7000-8000 stations. The product is optimized for best spatial coverage and use for water budget studies.

Methodology: SST data are used for the April-September period and precipitation data for the July-September period, both during 10 years (from 1998 to 2007). For a suitable comparison between SST and precipitation data sets, we first processed to calculate standardized anomalies as the ratio between the monthly anomalies and the corresponding standard deviation, in order to eliminate some seasonality effects in the data. It has to be kept in mind that a negative anomaly of SSTs corresponds to oceanic cooling whereas a positive anomaly is the signature of oceanic warming. For the precipitations, a negative (positive) anomaly indicates a rainfall decrease (increase), respectively.

Precipitations and SST statistical links have been checked as follow: (1) for a selected oceanic zone, the correlation between the SST and the precipitation standardized anomalies are computed during the upwelling period, i.e., from July to September (hereafter JAS for July-August-September). Oceanic zones include the coastal zone (8°W-5°E; 4°N-5°N) and the equatorial zone (12°W-0°E; 0°-2°S). The selected oceanic zones correspond to regions where SST variability (seasonal and interannual) is the highest observed in Equatorial Atlantic Ocean (Picaut, 1983; Servain *et al*., 1985). The equatorial zone is taken in the 0°N - 2°S latitude band because the equatorial upwelling is primarily centred around 1°S (Colin, 1988; Colin *et al*., 1993). (2) Correlations have first been computed with no temporal lag (0-lag, SST: JAS). In order to evidence an eventual time delay between SSTs and precipitations, SSTs have been lagged. The performed lags are taken from one to three months as follows: 1-lag for SST in June-August (hereafter JJA for June-July-August); 2-lag for SST in May-July (hereafter MJJ); 3-lag for SST in April-June (hereafter AMJ). The real interest in taking into account temporal lags is to consider time factors in both parameters comparison procedures. This can be considered as a physical predictive parameter quest action for some given regions.

RESULTS

Figure 1 represents the monthly averaged SSTs in the GG, from January to December, calculated over ten years (1998-2007) period. Two main periods can be distinguished: one cold period, from July to October, characterized by SSTs lower than 26°C over the GG with a minimum of 23 to 24°C in August-September around the equator and along the northern coasts and one warm period, from November to May, with SSTs ranging between 26 and 30°C, from South to North of the GG, with maximum SST values in March-April. The seasonal cycle of SSTs is characterized by cooling events occurring along the northern coast (Philander, 1979; Picaut, 1983) and around the equator between 20°W and 0°E of longitude, primarily localized south of the equator. While the equatorial upwelling is principally the response to basin scale oceanic response to wind forcing (Philander and Pacanowski, 1986), the dynamics of the coastal upwelling along the northern coast of the GG are still not well understood. Several attempts to numerically reproduce this important coastal upwelling present from early July through September have been attempted, with little success (Houghton, 1976; Adamec and O'Brien, 1978). Several processes have been advanced to explain this coastal upwelling of the Gulf of Guinea. Several studies privilege local and remote actions of the wind along with the role of equatorial and coastal Kelvin waves (Picaut, 1983;

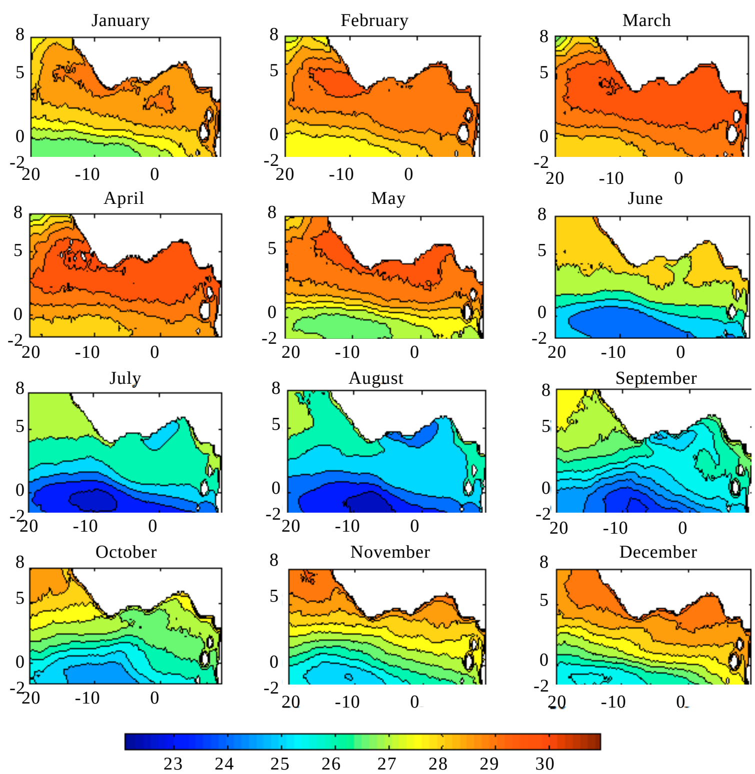

Fig. 1: Spatial evolution of monthly SSTs in the Guinea Gulf, from January to December, calculated over ten years (1998-2007) from TMI

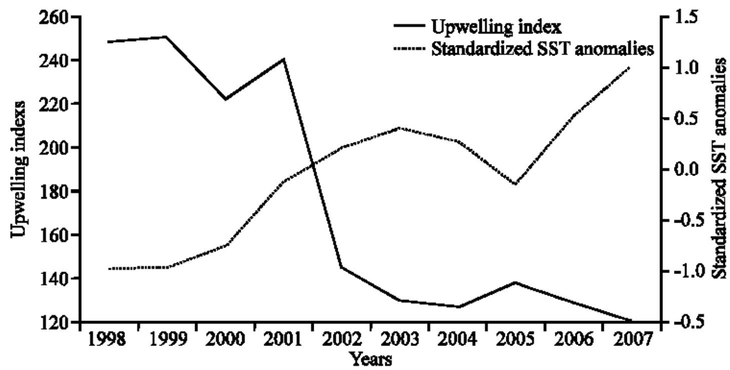

Fig. 2: Annual upwelling index and standardized SST anomalies calculated from TMI between 8°W-5°E; 4°N-5°N during 1998-2007

Moore, 1968; Lighthill, 1969; Moore *et al.*, 1978; O'Brien *et al.*, 1978; Adamec and O'Brien, 1978). Other ones put forward the potential role of the Guinea Current and cape effects (Ingham, 1970; Roy, 1995). However, the relative importance of these different processes responsible for this phenomenon still deserves to be better documented, that is beyond the scope of this study. Nevertheless, since it was shown that the sea surface temperature is a very good indicator of upwelling (Sadrinasab, 2009), the study of the SST evolution in boreal summer using weekly SSTs enabled us to compute the annual upwelling index from 1998 to 2007 between 8°W-5°E and 4°N-5°N. This index is considered as being the sum of the differences between a threshold SST of 26°C (Arfi *et al.*, 1991) and the lowest temperature of the week. This sum is multiplied by the total number of weeks of the cold season during which SSTs are lower than 26°C. The validation of the upwelling index is illustrated by the opposite evolutions of the upwelling index and the standardized SST anomalies over the JAS period: the stronger (weaker) upwelling index corresponds to the colder (hotter) SSTs (Fig. 2). The index indicates the intensity of the upwelling over the period of the calculations, as illustrated in Fig. 2 that shows a strong decrease from 2002, suggesting that the years 1998 to 2001 have been cooler than the years 2002 to 2007 during boreal summer and along the coasts. The high index from 1998 to 2001 could be explained by the important duration of the upwelling (1999, 2000, 2001) and by the high intensity of the cooling (1998). On the other hand, the low index observed from 2002 to 2007 is related with the duration of the coastal upwelling (2004 case), the weakness of the cooling (2002 case) and the combination of both parameters (2003, 2005, 2006 and 2007 cases). To illustrate this, some examples of monthly evolution of the coastal SST from July to October during 1998, 1999, 2002 and 2004 are plotted (Fig. 3). 1998 registered the lowest temperatures (SSTs < 25°C) in August while the upwelling in 1999 was prolonged until October. For the years with low upwelling index, SSTs were relatively warm during September 2004 (SSTs > 26°C), even though upwelling is generally intense in this month. That indicates a short duration of the upwelling. The cooling was weak in 2002 with SSTs above 25°C in September and August.

Figure 4 shows the evolution of the monthly climatology of the precipitation calculated over the 1998-2007 period in West Africa. The major rainy season in West Africa begins in April at approximately 5°N and reaches its maximum in June along the coasts. It moves gradually northward until approximately 15°N in July-September with a maximum in August. This northern area is characterized by only one rainy season (Sultan and Janicot, 2000). A second rainy season,

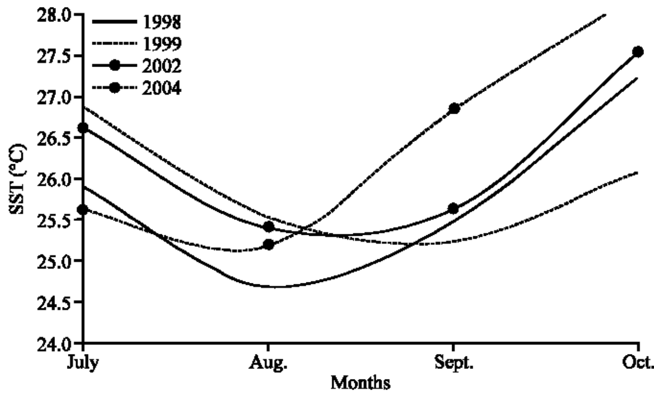

Fig. 3: Monthly evolution of the coastal SSTs from July to October during 1998, 1999, 2002 and 2004 calculated from TMI data

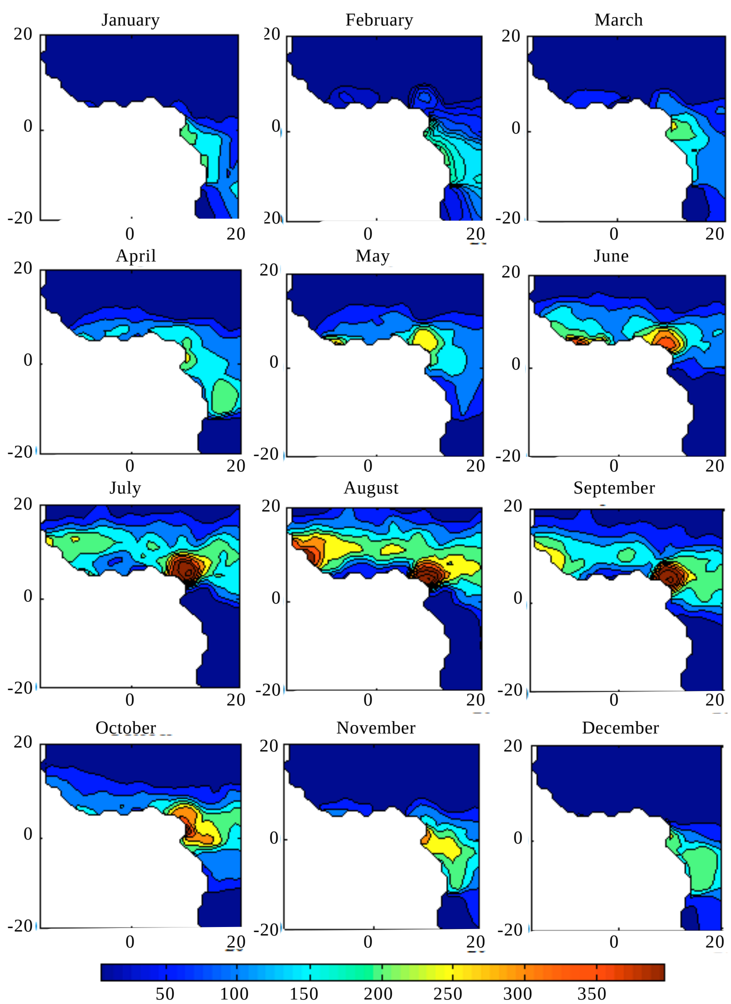

Fig. 4: Monthly means of GPCC precipitations of the Gulf of Guinea, from January to December calculated from 1998 to 2007

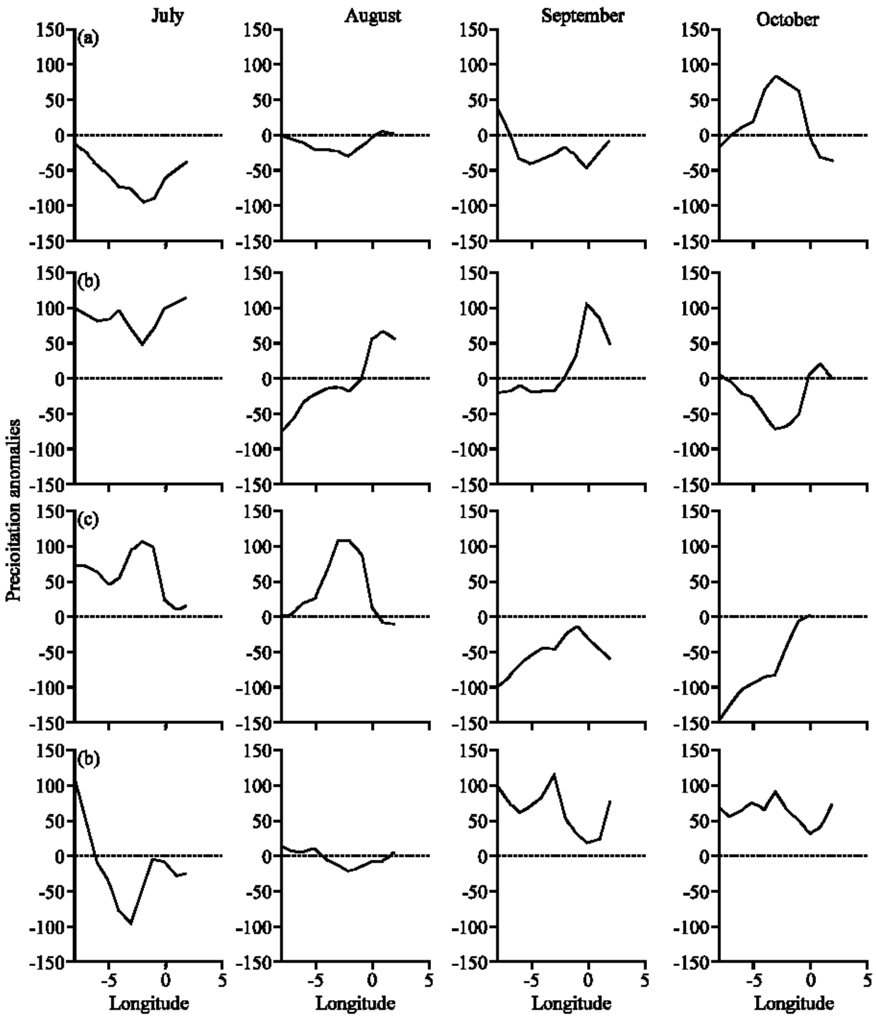

Fig. 5: Monthly precipitation anomalies from July to October along 8°W-5°E during (a) 1998, (b) 1999, (c) 2002 and (d) 2004 calculated from GPCC data

less intense than the April-June one, is observed along the coasts in October-November corresponding to the minor rainy season. This latitudinal evolution of the precipitation system is related to the oscillation of the ITCZ (Intertropical Convergence Zone) (Sultan and Janicot, 2000; Le-Barbe *et al.*, 2002). Indeed, the band of rain follows the movement of the ITCZ which itself moves according to the position of the thermal equator generally located at the latitude of the hottest air. In boreal summer, it moves northward with the top of the overheated continents to approximately 15°N. In boreal winter, it moves southward to reach approximately 5°N. The northern and the littoral regions of West Africa are also characterized by dry periods. Along the coastal regions, the major dry season begins in December and ends in March. In the Northern regions, it lasts from October to June. A minor dry season is observed along the littoral from July to September. Figure 5 presents the precipitation anomalies at 5°N along 8°W to 5°E of longitude for 1998, 1999, 2002 and 2004, i.e., the same years as for Fig. 3. It shows that in JAS, the months that sustained a relatively weak coastal cooling recorded a positive rainfall anomaly (as observed in 1999 and September of 2004). Conversely, the months during which the coastal upwelling has

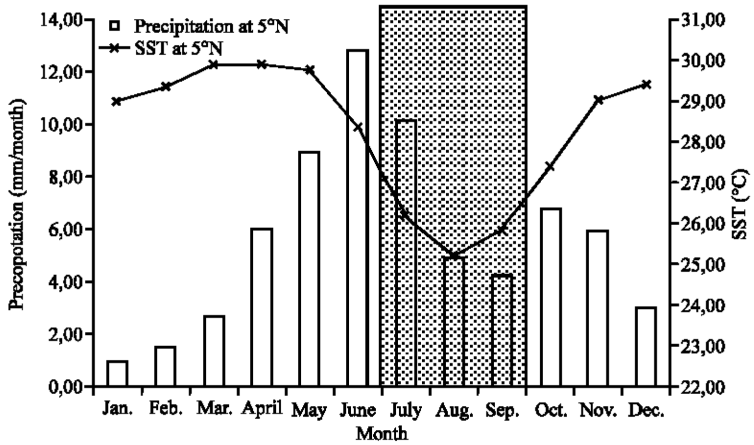

Fig. 6: Monthly mean of GPCC rain at 5°N and of the SSTs from TMI between 8°W-5°E; 4°N-5°N. The semi-transparent rectangle in gray indicates the period July-September

been intense recorded a negative rainfall anomaly (as observed in 1998). Some authors (Nicholson and Grist, 2003; Matthews, 2004) noted a link between the coastal SST and precipitations for the July-September period. Figure 6, derived from our data sets and showing the monthly mean evolution of the coastal SST and precipitation at 5°N, illustrates this conclusion. Indeed, it indicates that, at 5°N, the rains begin to decrease in August when the SST values are minimal (Fig. 2).

The following sections examine the possible links between SSTs and precipitations over West Africa. Concerning the precipitations, we focus on the continental areas of the littoral (South of 8°N and between 8°W and 5°E) and for the time period July-September which corresponds to the cooling period. The oceanic zones are defined between 8°W-5°E ; 4°N-5°N and 12°W-0°E; 0°N-2S.

Effect of the coastal SST (8°W-5°E; 4°N-5°N) on the GG coastal precipitations during the major upwelling season: Figure 7 shows the correlations between the standardized anomalies of SSTs and precipitations. At 1-lag (SST: JJA / rain: JAS; Fig. 7, left), a non-significant positive structure of correlations is observed at the coastline. These positive correlations are extended from west to east along the coastal line, though those which are recorded in the south-east of Ghana (~0.43 correlation) are at the bound of the significance. This result suggests a weak linkage between SST and precipitations with a 1 month lag. In fact June is a transition month between the oceanic warm and cold major seasons in the GG (Gallardo, 1993; Fig. 1). It is characterised by a gradual SST decrease, indicating the beginning of the coastal upwelling (Fig. 6).

At 0-lag, (SST: JAS/rain: JAS; Fig. 7, right), the regions with positive significant correlations are found along a diagonal from the west of Ghana to the South of Togo. There, the correlation coefficient is about 0.6, which indicates a simultaneous evolution of the SST and precipitations: an abnormal cooling of the SST (respectively an abnormal warming) is associated with a decrease of rainfall (respectively a rise). The SST influence on the precipitation is less spread, but rather concentrated in continental regions close to the oceanic zone where the upwelling is more intensified. However, the too localized character of the significant correlations structure may be due to a loss of information when the computations are done along the whole JAS period. For that reason, monthly correlations have been also computed.

Figure 8 shows the spatial distributions of monthly correlations between the coastal SST and the precipitations standardised anomalies without any time lag, in July, August and September. When imposing a time lag 1, whatever the month, the correlations obtained on the coastal line are not significant (not shown). July records non-significant correlation coefficients for lag 0. Positive

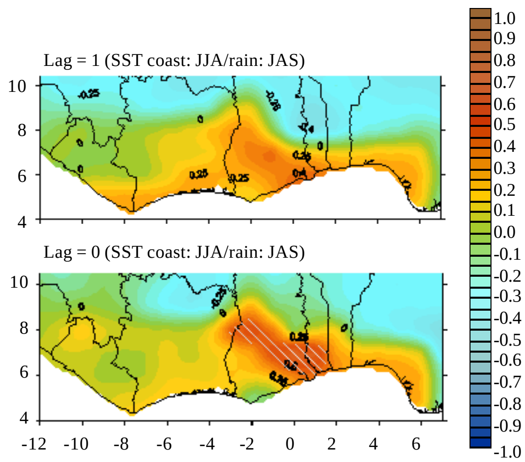

Fig. 7: Spatial distribution of correlations between coastal SSTs (8°W-5°E; 4°N-5°N) standardized anomalies and the coastline precipitations standardized anomalies during the time period July-September. Hachured zones correspond to significant correlations at 95%

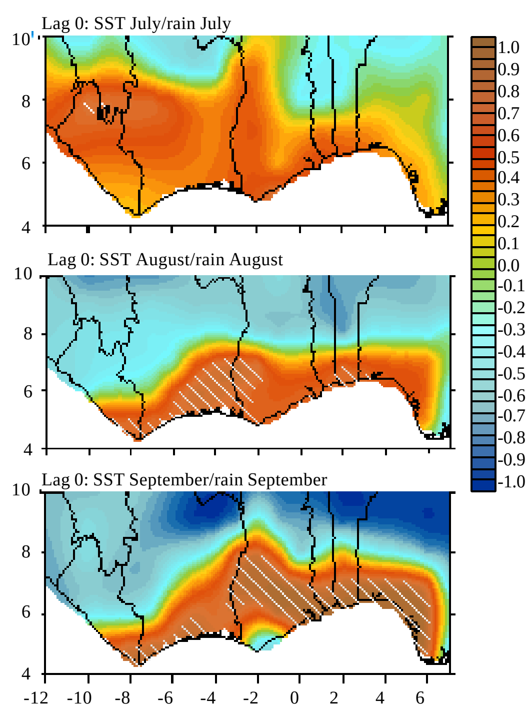

Fig. 8: Spatial distribution of monthly correlations between coastal SSTs (8°W-5°E; 4°N-5°N) standardized anomalies and the coastline precipitations standardized anomalies. Hachured zones correspond to significant correlations at 95%

correlation coefficients, even not significant in July on the coastal countries, show a weak link between a surplus (deficit) of precipitations and warm (cold) anomalies of the SSTs. On the contrary, the significant correlations registered in August and September suggests that the coastal rainfall is influenced by the coastal SSTs during these months. These positive correlations indicate a similar evolution of the SST with the coastal precipitations: a rise (respectively a decrease) of the precipitations in August or September is therefore associated with a rise (respectively a drop) of the coastal SSTs. This result complies with the works of Janicot *et al.* (1998) and Fontaine *et al.* (1999). This monthly analysis reveals that the significant correlation structure observed in Fig. 7 at lag 0 is effectively linked to a loss of information, induced by the July transition month.

Relationship between equatorial SST (12°W-0°; 0°-2°S) and coastal precipitations during the July-September period: Figure 9 shows the correlations between the standardised anomalies of equatorial SSTs and continental precipitations, with time lags of 0 to 3 months. All of them show positive correlations along the coasts, but only 3-lag correlation shows significant coefficients in the south-west of Ghana near the Cape of Three Points (3°W-1°W, 5°N-6°N). These results indicate that, at the beginning of the equatorial cooling (April-May), the equatorial SSTs and the precipitations around the Cape of Three Points are subject to a similar evolution. The equatorial upwelling starts in April-May (Colin, 1989) and is maximal during the July-September period (Fig. 1). The persistence of cold waters is supposed to contribute to the inhibition of the coastal rainfall in July-September, when precipitations are observed to be maxima farther in the north (Fig. 4). The coolness at the ocean/atmosphere interface induces the stabilization of the monsoon fluxes (Emanuel, 1985). This cooling reduces the convective movements in low troposphere and induces an inhibition effect of the coastline rainfall. As a result, the JAS precipitation near the Cape of Three Points could be predicted in accordance with the evolution of the equatorial SSTs in

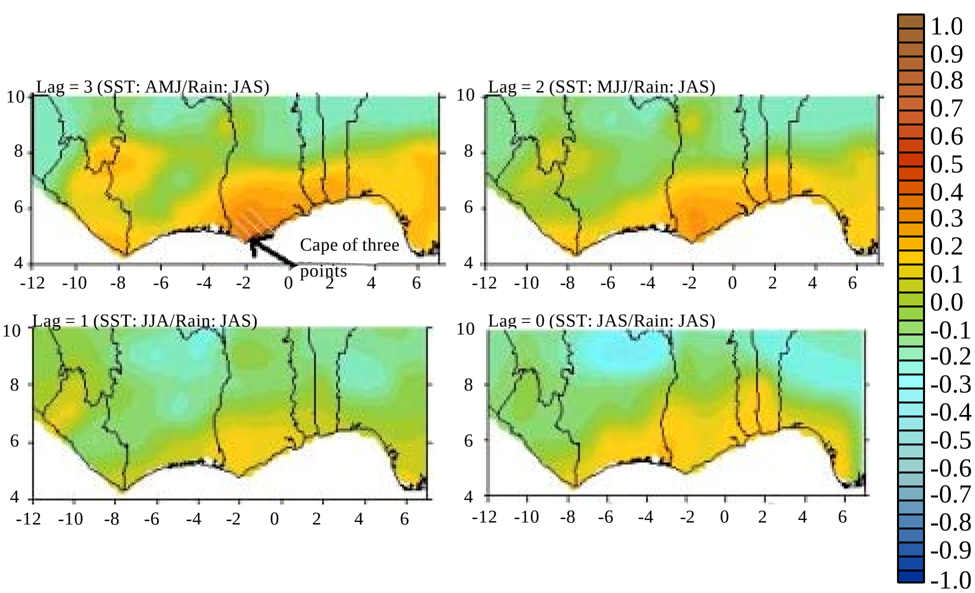

Fig. 9: Spatial distribution of correlations between equatorial SSTs (12°W-0°; 0°-2°S) standardized anomalies and the coastline precipitations standardized anomalies during the time period July-September. Hachured zones correspond to significant correlations at 95%

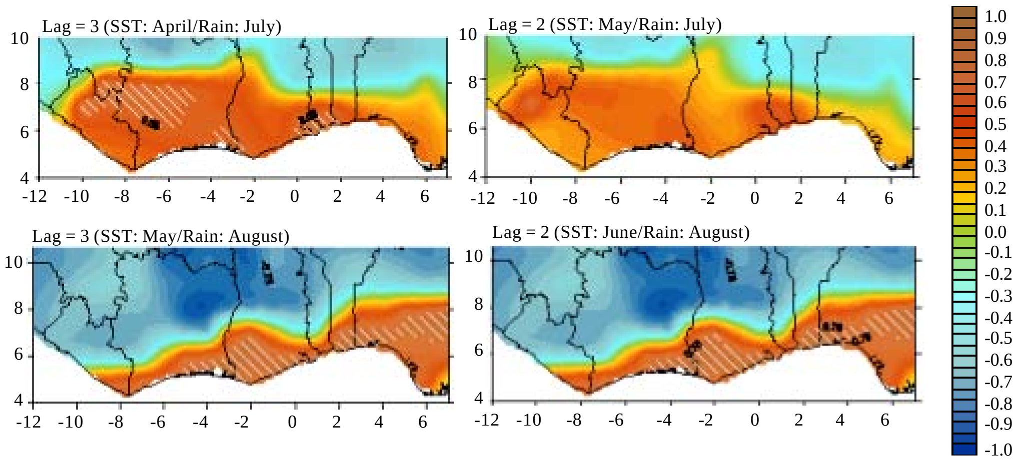

Fig. 10: Spatial distribution of monthly correlations between equatorial SSTs (12°W-0°; 0°-2°S) standardized anomalies and the coastline precipitations standardized anomalies. Hachured zones correspond to significant correlations at 95%

April-May. Any early cooling, occurring in April-May, can result in the decrease of coastal precipitations in July-September. On the other hand, any abnormal warming taking place in April-May, or a delay of the upwelling onset, can be one cause of an increase of precipitations in July-September, thus reducing the drought intensity during this time period.

The immediate action (at lag 0) of the equatorial SST anomalies has a quite null influence on the precipitation anomalies over the coastal regions whatever considered month (not shown). Figure 10 presents the monthly correlations between the standardised equatorial SST and the precipitations anomalies, with time lags of 2 and 3 months. This Fig. 10 shows an extension of positive and significant correlations in July for lag 3 (Figure 10, top-left) over the three regions delimited by the longitude / latitude ranges: 4°W-2°W / 5°N-7°N, 11°W-6°W / 6°N-8°N and 0-2°E / 6°N-7°N. In August, the 3-lag and 2-lag positive correlations are localized within the 5°N-8°N latitudinal band. In September, the correlation coefficients for lag 3 are negative and only significant in the South of Cote d'Ivoire (not shown). The observed correlations suggest that during July and August, an increase (respectively a decrease) of precipitations is related to an abnormal warming (respectively cooling) in April or May at the equator. If the equatorial cooling occurring in April or May intensifies in June, an increase of precipitations can be observed in the South of Cote d'Ivoire in September (Kouadio *et al*., 2002).

DISCUSSION

This study aims at studying the possible existing links between the oceanic surface conditions, namely the SSTs, in the Gulf of Guinea and the climatic conditions, namely the precipitations for this study, in West equatorial Africa during the boreal summer upwelling period. Previous works, as for instance Folland *et al*. (1991), Fontaine *et al*. (1998) and Camberlin *et al*. (2001), were mostly devoted to Sahelian regions. Furthermore, they do not clearly address the relative role of coastal and equatorial upwellings on precipitations over the northern coastal areas of the GG, that is the focus of the present study.

This relational study between SSTs and precipitations was realized considering two oceanic zones: the coastal upwelling zone (8°W-5°E; 4°N-5°N) and the equatorial upwelling zone (12°W-0°, 0°-2°S). Our statistical results show that the coastal SSTs have a weak influence over the coastal precipitations at the beginning of the upwelling and a more significant one during the intensification of the upwelling, i.e., in August and September. This influence results in a decrease of coastal precipitations when SST decreases in agreement with the works by Mitchell and Wallace (1992). The two types of situation, which indicate a positive (negative) anomaly of precipitation when SST increases (decreases), are in accordance with the positive correlations.

The equatorial cooling which occurs during April and May and the coastal precipitations around/near the Cape of Three Points during the JAS period are subject to the same evolution. But this relation goes diminishing as soon as the equatorial cooling persists.

This study points out *a priori* a non-combined influence of the SSTs of both oceanic zones on the littoral precipitation during the JAS period. The field of observed correlations displays the following structure between the equatorial Atlantic SSTs and the precipitation near the regions of Cape of Three Points: SST cooling associated with a decrease of the precipitation or SST warming linked to an increase of the precipitation. During the JAS period, only the coastal SSTs exert a significant influence on coastal precipitations. The most influenced continental zone is the south-east of Ghana with an extension towards the middle-east of Cote d'Ivoire. Concerning this continental area, the decrease of rainfalls during the JAS period is partly linked to the decrease of the SSTs. This work has been undertaken within the framework of the international program dedicated to the African monsoon (AMMA; eg Redelsperger *et al*., 2006) and its oceanic and air-sea exchanges component EGEE (e.g., Bourles, 2003) and has the merit to focus on the whole GG northern coastline. Our results highlight that both coastal and equatorial upwellings have a significant influence on the precipitations over that particular region during the JAS time period. These results deserve to be confirmed by model calculations in order to better understand the physical mechanisms at stake behind the SST and precipitation relationships.

ACKNOWLEDGMENTS

The authors would like to thank the IRD (Institut de Recherche pour le Developpement) Center in Brest (France) for supplying the observational data used in this study and IRD for funding three research periods of the 1st author in Brest. Based on a French initiative, AMMA was built by an international scientific group and is currently funded by a large number of agencies, including those in France, the UK, US and Africa. It has been the beneficiary of a major financial contribution from the European Community's Sixth Framework Research Programme. Detailed information on scientific coordination and funding is available on the AMMA International website http://www.amma-international.org.

REFERENCES

Adamec, D. and J.J. O'Brien, 1978. The seasonal upwelling in the Gulf of Guinea due to remote forcing. J. Phys. Oceanogr., 8: 1050-1060.

Arfi, R., O. Penzennec, S. Cissoko and M. Mensah, 1991. Variations Spatiale et Temporelle de la Resurgence Ivoiro-Ghaneenne. In: Pecheries Ouest-africaines: Variabilite, Instabilitie Changement, Cuiv, P. and C. Roy (Eds.). ORSTOM, Paris, pp: 162-172.

Bhat, G.S., G.A. Vecchi and S. Gadgil, 2004. Sea surface temperature of the bay of bengal derived from the TRMM microwave imager. J. Atmos. Oceanic Technol., 21: 1283-1290.

Bourles, B., M. D'Orgeville, G. Eldin, R. Chuchla, Y. Gouriou, Y. DuPenhoat and S. Arnault, 2002. On the thermocline and subthermocline eastward currents evolution in the Eastern equatorial atlantic. Geophys. Res. Lett., 29: 4-4.

Bourles, B., 2003. On the Gulf of guinea and the West African monsoon. Clivar Exchanges Lett., 27: 2-3.

Bourras, D., A. Weill, G. Caniaux, L. Eymard and B. Bourles *et al.*, 2009. Turbulent air-sea fluxes in the Gulf of guinea during the AMMA experiment. J. Geophys. Res., 114: 12-12.

Camberlin, P., S. Janicot and L. Poccard, 2001. Seasonality and atmospheric dynamics of the teleconnection between African rainfall and tropical sea-surface temperature: Atlantic vs. ENSO. Int. J. Climatol., 21: 973-1005.

Colin, C., 1988. Coastal upwelling events in front of the ivory coast during the FOCAL program. Oceanol. Acta, 11: 125-138.

Colin, C., 1989. On the variability in the Gulf of Guinea. New considerations on the mechanisms of upwelling. Doctoral Thesis, Museum National d'Histoire Naturelle in Paris, France.

Colin, C., Y. Gallardo, R. Chuchle and S. Cissoko, 1993. Climate and Oceanographic Environments on the Continental Shelf of the Cote d'Ivoire. In: Environment and Water Resources in Cote D'Ivoire: I-The Marine Environment, Marshall, P.E. and J.B.A. Kothias (Eds.). ORSTOM., Paris, pp: 75-110.

Eltahir, E.A.B. and C. Gong, 1996. Dynamics of the wet and dry years in West Africa. J. Climate, 9: 1030-1042.

Emanuel, K.A., 1985. An air-sea interaction theory for tropical cyclones. Part I: Steady-state maintenance. J. Atmos. Sci., 43: 585-605.

Folland, C., J. Owen, M.N. Ward and A.W. Colman, 1991. Investigation of seasonal rainfall in the Sahel region using empirical and dynamical methods. J. Forecast., 10: 21-56.

Folland, C.K., T.N. Palmer and D.E. Parker, 1986. Sahel rainfall and worldwide sea temperature 1901-1985. Nature, 320: 602-607.

Foltz, G.R., S.A. Grodsky, J.A. Carton and M.J. McPhaden, 2003. Seasonal mixed layer heat budget of the tropical Atlantic ocean. J. Geophys. Res., 108: 3146-3159.

Fontaine, B., N. Philippon and P. Camberlin, 1999. An improvement of June-September rainfall forecasting in the Sahel based upon region April-May moist static energy content (1968-1997). Geophys. Res. Lett., 26: 2041-2044.

Fontaine, B., S. Trzaska and S. Janicot, 1998. Evolution of the relationship between near global Atlantic SST modes and the rainy season in West Africa: Statistical analysis and sensitivity experiments. Climate Dynam., 14: 353-368.

Gallardo, Y., 1993. Les Conditions Oceaniques du Maximum des Pluies Littorales Ivoiriennes en Juin. ORSTOM, Ivory Coast, pp: 63-73.

Gouriou, Y. and G. Reverdin, 1992. Isopycnal and diapycnal circulation of the upper equatorial Atlantic ocean in 1983-1984. J. Geophys. Res., 97: 3543-3572.

Gu, G. and R.F. Adler, 2004. Seasonal evolution and variability associated with the West African monsoon system. J. Climate, 17: 3364-3377.

Hastenrath, S., 1984. Interannual variability and annual cycle: Mechanisms of circulation and climate in the tropical Atlantic sector. Mon. Weather Rev., 112: 1097-1107.

Houghton, R.W., 1976. Circulation and hydrographic structure of the Ghana continental shelf during the 1974 upwelling. J. Phys. Oceanogr., 6: 909-924.

Ingham, M.C., 1970. Coastal upwelling in the Northwestern of Gulf of Guinea. Bull. Mar. Sci., 20: 1-34.

Janicot, S., V. Moron and B. Fontaine, 1996. Sahel droughts and ENSO dynamics. Geophys. Res. Lett., 23: 515-518.

Janicot, S., A. Harzallah, B. Fontaine and V. Moron, 1998. West African monsoon dynamics and Eastern equatorial Atlantic and pacific SST anomalies (1970-88). J. Climate, 11: 1874-1882.

Kouadio, Y., D.A. Ochou and J. Servain, 2002. Atlantic influence on the rainfall variability in cote d'Ivoire. Geophys. Res. Lett., 30: 8005-8005.

Lamb, P.J., 1978. Large-scale Tropical Atlantic surface circulation patterns associated with sub-saharian weather anomalies. Tellus, 30: 240-251.

Lamb, P.J. and R.A. Peppler, 1992. Further case studies of tropical Atlantic surface atmospheric and oceanic patterns associated with sub-Saharan drought. J. Climate, 5: 476-488.

Le Barbe, L., T. Lebel and D. Tapsoba, 2002. Rainfall variability I West Africa during 1950-1990. J. Climate, 15: 187-202.

Lighthill, M.J., 1969. Dynamic response of the Indian to the onset of the Southern monsoon. Phil. Trans. R. Soc. Lond., 265: 45-93.

Lough, J.M., 1986. Tropical Atlantic sea surface temperature and rainfall variations in Subsaharan Africa. Mon. Weather Rev., 114: 561-570.

Maloney, E.D. and A.H. Sobel, 2004. Surface fluxes and ocean coupling in the tropical intraseasonal oscillation. J. Climate, 17: 4368-4386.

Maloney, E.D., D.B. Chelton and S.K. Esbensen, 2008. Subseasonal SST variability in the tropical eastern North pacific during boreal summer. J. Climate, 21: 4149-4167.

Marin, F., G. Caniaux, H. Giordani, B. Bourles, Y. Gouriou and E. Key, 2009. Why were sea surface temperature so different in the eastern equatorial Atlantic in June 2005 and 2006. J. Phys. Oceanogr., 39: 1416-1431.

Matthews, A.J., 2004. Intraseasonal variability over tropical Africa during northern summer. J. Climate, 17: 2427-2440.

Mitchell, T.P. and J.M. Wallace, 1992. The annual cycle in equatorial convection and sea surface temperature. J. Climate, 5: 1140-1156.

Moore, D.W., 1968. Planetary-gravity waves in an equatorial ocean. Ph. D. Thesis, Harvard University.

Moore, D., P. Hisard, J. McCreary, J. Merle and J. O'Brien *et al.*, 1978. Equatorial adjustment in the Eastern Atlantic. Geophys. Res. Lett., 5: 637-640.

Moron, V., S. Bigot and P. Roucou, 1995. Rainfall variability in the subequatorial America and Africa and relationship with the main Sea surface temperature modes (1951-1990). Int. J. Climatol., 15: 1297-1322.

Nicholson, S.E. and J.P. Grist, 2003. The seasonal evolution of the atmospheric circulation over West Africa and Equatorial Africa. J. Climate, 16: 1013-1030.

O'Brien, J.J., D. Adamec and D.W. Moore, 1978. A simple model of upwelling in the Gulf of Guinea. Geophys. Res. Lett., 5: 641-644.

Opoku-Ankomah, Y. and I. Cordery, 1994. Atlantic sea surface temperatures and rainfall variability in Ghana. J. Climate, 7: 551-558.

Philander, S.G.H. and R.C. Pacanowski, 1986. A model of the seasonal cycle in the tropical Atlantic ocean. J. Geophys. Res., 91: 14192-14206.

Philander, S.G.H., 1979. Upwelling in the Gulf of Guinea. J. Marine Res., 37: 23-33.

Picaut, J., 1983. Propagation of the seasonal upwelling in the Eastern equatorial Atlantic. J. Phys. Oceanogr., 13: 18-37.

Redelsperger, J.L., A. Diedhiou, C. Flamant, S. Janicot and J.P. Lafore *et al.*, 2006. AMMA, une etude multidisciplinaire de la mousson ouest-africaine. La Meteorol., 54: 22-32.

Rowell, D.P., C.K. Folland, K. Maskell and M.N. Ward, 1995. Variability of summer rainfall over tropical North Africa (1906-92): Observations and modeling. Quart. J. Royal Meteorol. Soc., 121: 669-704.

Roy, C., 1995. The Cote d'Ivoire and Ghana Coastal Upwellings: Dynamics and Changes. In: Dynamics and Use of Sardinella Resources from Upwelling off Ghana and Ivory Coast, Bard, F.X. and K.A. Koranteng (Eds.). ORSTOM, Paris, pp: 346-361.

Rudolf, B., H. Hauschild, W. Rueth and U. Schneider, 1994. Terrestrial Precipitation Analysis: Operational Method and Required Density of Point Measurements. In: Global Precipitations and Climate Change, Desbois, M. and F. Desalmond (Eds.). Vol. 26, Springer-Verlag, Berlin, Heidelberg, pp: 173-186.

Sadrinasab, M., 2009. Three-dimensional numerical modeling study of the coastal upwelling in the Persian Gulf. Res. J. Environ. Sci., 3: 560-566.

Sengupta, D., B.N. Goswami and R. Senan, 2001. Coherent intraseasonal oscillations of the ocean and atmosphere during the Asian summer monsoon. Geophys. Res. Lett., 28: 4127-4130.

Servain, J., J. Picaut and A.J. Busalacchi, 1985. Interannual and seasonal variability of the tropical Atlantic ocean depicted by sixteen years of sea-surface temperature and wind stress. Elsevier Oceanogr. Ser., 40: 211-237.

Servain, J., I. Wainer, H.L. Ayina and H. Roquet, 2000. The relationship between simulated climatic variability modes of the tropical Atlantic. Int. J. Climatol., 20: 939-953.

Sobel, A.H. and H. Gildor, 2003. A simple time-dependent model of SST hot spots. J. Climate, 16: 3978-3992.

Sultan, B. and S. Janicot, 2000. Abrupt shift of the ITCZ over West Africa and intra-seasonal variability. Geophys. Res. Lett., 27: 3353-3356.

Vizy, E.K. and K.H. Cook, 2001. Mechanism by which Gulf of Guinea and Eastern North Atlantic sea surface temperature anomalies can influence African rainfall. J. Climate, 14: 795-821.

Wagner, R.G. and A.M. da Silva, 1994. Surface conditions associated with anomalous rainfall in the Guinea costal region. Int. J. Climatol., 14: 179-199.

Wentz, F.J., C. Gentemann and D. Smith, 2000. Satellite measurements of sea surface temperature through clouds. Science, 288: 847-850.