THE WEST AFRICAN MONSOON

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Abstract

This chapter provides a description of the West African Monsoon (WAM) and of its variability at a wide range of space and timescales. After a summarizing the current understanding of its annual cycle, we analyze its variability on interannual and longer timescales including how the WAM may change in the coming century. The intraseasonal variability and its understanding are then addressed. Finally we summarize the recent progresses made concerning weather systems and process studies.
It is also highlighted the increased interest and research activity concerned with the WAM that has developed substantially in the past decade. Much of this can be attributable to the AMMA project (Redelsperger et al, 2006) and the contributing scientists who were able to mobilize resources to support observations and research and who continue to help coordinate international research on the WAM (see http://www.amma-international.org).

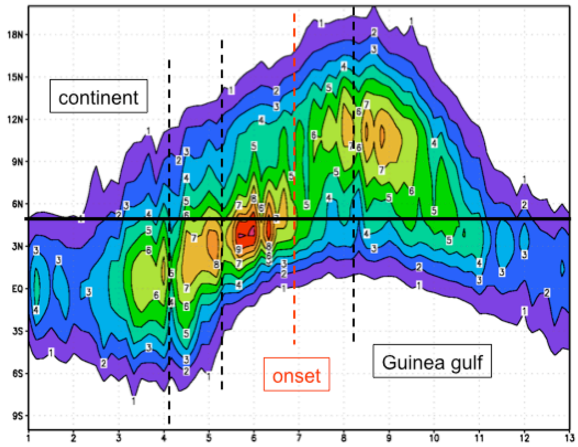
1. INTRODUCTION

The West African monsoon (WAM) is a coupled atmosphere-ocean-land system characterized by summer rainfall over the continent and winter drought. The processes that couple the land, ocean, and atmosphere involve multiple interacting space- and time-scales. Many of the key scientific questions that relate to these scale interactions cannot be answered using routinely available observations and reanalysis data sets. This is due to a combination of the sparsity of the routine observing network over and around West Africa, the need for specialized observations of various climate system parameters, and the known deficiencies of GCMs used for weather and climate prediction and relied upon for producing reanalysis.

As with all monsoon systems, the evolving ocean and land conditions are crucially important for determining the nature of the WAM and its variability. In particular, prospects for improving seasonal-to-interannual prediction of the WAM heavily rely on the potential predictability of these surface conditions, our ability to observe key surface variables needed to initialize dynamical models, and the skill of these models to simulate subsequent evolution of the surface variables.

In this chapter, we begin in section 2 by summarizing the current understanding of the mean annual cycle of the WAM system. It is very likely that problems GCMs and CGCMs have in predicting the WAM climate arise through a misrepresentation of the basic mean annual cycle. In this regard it is important that we understand better the key atmosphere, land, ocean interactions that take place in association with the annual cycle. In section 3 we will focus our attention on the variability of the WAM on interannual and longer timescales including how the WAM may change in the coming century. In sections 4 and 5 we discuss the variability of the WAM at intraseasonal and weather timescales respectively. The motivation for this is two-fold: (i) There is a need to better understand key processes associated with the interactions between atmosphere, land and ocean as well as between dynamics and convection – fundamentally these need to be studied at intraseasonal and shorter timescales. We believe that model improvements at these timescales will very likely have positive implications for longer timescales ; and (ii) Almost all societal applications require information about the statistics of weather systems and so it is important that

Fig. 1 The mean seasonal cycle of rainfall over West Africa through a latitude cross-section. Precipitation values (mm.day^{-1}) from CMAP dataset are averaged over 10°W - 10°E and over the period 1979-2000. The black horizontal line represents the Guinean coast. Vertical lines represent the different steps of the annual cycle, detected through Varimax Principal Component Analysis (Louvet et al. 2003). The red vertical line corresponds to the summer monsoon onset



we improve our understanding, and ultimately our ability to predict, this variability.

2. THE MEAN ANNUAL CYCLE

The mean seasonal cycle of rainfall over West Africa is presented on Fig. 1 through a time-latitude cross-section. It corresponds to a south-north-south displacement of the Inter-Tropical Convergence Zone (ITCZ), which is not a smooth one but is characterized by a succession of active phases and pauses in the convective activity. These different steps have been identified statistically by Louvet et al. (2003). Two of them are located at the time of the first rainy season along the Guinean Coast in April and May, the third one is what is called the “monsoon onset” at the end of June, and a weaker one occurs during the monsoon season in August. The monsoon onset is the strongest one and corresponds to a weakening of the convective activity associated with an abrupt shift to the north of the ITCZ, from 5°N to 10°N (Le Barbé et al. 2002, Sultan and Janicot 2003, Gu and Adler 2004). Over the period 1968-2004, its average date is June 24th, with a standard deviation of 7 days.

The coupled air-sea character of the African monsoon is well-known and could regulate the different steps of the ITCZ annual cycle. The annual cycle of the sea surface temperature (SST) in the Gulf of Guinea is asymmetrical with a rapid cooling from the highest SST in April to the lowest SST in August, and a gradual increase up to the next April. The SST warming in the beginning of the year can explain the rainfall increase observed over the gulf of Guinea starting in March. The fact that SST begins to cool in April, at the onset of the first rainy season along the Guinean Coast, is probably not a coincidence. This evolution results from positive feedbacks between the enhancement of the monsoon winds above the Guinea Gulf associated with the convection enhancement in the ITCZ, the set-up of the equatorial upwelling, the extension of the cooling in the southern tropical Atlantic associated with the strengthening of the Santa Helena anticyclone and the enhancement of the southern Hadley circulation, and the occurrence of low-level stratus clouds over these cold waters (Okumura and Xie 2004). This scenario has been examined recently both through new observational datasets and modelling experiments. Gu and Adler (2004) confirmed it by using recent satellite observation of the 1998-2003 high-quality Tropical Rainfall Measuring Mission (TRMM), water vapour and cloud liquid water, TMI SST, and QuikSCAT surface wind products. Okumura and Xie (2004) tested the impact of the SST in the Guinea Gulf on the West African monsoon. They compared in modelling experiments the evolution of the monsoon forced in an atmosphere model by the annual cycle of the SST with this evolution where the annual cycle of the SST is held constant in time at the mid-April values. They showed that the equatorial cooling exerts a significant influence on the African monsoon by intensifying the southerly winds in the Guinea Gulf and pushing the continental rainband inland. This evolution feeds back positively: first, it contributes to trigger this oceanic cooling in the east; second, easterly winds accelerate in the equatorial Atlantic during northern summer, inducing local upwelling and raising the thermocline in the east, contributing to the westward propagation of the cool equatorial SST.

In a set of experiments, Biasutti et al. (2003, 2004, 2005) investigated the mechanisms controlling the annual cycle of the ITCZ over West Africa and the tropical Atlantic by comparing the relative importance of insolation over land and of the SST. As for Okomura and Xie (2004), they compared the evolution of the ITCZ in an atmospheric model between a realistic annual cycle and an annual cycle where SST and/or insolation is held constant in time from March onward. They concluded that West African rainfall is significantly influenced by SST through the advection of marine boundary layer temperature anomalies over Africa which causes the development of sea level pressure and surface wind convergence anomalies. They showed also that the seasonal changes in insolation control the seasonal changes in the net budget of energy input in the atmospheric column, which is balanced by horizontal energy export in the thermally direct circulation associated with convection in the ITCZ. This modulates the moisture advection inland and controls the rainfall production over West Africa.

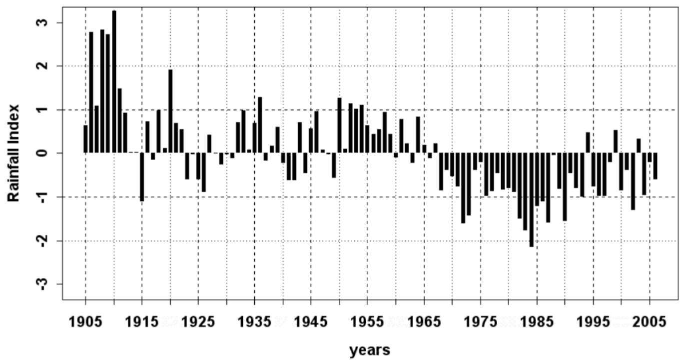
The mechanism associated with this abrupt ITCZ shift is however still unclear. Another hypothesis highlights the role of the Saharan thermal low, which increases at the time of the onset, leading to higher moisture advection inland, and which could be due to some interactions with the North Africa orography (Sultan and Janicot, 2003, Drobinski et al. 2005) combined with the spatial distribution of albedo and net shortwave radiative budget at the surface (Ramel 2004). Recently Hagos and Cook (2007) provided a more comprehensive explanation based on the concept of inertial instability of the monsoon system (Thomas and Webster 1997). Because of the distribution of albedo and surface moisture, a sensible heating maximum is in place over the Sahel region throughout the spring. In early May, this sensible heating drives a shallow meridional circulation and moisture convergence at the latitude of the sensible heating maximum, and this moisture is transported upward into the lower free troposphere where it diverges. During the second half of May, the supply of moisture from the boundary layer exceeds the divergence, resulting in a net increase of moisture and condensational heating into the lower troposphere. The resulting pressure gradient introduces an inertial instability, which abruptly shifts the midtropospheric meridional wind convergence maximum from the coast into the continental interior at the end of May. This in turn introduces a net total moisture convergence, net upward moisture flux and condensation in the upper troposphere, and an enhancement of precipitation in the continental interior through June. Because of the shift of the meridional convergence into the continent, condensation and precipitation along the coast gradually decline.

It is still lacking a comprehensive view of the annual cycle of the WAM combining the internal atmospheric dynamics with land and ocean atmosphere interaction processes. This is true for its onset and also to better explain the asymmetry in latitude of the installation/retreat phases of this monsoon system. This is one of the objectives of the AMMA (African Monsoon Multidisciplinary Analyses) project which has implemented in West Africa over the seasonal cycles 2005-2007 a network of enhanced radiosounding stations, a network of flux stations measuring the radiative budget at the surface, sensible and latent heat fluxes, and soil moisture profiles, in parallel with two oceanographic campaigns per year in the Guinea Gulf.

3. VARIABILITY OF THE WEST AFRICAN MONSOON AT INTERANNUAL AND LONGER TIMESCALES

Fig. 2 shows the well-known time series of rainfall anomalies over the Sahel, updated over the period 1905-2006. While the first part of the 20th century has been characterized by a succession of short wet and dry periods, the second part of the century has known a very unusual evolution of rainfall with a 20-year wet period followed by another 20-year dry period. This long-term negative trend of rainfall has an amplitude that has not been observed anywhere in the world during this century. Since the mid-1980's, we see a tendency towards some recovery even if the rainfall anomalies are still mostly negative. This tendency is more marked over the eastern part of the Sahel than the western part (not shown). For the Gulf of Guinea coastal region, the interannual variability dominates the rainfall time series for the last century (e.g., Rowell et al. 1995, Ward 1998).

Fig. 2 The corresponding May-September Sahelian rainfall index, expressed in normalized anomalies, is computed from 1905 to 2006 by averaging rainfall data between 13°N - 17°N/, 17°W - 22°E. Source of the data: AGRHYMET.



3.1 Understanding of Interannual Variability

The research of the last 30 years has established that the interannual variability of WAM seasonal rainfall exhibits distinct spatial modes of behavior, and that some of those modes are forced by large SST anomaly patterns. WAM seasonal rainfall anomalies tend to be of either (a) opposite sign between the Sahel-Soudan zone and Gulf of Guinea coastal region (i.e., a "dipole"; Lamb 1978a; Nicholson, 1980) or (b) the same sign across all of the Subsaharan West Africa (Nicholson, 1980). The dipole rainfall behavior has been linked to the interannual variability of tropical Atlantic SST anomaly patterns (e.g., Lamb 1978 a, b; Lamb and Peppler 1991, 1992; Rowell et al. 1995; Ward 1998). Sahel-Soudan drought accompanied by Gulf of Guinea wetness is associated with cold (warm) SST anomalies north (south) of 10°N, and vice versa. In contrast, the extension of drought conditions all the way from the Sahara Desert to the Gulf of Guinea coast has been found to coincide with El Niño events in the tropical Pacific Ocean (e.g., Folland et al. 1986; Palmer 1986, Palmer et al. 1992, Rowell et al. 1995; Ward 1998, Giannini et al. 2003).

Colder SSTs in the Gulf of Guinea region tend to be associated with a stronger moist static energy contrast between the land and the ocean and a stronger direct meridional circulation. Rainfall is increased in the Sahel and reduced in the Guinea coastal region. When the SSTs are warmer in the Gulf of Guinea the reverse is true. Zheng et al. (1999) also emphasized a second impact of the SST anomalies. While a stronger direct circulation may be expected with colder SSTs in the Gulf of Guinea, the boundary layer will be expected to be drier due to reduced evaporation (Lamb 1983). The surface fluxes and moisture transports between the ocean and the land need to be investigated along with the processes that determine the variability and predictability of SSTs themselves.

At the interannual time scale, the impact of El Niño events of Sahel drought has been described in several papers including for the most recent ones Rowell (2001), Janicot et al. (2001), Giannini et al. (2003). Rowell (2001) examined in details the mechanisms of this atmospheric teleconnection between the Pacific Ocean and Africa. He demonstrated, as for the mechanism linking MJO and African convection variability (Matthews 2004), the role of the development of equatorial eastward propagating Kelvin wave from east Pacific convective heating anomalies, and westward propagating Rossby waves from the Indian Ocean in response to the anomalous west Pacific – Indian Ocean SST gradients via convective heating anomalies over the Indian Ocean. These interact over Africa to enhance large-scale subsidence over the Sahel and reduce seasonal rainfall totals. Janicot et al. (2001) showed that this Sahel – El Niño teleconnection has not been strong during the whole second part of the last century but has been significantly modulated by the decadal time scale SST anomaly pattern during this period. The long-term warming of the global SST mode, not only favours the long-term drying over the Sahel, but helps also to enhance the atmospheric teleconnection pattern linking El Niño events to Sahel rainfall deficits after 1980, through a fill-in of the monsoon trough and a moisture advection deficit over West Africa. An additional remote mechanism, which bridges the ENSO signal into the tropical Atlantic, is the tropospheric wave train emanating from the tropical Pacific and developing in the southern hemisphere (Huang 2004).

Another ocean – African monsoon relationship at the regional scale concerns the role of the SST anomalies in the Mediterranean Sea. Rowell (2003) explored this teleconnection by forcing an atmospheric model in northern summer with idealized SST anomalies in this basin (positive (negative) anomalies up (down) to 2°C) with climatological SST elsewhere. This experiment shows that positive Mediterranean SST anomalies leads to a wetter Sahel: local evaporation enhances over this basin, increasing the moisture content in the lower troposphere; this additional moisture is advected southward across the eastern Sahara by the mean flow, leading to enhanced low-level moisture convergence over the Sahel which feeds enhanced rainfall. Then positive feedbacks between convection and the local atmospheric circulation help to extend this anomaly. These results have been confirmed by Fontaine et al. (2009), and Jung et al. (2006) confirmed also this link for the summer 2003.

Also, we are aware that the substantial multi-decadal rainfall variability experienced in the Sahel during the 20th century has been associated with an interhemispheric SST anomaly difference, especially marked in the Atlantic, and including warmer-than-average low-latitude water in the Indian oceanic basin (e.g., Ward 1998, Janicot et al. 2001, Bader and Latif 2003, Giannini et al. 2003). More recently the Atlantic Multidecadal Oscillation (AMO) pattern has been highlighted (Sutton et Hodson 2005), strongly associated to the strength of the Thermo-Haline Circulation (THC; Gray et al. 2004). The AMO appears to present a dominant part of the decadal scale SST variability which has been impacting strongly Sahel rainfall (Knight et al. 2006, Baines and Folland 2007). A present debate concerns the relative impact of AMO and Indian Ocean evolution on the recent recovery of Sahel rainfall since the mid-1980s (Hagos and Cook 2008). This is a key issue for the next future of West African climate since it is known that one of the impact of Green House Gaz (GHG) concentration might be a weakening of the THC which would impact the AMO pattern and lead to Sahel rainfall decreasing.

3.2 Scenarios of the Climate Change Impact on the West African Monsoon

Simulations of the climate change impact on the West African monsoon must be considered with great caution. Such scenarios are based on integrations of ensembles of coupled ocean-atmosphere climate models combined with projections of various socio-economical developments. At the global scale, uncertainties come for an equal part from weaknesses of the models and incomplete knowledge of the future economical orientation. For instance Joly et al. (2007) showed that the models used for IPCC4 simulations are unable to reproduce accurately the relationships between Sahel rainfall and ENSO at interannual time scale as well as between Sahel rainfall and the interhemispheric SST gradient at decadal time scale. At the regional scale, uncertainties are larger, due in particular to the inability of the models to simulate accurately the small-scale processes as those associated to water cycle, convection, cloud-radiation and cloud-moisture feedbacks among others. Another problem, specific to the simulation of the West African monsoon, is the existence of a warm bias in the SST of the eastern oceanic basins in the coupled models. This is true for the eastern part of the southern tropical and equatorial Atlantic, which induces a weaker than normal African monsoon as we have seen above.

In spite of these drawbacks in climate modeling, climate scenarios for the future are now regularly produced and scrutinized. The scenarios which have been produced for the IPCC4 Report in 2007 do not provide a coherent response for West Africa and the Sahel region (Held et al. 2005, Cook and Vizy 2006, Lau et al. 2006, Biasutti and Giannini 2006). A primary question is to know whether the recent drought in the 1970s and 1980s (and now the partial recovery if it is confirmed) is more or less a consequence of the global increase of GHG or if it still comes from natural climate variability meaning that the global warming impact has been up now weak. Some studies have linked to decadal scale interhemispheric SST pattern evolution including the Indian Ocean warming to GHG and anthropogenic aerosols increase (Knutson et al. 1999, Stott et al. 2000, Rotstayn and Lohmann 2002) while others argued for natural variability (Hoerling et al. 2006). Biasutti and Giannini (2006) explored the whole IPCC4 dataset and estimated that at least

30% of the observed long-term drought during the 20th century was due to external forcing, that is most likely anthropogenically forced. They also highlighted the significant role of aerosols (mainly located in the northern hemisphere) in forcing the Sahel drought since simulations in which GHG are the only forcing do nor provide a consistent 20th century response. This is in agreement with Hoerling et al. (2006) which suggested that the recent decrease of aerosols may explain the recent rainfall recovery in the Sahel. So considering all these results and model drawbacks, future West African climate remains highly uncertain up to now.

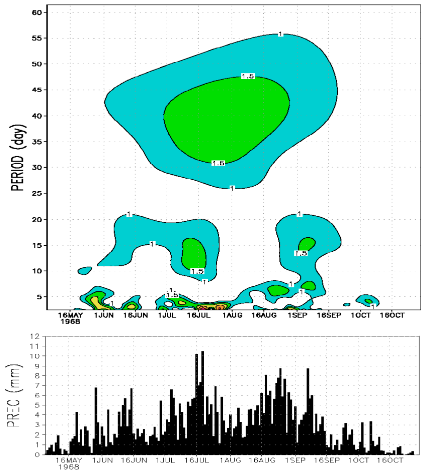


Fig. 3 Daily rainfall time series form 1st of May to 31st of October 1968 averaged over 2.5°W - 2.5°E/12.5°N - 15°N (mm) and associated wavelet diagram. Data set from IRD.

4. INTRASEASONAL VARIABILITY

In this section intraseasonal variability refers to timescales shorter than the season but longer than the timescales associated with individual weather systems such as synoptic African easterly waves (Reed et al, 1977, Berry and Thorncroft, 2005, Kiladis et al, 2006), synoptic convectively coupled Kelvin waves (Mounier et al, 2007, Mekonnen et al, 2009) and the ubiquitous Mesoscale Convective Systems (MCSs, Laing and Fritsch, 1993, Hodges and Thorncroft, 1997). There are two distinct periods that typify the observed intraseasonal variability : 10-25 days, the causes of which remain uncertain (Sultan et al. 2003, Mounier and Janicot 2004, Mounier et al, 2008, Taylor 2008, Janicot et al. 2009b) and 25-60 days which, although climatologically weaker, may be related to the relatively better understood MJO (Mathews, 2004, Maloney and Shaman, 2008, Janicot et al. 2009a, Pohl et al. 2009, Lavender and Matthews 2009). The Fig. 3 shows an example of such variability over the Sahel in 1968. Superimposed on the seasonal cycle of Sahel rainfall characterized by low rainfall in May and October and the highest values during the northern summer, sequences of more than 10 days of persistent high or low rainfall amounts are evident. In particular low values between mid-July and mid-August correspond to a 40-day variability signal.

4.1 Understanding of intraseasonal Variability

While a considerable amount of research has been carried out on the relatively long seasonal-to-decadal timescales (Lamb 1978ab, Nicholson 1978, Ward, 1998, Giannini et al, 2003) our knowledge and understanding of the intraseasonal variability and that associated with extremes is very limited. Such extremes are important in their own right, substantially impacting water and food resources as well as infrastructure. A recent extreme wet period in the West African region occurred during the summer of 2007 and lead to severe flooding and the displacement of more than 500,000 people (this extreme event also affected other parts of tropical Africa). Long dry spells at certain sensitive times in the growing season, especially around the monsoon onset, can also severely impact food security. We continue to lack information about the nature, the causes and predictability of such variability and related extremes in the West African region.

While there is considerable motivation to study intraseasonal variability in its own right, it is important to remember that the sum of these intraseasonal events comprises the seasonal mean. Indeed, the limited work that has taken place on intraseasonal timescales over West Africa (Mounier et al. 2008; Janicot et al, 2009a) has highlighted structures that resemble the patterns of variability identified at interannual and decadal timescales (e.g. Chen et al. 2001, Chelliah and Bell, 2004). It is clear that systematic errors in dynamical models used for climate prediction are often manifested on short timescales (from hours-to-days) which further motivates the study of intraseasonal variability of rainfall in this region from observational as well as modeling perspectives.

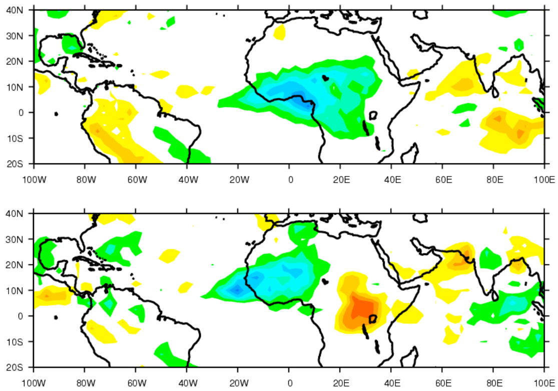


Fig. 4 (top left) Composite pattern of the "QBZD mode" in terms of non-filtered OLR modulation (W.m^{-2}). (bottom left) Same but for the Sahel mode. (right) Same but for the "MJO mode". See mode details in the text.

Only a handful of recent studies have been concerned with intraseasonal variability of the WAM. Sultan et al. (2003) and Mounier and Janicot (2004) were the first to highlight, in a climatological sense, the presence of significant intraseasonal variability over the Sahel. Two distinctive timescales have been identified, 10-25 days and 25-60 days with the former contributing most to the total variance. Mounier et al (2008) explored in more detail the nature of the 10-25 day variability. They proposed the existence of a Quasi Biweekly Zonal Dipole (QBZD; Fig. 4 top left) structure based on a regional EOF analysis. This mode is the leading one over sub-Saharan Africa. The "dipole" aspect of this refers to a nearly standing oscillation with an out-of-phase relationship that African rainfall has with rainfall in the Western Atlantic and Americas. It is controlled both by equatorial atmospheric dynamics and by land surface processes over Africa, inducing combined fluctuations in surface temperatures, surface pressure and low-level zonal winds off the coast of West Africa. When convection is at a minimum over West and Central Africa, a lack of cloud cover results in higher net shortwave flux at the surface, which increases surface temperatures and lowers surface pressures. This creates an east-west pressure gradient at the latitude of both the ITCZ (10°N) and the Saharan heat low (20°N), leading to an increase in eastward moisture advection inland. The arrival from the Atlantic of the positive pressure signal associated with a Kelvin wave pattern amplifies the low-level westerly wind component and the moisture advection inland leading to an increase in convective activity over West and Central Africa. Then the opposite phase of the dipole develops.

The second EOF mode of 10-25-day variability (the "Sahel mode"; Fig. 4 bottom left) has been detected by Sultan et al. (2003) and Mounier and Janicot (2004). The convection increase in the African ITCZ is associated with a propagating mode appearing first over Central Africa, moving to the north towards the Sahelian latitudes, and then propagating westward towards the eastern tropical Atlantic. This pattern is associated with a cyclonic circulation positioned ahead of the enhanced convective pole increasing the advection of moisture towards this pole. Taylor (2008) showed that land-atmosphere interaction processes contribute significantly to the maintenance and the westward propagation of the convective pole. It may also be associated with the occurrence of a convectively coupled equatorial Rossby wave (Janicot et al. 2009b). At this time range Vizy and Cook (2009) also showed that strong cold air surges coming from the Mediterranean area can lead to African monsoon breaks through a southwestward propagation but this signal does not seem to be linked to the "Sahel" mode.

Mathews (2004), then Lavender and Matthews (2009), showed that dry/wet spells over West Africa at the 25-60-day time scale could be linked to a combination of equatorial Kelvin and Rossby waves forced in the Pacific by the MJO. Janicot et al (2009a) and Pohl et al. (2009) confirmed the relationship with the MJO through the dynamics of active/break cycles of the Indian monsoon (see OLR pattern on Fig. 4 right), showed higher impact on dry spells over sub-Saharan Africa, and highlighted the dominant role in this connection of a convectively coupled equatorial Rossby wave (Janicot et al. 2009b).

Sultan et al. (2009) presented a first attempt of medium-lead forecast of these intraseasonal oscillations using the singular-spectrum analysis combined with the maximum entropy method. In an operational context the predictability of individual intraseasonal modes is high at 5 and 10 days range but the forecast skill for the whole intraseasonal signal remains weak. As shown for other such statistical schemes the forecast skill is good when the intraseasonal signal is well defined, and the forecast fails when the characteristics of this signal is changing rapidly

4.2 Seasonal forecasting of the West African monsoon

The above results provide the basis for the seasonal prediction of WAM rainfall from the expected evolution of tropical Atlantic and Pacific SST anomaly patterns during the months leading up to the monsoon season. Both statistical and numerical modeling approaches have been employed with mixed success. A major challenge that persists is the difficulty of predicting the evolution of the tropical Atlantic SST anomaly pattern prior to and during the WAM season. Unfortunately, at present, neither observations nor models are fully adequate to characterize quantitatively the variability and predictability of SSTs in the tropical Atlantic. The problem is rooted in prediction inadequacy for the average state and the seasonal cycle that propagates through to the interannual time-scales.

In contrast to the above dominant role that tropical Atlantic and Pacific SST anomaly patterns have been found to play for the interannual variability of WAM rainfall, the contributions of land-atmosphere interactions are considered to be of subordinate (but possibly amplifying) importance on this time-scale (e.g., Douville 2002, Giannini et al. 2003). We are hindered in investigating the role of the land due to the lack of appropriate large-scale multiyear observations of land surface conditions and low confidence in our ability to model the complex interactions between the land surface and the atmosphere. In order to assess the role of the land on these time-scales, it is necessary to consider the significance of potential feedbacks. Fundamentally, rainfall anomalies are expected to result in anomalies in the surface conditions, notably through impacting soil moisture and vegetation. These surface anomalies will impact the energy and water budgets. It is important to investigate whether, through such impacts, the rainfall anomaly is perpetuated through a positive feedback process or, alternatively, how it is damped. A positive feedback process has been shown to operate at the mesoscale (e.g. Taylor and Lebel 1998) and there is also some evidence to suggest that the seasonal WAM rainfall is impacted by the rainfall at the end of the previous rainy season (Philippon and Fontaine 2002). Recently Douville et al. (2007) has reactivated this debate by showing that the phasing of equatorial Pacific and Guinea Gulf SST might explain such a “memory” effect.

Fontaine et al. (1999) showed that using information related to land surface properties during spring (meridional gradient of low-level entropy over West Africa and over the Guinean Coast) enables to increase significantly the scores of statistical seasonal forecast skills. This is true also when statistical adaptations using such atmospheric dynamical parameters are applied on ocean-atmosphere models seasonal forecasts. Bouali et al. (2008, 2009) have shown that the direct WAM rainfall seasonal forecast from DEMETER and ENSEMBLES multimodel ensemble system is poor and they have demonstrated that it is possible to increase significantly this skill through model-output-statistics procedures using such parameters.

5. WEATHER SYSTEMS AND PROCESS STUDIES IN THE WAM

The weak skill of GCMs and NWP models in catching the West African characteristics and more specifically its associated precipitation pattern and variability, strongly motivated the international AMMA program. Through its multi-scale approach, AMMA promoted the idea that a better knowledge of physical processes involved in the complex chain resulting in rainy weather systems is necessary to improve their representation in GCM and NWP models. The Specific Periods of Observation (SOP) in 2006 is one facet of this effort (Redelsperger et al. 2006), whose current scientific exploitation has fostered the research to understand the functioning of the WAM system involving numerous couplings between the atmosphere, the continental surface, the ocean and aerosols. The Fig. 5 provide a schematic 3D view of the WAM system, exhibiting some key features that we will review below to evaluate the progress performed and to identify the still open questions to be answered.

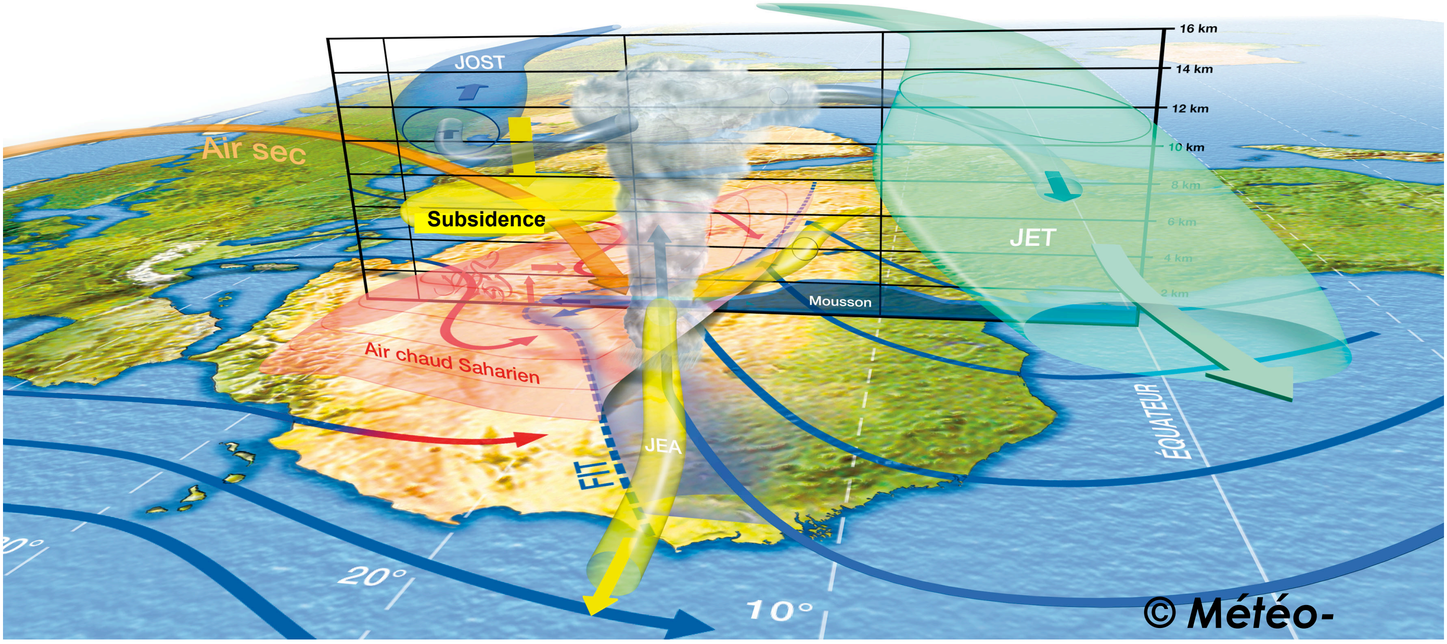


Fig. 5 3D schematic view of the West African Monsoon (see text for details). Main features are indicated in French. Their transation in English are the following: ITD (InterTropical Discontinuity) for FIT, 'Warm Saharian Air" for "Air chaud Saharien", AEJ (African Easterly Jet) for JEA, TEJ (Tropical Easterly Jet) for JET, Subtropical Westerly Jet JOST, "Dry Air" for "Air sec".

5.1 Structure and Circulation at Low-Levels

The Saharan Heat Low (SHL) is a major feature of the WAM (Thorncroft and Blackburn 1999, Thorncroft et al. 2003, Parker et al. 2005) corresponding to a deep dry-convective ABL (red dome in Fig. 5). The SHL minimum of pressure located on its southern flank drives the convergence of 2 opposite low level flows along the Inter Tropical Discontinuity (ITD dashed blue line): *i.e.* the northerly dry and hot flux (or Harmattan) and the south-westerly moist and fresher monsoon flux. The resulting strong baroclinicity across this discontinuity is responsible for the mid-level African Easterly Jet (AEJ yellow tube). Moreover the low-level negative PV anomaly in the SHL region (weak vertical stability) coupled with the positive PV anomaly below convection in the ITCZ (Inter Tropical Convergence Zone) to the south of the ITD (~5°) is responsible for the PV sign-reversal that supports AEW growth (Thorncroft and Blackburn 1999).

The AMMA SOP in 2006, (together with the UK GERBILS campaign in 2007), provided a unique documentation of the SHL and ITD with a wide range of measurements (lidar, dropsondes, aerosols, radiation…) from ground, aircraft and satellites (CALIPSO). First results stress the role of ITD, turbulence and convective gusts to mobilize and uplift mineral dusts (Flamant et al. 2007, Bou

Karam et al. 2007). Pospichal and Crewell (2007) illustrated the pronounced diurnal cycle of the ITD during the transition period, with strong gradients in PBL. Also the Saharan ABL structure, its seasonal evolution and its radiative budget have been analyzed using the unique dataset acquired in Tamanrasset (Cuesta et al., 2007). Sultan and Janicot (2003) suggested that the monsoon onset, initiated by the amplification of the SHL, could be due to interactions with the northern orography of the Atlas and Hoggar Mountains. Drobinski et al. (2005) confirmed this explanation using a linear model. Drobinski et al. (2009) illustrated in a case study the role played by orography on the late northward monsoon propagation in 2006. Idealized approaches (Chou et al. 2001, Peyrille and Lafore 2007) or regional simulations (Ramel *et al.* 2006) allowed a better understanding of processes involved in the HL formation, equilibrium and its diurnal cycle such as ventilation, turbulence in the ABL and albedo. Recently Lavaysse et al. (2009) developed a 35-year climatology of the SHL that included its seasonal migration over Africa and variability. Despite the recent progress, further studies are needed to understand the internal functioning of the SHL (turbulence, aerosols, surface, ventilation) and its interplay with the other WAM components such as monsoon bursts, convection, AEJ and AEWs. For instance Couvreux et al. (2009) noted that the monsoon pulsations were closely related to SHL oscillations of the SHL low magnitude.

The AMMA SOP also allowed documenting in detail the structure and evolution of the ABL and the monsoon fluxes at different latitudes from the coast to the northern flank of the Sahel. Of particular interest is the atmosphere budget allowed for the whole 2006 year by the RADAGAST (Radiative Atmospheric Divergence using ARM Mobile Facility, GERB data and AMMA Stations) experiment in Niamey (12.5°N). Slingo et al. (2008, 2009) provided an overview for the 2006 year of the atmosphere structure, of radiative fluxes at its top and at the surface and examined the factors that control these fluxes. Similarly Guichard et al. (2009) analyzed seasonal and diurnal cycles of the thermodynamics and of the radiative budget at the surface on the Malian Gourma site (15.3°N) for several years. The relative role of aerosols, clouds, atmospheric water vapour and temperature, together with surface precipitation and the growth of vegetation (via surface emission and albedo) on the monsoon settlement during the season is provided through the analysis of the net radiation at the surface. It shows how the combination of these processes combine to finally allowing for an increase in the monsoon moist energy needed for convection in the Sahel. Lothon et al. (2008) detailed the diurnal cycle of the low troposphere from profilers and radiosondings at different locations. In particular the importance of the nocturnal low level jet all over the year for the northward moisture transport, maximal around 0500 UTC and centered at 400 m height above the ground in Niamey, with a maximum around 10 m s^{-1}. Current works concentrate on the turbulent transports within the ABL and on the mixing between the monsoon moist layer the overlaying Harmattan dry layer, occurring in a sheared layer documented by aircraft measurements (Canut et al. 2009) and LES. The importance of this mixing for the northward propagation of the monsoon has been shown by Peyrillé (2006). An important goal is to improve the representation of above ABL processes in current parameterizations and test their impact on the skill of GCM and NWP in representing the monsoon seasonal cycle. The above unique dataset provide a valuable ground truth for assessing models over this area. Other issues under consideration concern the interaction with aerosols and the coupling with the surface.

5.2 African Easterly Waves

African easterly waves (AEWs) are the major synoptic weather systems in the West African summer monsoon. They are westward traveling waves originating west of 20°E, and develop through both barotropic and baroclinic energy conversions as they move along the African easterly Jet (AEJ) located around 600 hPa. Their wavelength varies between 2000 and 4000 km and they have a westward phase speed of about 8 m s^{-1} (leading to spectral periodicities between 3 and 5 days). Observations and modeling studies also indicate that AEWs significantly interact with moist convection (Duvel 1990, Diedhiou et al. 1999). Kiladis et al (2006) recently highlighted the composite structure of AEWs and how the phase relationship with convection varies with longitude. Over the continent convection tends to be in the northerlies ahead of the trough. Near the West coast it tends to be in the trough and just downstream of the continent it shifts into the southerlies. A complete understanding of the reasons for these phase shifts, how AEWs interact with convection and, in particular, how AEWs and MCSs interact is still lacking.

Berry and Thorncroft (2005) recently carried out a case study of an AEW and, based on this and previous related studies in the literature, promoted a PV-_ approach for describing and understanding AEW life-cycles that takes account of both the synoptic and embedded sub-synoptic scale aspects of AEWs, key for understanding scale interactions. A conceptual AEW life-cycle is envisaged with 3 phases: (i) Initiation, (ii) Baroclinic developments and (iii) West coast developments (see Fig. 6):

Phase I: Initiation: Genesis of AEWs are often preceded by several mesoscale convective systems (MCSs) triggered over the elevated terrain of Darfur around 25^{o}E (although triggering can take place in other locations). It is hypothesized that the dynamical response to the heating associated with this outbreak of convection close to the axis of the African easterly jet (AEJ) results in downstream development of baroclinically growing AEWs. This hypothesis is supported by the statistical analyses of Mekonnen et al (2006), Kiladis et al (2006) and the modeling work of Hall et al (2006) and Thorncroft et al (2008). An alternative hypothesis for the role of upstream convection was proposed by Hsieh and Cook (2005, 2007) who suggested that the key atmospheric response involved the establishment of a more unstable AEJ (cf Schubert et al, 1991). Recently Leroux and Hall (2009) analyzed the optimal structure of the AEJ favouring AEW development.

(a) Phase I (b) Phase II (c) Phase III

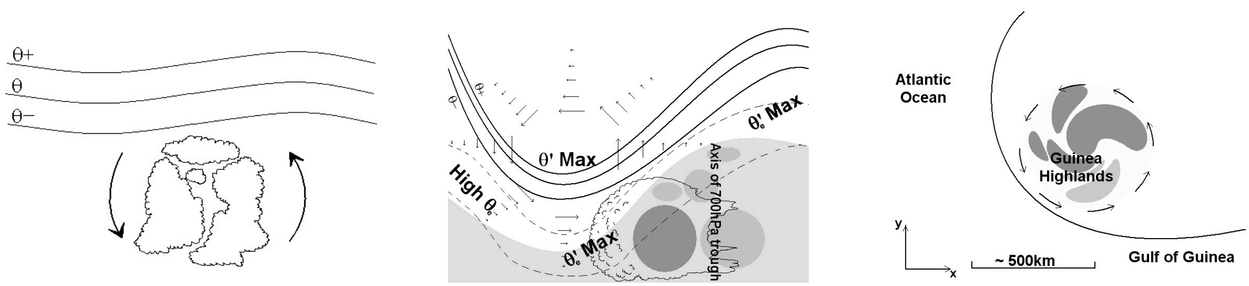


Fig. 6 Schematic illustrating 3 phases of observed AEW life-cycles (see text for details).

Phase II: Baroclinic Development: The second phase of the AEW life cycle commences as the synoptic scale perturbation to the low-level _ gradient interacts with a synoptic scale perturbation to the AEJ level PV strip. For approximately 2 days over West Africa, these perturbations move together in a configuration that is consistent with baroclinic growth. Isolated sub-synoptic PV maxima generated by convection ahead of and within the PV-trough contribute to the strength of the PV-trough and the baroclinic interaction. We know little about the nature of the sub-synoptic PV structures, important for understanding these key scale interactions.

Phase III: West Coast Developments: As the AEWs approach the West African coast significant convection often develops over the Guinea highlands. The Guinea highlands PV anomalies can then merge with the sub-synoptic scale PV anomalies embedded in the propagating AEW. Analysis indicates that the resulting coherent structures can be moist and warm core before leaving the West African coast. This is therefore an ideal structure for triggering tropical cyclogenesis and highlights the importance of considering the nature and variability of these systems, including their upstream structure, in more detail. Hopsch et al (2007) have studied the significance of these events on tropical cyclogenesis events in a statistical sense.

5.3 Kelvin Waves

Kelvin waves have been recently detected over West and Central Africa during northern summer following the wavenumber-frequency filtering procedure developed by Wheeler and Kiladis (1999) and applied to the NOAA OLR data set over Africa (Mounier et al. 2007, Mekonnen et al, 2008). Fig. 7 shows the sequence of the composite fields from to-2 days to to+2 days of non-filtered OLR (shaded) and 925 hPa geopotential heights and winds for the difference between the dry and wet Kelvin phase. A wet (dry) Kelvin phase designates a Kelvin wave occurrence over

West and Central Africa associated with negative (positive) OLR anomalies, that is higher (weaker) convective activity in the ITCZ (see Mounier et al. (2007) for more details). The high range of the OLR differences between wet and dry Kelvin phases means that Kelvin waves are able to significantly modulate the convective activity in the ITCZ during the summer monsoon. In contrast to the AEWs, the Kelvin wave signal is propagating eastward with a mean phase velocity of about 15 m s^{-1} and it reveals a mean periodicity of about 6 days. Its zonal extension corresponds to an approximate wavelength of 8000 km. A large part of the convective activity over West and Central Africa can be influenced by the crossing of such a Kelvin wave. Its dynamical pattern is very similar to theoretical equatorially trapped Kelvin wave solution on an equatorial beta-plane computed using a shallow water model (Matsuno 1966). The passage of such a wave is preceded by easterly and followed by westerly wind anomalies, in phase with negative and positive geopotential perturbations respectively. Most of the flow is in the zonal direction as predicted by theory, although the strong monsoonal heating over West Africa does favor meridional southerly inflow. This wind field anomaly pattern leads to an enhancement of inland moisture advection by the zonal flow component. The convergence of westerly and easterly flows contributes then to the enhanced convection over West and Central Africa. The dynamical fields tend to be much more symmetric about the equator, with large Southern Hemisphere signals, even though the convection is concentrated well north of the equator, at the latitude of the ITCZ.

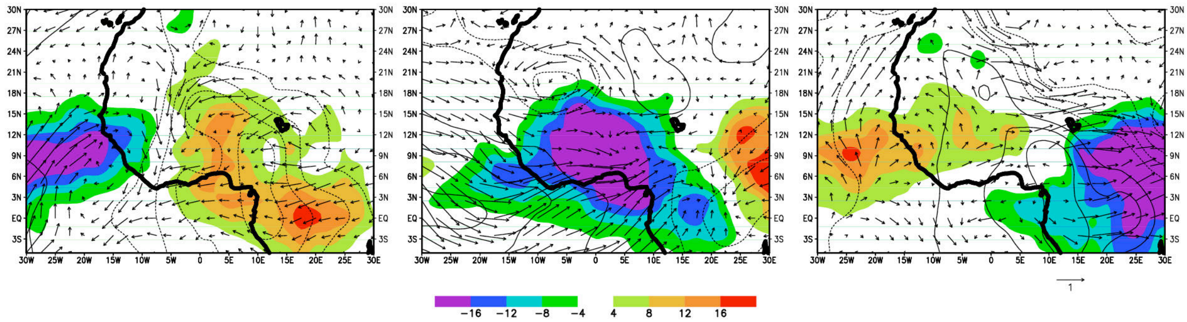


Fig. 7 Sequence of composite fields of non-filtered OLR (shaded) and 925 hPa geopotential heights (1 mgp) and winds (m/s) for the difference between the wet and dry Kelvin phase for t_0-2, t_0 and t_0+2 days where t_0 represents the date of the highest signal over 7.5°N-12.5°N/-10°W-10°E (see Mounier et al. 2007).

5.4 Tropical-Extratropical Interactions

Knippertz (2007) made a review of tropical-extratropical interactions (TEI) related to upper-level through at low latitudes, including theoretical, modeling and observational studies. The penetration of an upper-level trough into the Tropics is often associated with enhanced convection and the formation of an east- and poleward stretching elongated band of upper- and midlevel clouds, usually referred to as a ‘tropical plume’ (TP). He provided a schematic model (see his Fig. 11) of the synoptic situation during these precipitation events over West Africa in connection with TP occurring in boreal winter and spring. A complete statistic of Tropical TP days in 2006 has been undertaken, with an identification of 28 TEI events. Although less frequent and intense during the monsoon season, forcing mechanisms by the subtropical jet (STJ) and troughs at low latitudes may play a role in helping triggering convection, in particular around the monsoon onset. Recently Knippertz and Fink (2008) suggested a coupling between the TP and the Heat Low: *i.e.* through a local enhancement of the greenhouse effect due to the high clouds and a higher water vapor Content, the TP contributes to the pressure drop.

For the 1992 year Roca *et al.* (2005) showed that dry events in the mid-troposphere (< 5%) over Sahel originated in the upper levels (200–250 hPa) on the anticyclonic side of the polar jet stream at 50°N. The mean trajectory of these “dry intrusion” sinking while turning to the south below the STJ acting as a PV barrier, to finally turning to the SW while gently descending over Sahara to reach the Sahel band in the upper part of the AEJ. A composite study shows that “dry intrusions” favor the development of long-lived (> 12 hr) organized MCS, contrary to results over the TOGA-COARE region (Sherwood 1999, Parsons et al. 2000) where dry intrusion reduce the convective activity. The low level shear coexisting with the dry air in the Sahel region is suggested as the main reason of this difference of behavior. The "extratropical dry intrusion" have been monitored during the AMMA SOP (Thorncroft *et al.* 2007) and revealed that they can either organize or suppress convection for reasons that are currently not understood. Thus above TEI interactions and "extra-tropical dry intrusions" can enhance the predictability of the WAM system and need further investigation.

5.5 MCS and Anvils

Convection is organized in the ITCZ to the south of the ITD (Fig. 5). When active their upper-level anticyclonic and divergent flow feed the TEJ and STJ on their southern and northern flanks respectively as shown by CRM simulation or idealized models (Diongue et al. 2002, Peyrillé et al. 2007). Fast-moving MCS (Mathon et al. 2001, 2002) account for most of the rain over Sahel but less organized and short-lived systems to the south need attention. The understanding, forecasting and representation in GCMs of the SCO is one of the challenges for the WAM. Although the theory of such systems is now well established for this region (Rotunno et al 1988, Moncrieff and So 1989, Lafore and Moncrieff, 1989), their interplays with the above large scale features (AEJ, AEWs, vortices, dry air, troughs…) and with the surface stay unclear. Of particular importance is their triggering, as the convective inhibition is large over the Sahel. AMMA SOP brought a huge dataset to tackle this key issue (satellite data, radiosondes network, surface and airborne observations, 4 Doppler radars with 3 including dual polarization). Of particular interest is the MIT radar operated during two seasons (2006 and 2007) with clear air measurement capacities up to 3 km. The treatment of such amount of data represents a huge effort, but will be a unique source of information for future studies. First results such as Chong (2009) confirm the squall-line conceptual model we have and the specific nature of precipitation above and under the melting level. The Cloud Resolving Models now can realistically simulate such systems (Barthe et al. 2009) offering an adequate tool to study the interactions of African MCSs with their environment. Also AMMA SOP provides a unique documentation of the microphysical and radiative properties of tropical anvils owing to the combination of radar, lidar and in situ observations, from ground, aircraft and satellite (Calipso). Bouniol et al. (2008) evaluated the CLOUDSAT reflectivity against the airborne radar. Combined with statistics (Protat et al. 2009) and simulations of anvil, it is now possible to study their characteristics and role on the WAM by modulating the atmosphere and surface radiative budget driving thus the surface fluxes.

6. FINAL COMMENTS

This chapter has provided a description of the WAM and its variability at a wide range of space and timescales. It has also highlighted the increased interest and research activity concerned with the WAM that has developed substantially in the past decade. Much of this can be attributable to the AMMA project (Redelsperger et al, 2006) and the contributing scientists who were able to mobilize resources to support observations and research and who continue to help coordinate international research on the WAM (see http://www.amma-international.org). In particular, the AMMA Special Observing Period that took place during 2006 has provided us with a unique dataset to study the multiple scales and related processes that characterise the coupled atmosphere-land-ocean monsoon system. While AMMA has already brought new and exciting insights, more effort is required to exploit this vast dataset and to ensure appropriate pull-through to improve models used for weather and climate prediction and their impacts.

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http://www.amma-international.org.

References

Albignat, J. P., and R. J. Reed, 1980: The origin of African wave disturbances du¬ring Phase III of GATE. Mon. Wea. Rev., 108, 1827–1839.

Bader, J. and M. Latif, 2003: The impact of decadal-scale Indian ocean SST anomalies on Sahelian rainfall and the North Atlantic Oscillation. *Geophys. Res. Lett.,* 30, 2169, doi:10.1029/2003GL018426.

Baines, P.G. and C.K. Folland, 2007: Evidence for a rapid global climate shift across the late 1960s. *J. Climate*, **20**, 2721-2744.

Barthe C., N. Asencio, J.-P. Lafore, M. Chong, B. Campistron and F. Cazenave 2009: Multi-scale analysis of the 25-27 July 2006 convective period over Niamey: comparison between Doppler radar observations and simulations. *Q. J. Roy. Meteorol. Soc.*, in revision.

Berry, G. J., and C. D. Thorncroft, 2005: Case study of an intense African Easterly Wave. Mon. Wea. Rev., 133, 752–766

Biasutti, M, D. S. Battisti, and E. S. Sarachik (2003) The Annual Cycle over the Tropical Atlantic, South America, and Africa. *Journal of Climate*, 16, 2491-2508.

Biasutti, M, D. S. Battisti, and E. S. Sarachik (2004) Mechanisms controlling the annual cycle of precipitation in the Tropical Atlantic sector in an AGCM. *Journal of Climate*, 17, 4708-4723.

Biasutti, M, D. S. Battisti, and E. S. Sarachik (2005) Terrestrial influence on the annual cycle of the Atlantic ITCZ in an AGCM coupled to a slab ocean model.. *Journal of Climate*, **18**, 211-228.

Biasutti, M. and A. Giannini, 2006: Robust Sahel drying in response to late 20th century forcings. *Geophys. Res. Lett.*, **33**, doi:1029/2006GL026067.

Bouali, L., N. Philippon, B. Fontaine and J. Lemond, 2008: Performance of DEMETER calibration for rainfall forecasting purposes: Application to the July-August Sahelian rainfall. J. Geophys. Res., **113**, D15111, doi:10.1029/2007JD009403.

Bouali, L., B. Fontaine and N. Philippon, 2009: Validation and predictability of the July-August Sahelian rainfall in DEMETER coupled simulations. *Climate Dyn*., in press

Bou Karam D., C. Flamant, P. Knippertz, O. Reitebuch, J. Pelon, M. Chong, and A. Dabas, 2008: Dust emissions over the Sahel associated with the West African Monsoon inter-tropical discontinuity region: a representative case study, *Q. J. Roy. Meteorol. Soc.,* **134,** 621-634.

Bouniol, D., A. Protat, A. Plana-Fattori, M. Giraud, N. Grand et J.-P. Vinson, 2008: Comparison of airborne and spaceborne 95 GHz radar reflectivity and evidence of multiple scattering effect in spaceborne measurements. *J. Atmos. Oceanic Technol*., 25(11), 1983-1995.

Burpee RW (1972) The origin and structure of easterly waves in the lower troposphere of North Africa. J. Atmos. Sci. 29: 77-90

Burpee, R.W.,1974: Characteristics of North African easterly waves during the summers of 1968 and 1969. J. Atmos. Sci., 31, 1556–1570.

Burpee, R.W., 1975: Some features of synoptic-scale waves based on a compositing analysis of GATE data. Mon. Wea. Rev., 103, 921–925.

Canut, G., M. Lothon, F. Said and F. Lohou, 2009: Observation of entrainment at the interface between monsoon flow and Saharan Air Layer. *Q. J. R. Meteorol. Soc.*, Accepted.

Carlson, T. N., 1969a: Synoptic histories of three African disturbances that developed into Atlantic Ocean hurricanes. Mon. Wea. Rev., 97 (3), 256–276.

Carlson, T. N., 1969b: Some remarks on African disturbances and their progress over the tropical Atlantic Ocean. Mon. Wea. Rev., 97 (10), 716–726.

Chelliah, M. and G.D. Bell, 2004. Tropical multidecadal and interannual climate variability in the NCEP-NCAR reanalysis. *J. Climate*, **17**, 1777-1803.

Chen, T.C., J. Yoon, K. J. St Croix and E.S. Takle, 2001. Suppressing impacts of the Amazonian deforestation by the global circulation change. *Bull. Amer. Meteor. Soc*, **82**, 2209-2216.

Chong M., 2009: The 11 August 2006 squall-line system as observed from MIT Doppler radar during the AMMA SOP. Quart. J. Roy. Meteor. Soc., 135, 18 pages.

Chou, C. and J. D. Neelin, 2003. Mechanism limiting the northward extend of the northern summer monsoons over North America, Asia and Africa. *J. Climate*, 16:406–425.

Cook, K. and E. K. Vizy, 2006: Coupled model simulations of the west african monsoon system: 20th and 21st century simulations. *J. Climate*, **19**(15), 3681–3703.

Couvreux, F., F. Guichard, O. Bock, B. Campistron, J.-P. Lafore and J.-L. Redelsperger, 2009: Synoptic variability of the monsoon flux over West Africa prior to the onset. *Q. J. R. Meteorol. Soc.,* Accepted.

Cuesta, J., E. Dimitri, M. Mimouni, P.H. Flamant, C. Loth, F. Gibert, F. Marnas, A. Bouklila, M. Kharef, B. Ouchène, M. Kadi, and C. Flamant, 2008: Multi-platform observations of the seasonal evolution of the Saharan atmospheric boundary layer in Tamanrasset, Algeria, in the framework the African Monsoon Multidisciplinary Analysis field campaign conducted in 2006, *J. Geophys. Res.*, **113,** D00C07, doi:10.1029/2007JD009417.

Diedhiou A. S., Janicot, A. Viltard, P. de Felice and H. Laurent, 1999. Easterly wave regimes and associated convection over West Africa and tropical Atlantic: results from NCEP/NCAR reanalysis. *Climate Dynamics*, **18**, 241-253.

Diedhiou A. S., Janicot, A. Viltard, and P. de Felice, 2001. Composite patterns of easterly disturbances over West Africa and tropical Atlantic: a climatology from the 1979-1995 NCEP/NCAR and ECMWF reanalysis. *Climate Dynamics*, **15**, 795-822.

Diongue, A., J. P. Lafore, J. L. Redelsperger, and R. Roca, 2002: Numerical study of a Sahelian synoptic weather system: Initiation and mature stages of convection and its interactions with the large-scale dynamics. *Quart. J. Roy. Meteor. Soc.,* 128, 1899–1928.

Douville H, 2002: Influence of soil moisture on the Asian and African monsoons. Part II: Interannual variability. *J. Climate* **15**, 701-720

Douville, H., S. Conil, S. Tyteca and A. Voldoire, 2007: Soil moisture memory and West African predictability: artefact or reality ? *Climate Dyn*., **28**, 723-742.

Drobinski P., B. Sultan, and S. Janicot (2005) Role of the Hoggar mountain in West African monsoon onset. *Geophysical Research Letters*, **32**, L01705, doi :10.1029/2004GL020710

Drobinski, P., S. Bastin, S. Janicot, O. Bock, A. Dabas, P. Delville, O. Reitebuch and S. Sultan, 2009: On the late northward propagation of the West African monsoon in summer 2006 in the region of Niger/Mali. J. Geophys. Res., **114**, D09108, doi :10.1029/2008JD011159

Duvel, J.P., 1990. Convection over tropical Africa and the Atlantic ocean during northern summer. Part II: Modulation by easterly waves. *Mon. Wea. Rev.,* **118**, 1855-1868.

Fink, A. H., and A. Reiner, 2003: Spatio-temporal variability of the relation between African easterly waves and West African squall lines in 1998 and 1999. J. Geophys. Res., 108(D11), 4332, doi:10.1029/2002JD002816.

Fink, A. H., D. G. Vincent, P. Reiner, and P. Speth, 2004: Mean state and wave disturbances during phases I, II, and III of GATE based on ERA-40. Mon. Wea. Rev., 132 (7), 1661–1683.

Fink, A. H., D. G. Vincent, and V. Ermert, 2006: Rainfall Types in the West African Soudanian Zone during the Summer Monsoon 2002. Accepted for publication in Mon. Wea. Rev.

Flamant C., J.-P. Chaboureau, D. J. Parker, C. M. Taylor, J.-P. Cammas, O. Bock, F. Timouck, and J. Pelon, 2007: Airborne observations of the impact of a convective system on the planetary boundary layer thermodynamics and aerosol distribution in the inter-tropical discontinuity region of the West African Monsoon, Q. J. Roy. Meteorol. Soc., 133, 1175-1189, doi: 10.1002/qj.97.

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

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 (1968-1977). *Geophys. Res. Lett.,* **26**(14), 2041-2044.

Fontaine, B., J. Garcia-Serrano, P. Roucou, B. Rodriguez-Fonseca, T . Losada, F. Chauvin, S. Gervois, S. Sivarajan, P. Ruti and S. Janicot, 2009 : Impacts of warm and cold situations in the Mediterranean basin on the West African monsoon : Observed connection patterns (1979-2006) and climate simulations. *Climate Dyn*., accepted.

Fortune, M., 1980: Properties of African squall lines from time-lapse satellite imagery. Mon. Wea. Rev., 108, 153–168.

Fyfe, J. C., 1999: Climate simulations of African easterly waves. J. Climate, 12, 1747–1769.

Giannini, A., R. Saravanan, and P. Chang, 2003: Oceanic forcing of Sahel rainfall on interannual to interdecadal time scales. *Science*, **302**, 1027-1030.

Gray, S.T., L.J. Graumlich, J.L. Betancourt and G.T. Pederson, 2004: A tree-ring reconstruction of

the Atlantic multidecadal oscillation since 1567 A.D. *Geophys. Res. Lett.*, 31, doi:1029/2004GL019932.
Grist JP, Nicholson SE , Barcilon AI (2002) Easterly waves over Africa. Part II: Observed and modeled contrasts between wet and dry years. Mon. Wea. Rev. 130: 212-225
Grodsky S. A. and J.A. Carton, 2001. Coupled land/atmosphere interactions in the West African Monsoon. *Geophys. Res. Lett.,* **28**, 1503-1506.
Grueber A., 1974. The wavenumber-frequency spectra of satellite-measured brightness in the Tropics. *J. of Atmos. Sci.*, 31, 1675-1680.
Gu G. and R.F. Adler, 2004: Seasonal evolution and variability associated with the West African monsoon system. *J. Climate*, 17, 3364-3377.
Guichard, F., L. Kergoat, E. Mougin, F. Timouk, F. Baup, P. Hiernaux and F. Lavenu , 2009: Surface thermodynamics and radiative budget in the Sahelian Gourma: seasonal and diurnal cycles, J. Hydrology (AMMA special issue), doi:10.1016/j.jhydrol.2008.09.007.
Hagos, S.M. and K.H. Cook, 2007: Dynamics of the West African monsoon jump. *J. Climate*, 20, 5264-5284.
Hagos, S.M. and K.H. Cook, 2008: Ocean warming and late-twentieth century Sahel drought and recovery. *J. Climate*, **21**, 3797-3814.
Hall, N. M. J., G. N. Kiladis, and C. D. Thorncroft, 2006.Three-dimensional structure and Dynamics of African easterly waves. Part II: Dynamics. *J. Atmos. Sci.*, 63, 2231-2245.
Hayashi Y ,1979. A generalized method for resolving transient disturbances into standing and travelling waves by space-time spectral analysis. *J Atmos Sci* 361017-1029
Held, I. M., T. L. Delworth, J. Lu, K. L. Findell and T. R. Knutson, 2005: Simulation of Sahel drought in the 20th and 21st centuries. Proc. Natl. Acad. Sci., **102**(50), 17891–17896.
Hodges, K.I. and C.D. Thorncroft, 1997. Distribution and statistics of African mesoscale convective weather systems based on the ISCCP METEOSAT imagery. *Mon. Wea. Rev.,* 125, 2821-2837.
Hopsch, S.B., Thorncroft, C. D.**,** Hodges, K. I. and Aiyyer, A. 2007 West African storm tracks and their relationship with tropical cyclones, J. Climate, 20, 2468-2483.
Hoerling, M., J. Hurrell, J. Eischeid and A. Phillips, 2006: Detection and attribution of 20th century northern and southern African rainfall change. *J. Climate*, **19**(16), 3989–4008.
Hsieh, J.-S. and K.H.Cook, 2005: Generation of African easterly wave disturbances: Relationship to the African easterly jet. Mon. Wea. Rev., 133, 1311-1327.
Hsieh, J.-S. and K.H.Cook, 2007: A study of the energetics of African Easterly Waves using a regional climate model. J. Atmos. Sci. 64, 421-440.
Huang, B., 2004: Remotely forced variability in the tropical Atlantic ocean. *Climate Dyn.*, 23, 133-152.
Janicot, S., S. Trzaska, and I. Poccard, 2001: Summer Sahel-ENSO teleconnection and decadal time scale SST variations. *Climate Dyn.*, **18**, 303-320.
Janicot, S. and B. Sultan, 2001. Intra-seasonal modulation of convection in the West African monsoon. *Geophys. Res. Lett.,* 28, 523-526.
Janicot, S., F. Mounier, N. Hall, S. Leroux, B. Sultan and G. Kiladis, 2009a: The West African monsoon dynamics. Part IV: Analysis of 25-90-day variability of convection and the role of the Indian monsoon. *J. Climate,* **22**, 1541-1565.
Janicot, S., F. Mounier, S. Gervois, B. Sultan and G. Kiladis, 2009b: The dynamics of the West African monsoon. Part V: The role of convectively coupled equatorial Rossby waves. Submitted to *J. Climate*.
Joly, M., A. Voldoire, H. Douville, P. Terray and J.F. Royer, 2007: African monsoon teleconnections with tropical SSTs: validation and evolution in a set of IPCC4 simulations. *Climate Dyn.*, DOI 10.1007/s00382-006-0215-8
Jung, T. L. Ferranti and A. Tompkins, 2006: Response to the summer of 2003 Mediterranean SST anomalies over Europe and Africa. *J. Climate*, **19**, 5439-5454.
Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woolen, Y. Zhu, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K.C. Mo,Ropelewski, C., J. Wang, A. Leetma, R. Reynolds, R. Jenne and D. Joseph, 1996. The NCEP/NCAR 40-year reanalysis project. *Bull. Amer. Meteorol. Soc.*, **77**, 437-471.
Kanamitsu M., W. Ebisuzaki, J. Woollen, S.K. Yarg, J.J. Hnilo, M. Fiorino and G.L. Potter, 2002. NCEP/DOE AMIP-II reanalysis (R-2). *Bull. of Amer. Met. Soc .,* **83**, 1631-1643.

Knight, J.R., C.K. Folland and A.A. Scaife, 2006: Climate impacts of the Atlantic multidecadal oscillation. *Geophys. Res. Lett*., **33**, doi:10.1029/2006GL026242.

Kiladis, G.N. and K.M. Weickmann, 1997. Horizontal structure and seasonality of large-scale circulations associated with submonthly tropical convection. Mon. Wea. Rev., 125, 1997-2013.

Kiladis, G.N., C.D. Thorncroft and N.M.J. Hall, 2005. Three-Dimensional Structure and Dynamics of African easterly waves. Part I: Observations. *J. Atmos. Sci*., 63, 2212-2230

Knippertz, P. and A. H. Fink, 2008: Dry-Season Precipitation in Tropical West Africa and its Relation to Forcing from the Extratropics. *Monthly Weather Review*. 136, 3579–3596 DOI: 10.1175/2008MWR2295.1

Knippertz, P. 2007: Tropical–extratropical interactions related to upper-level troughs at low latitudes. *Dynamics of Atmospheres and Oceans*. 43 (2007) 36–62. doi:10.1016/j.dynatmoce.2006.06.003

Knutson, T. R., D. L. Delworth, K. W. Dixon and R. J. Stouffer, 1999: Model assessment of regional temperature trends (1949-1997). *J. Geophys. Res*., **104**, 30,981–30,996.

Lafore, J.-P., and M. W. Moncrieff, 1989: A numerical investigation of the organization and interaction of the convective and stratiform regions of tropical squall lines. *J. Atmos. Sci.,* 46, 521–544

Laing, A.G. and J.M. Fritsch, 1993. Mesoscale convective complexes in Africa*. Mon. Wea. Rev.,* 121, 2254,-2263.

Landsea, CW, Gray WM (1992) The strong association between western Sahel monsoon rainfall and intense Atlantic hurricanes. J. Climate, 5:435-453.

Lamb, P., 1978a: Large scale tropical Atlantic surface circulation patterns associated with sub-Saharan weather anomalies. *Tellus*, **30**, 240-251.

Lamb, P., 1978b: Case studies of tropical Atlantic surface circulation patterns during recent sub-Saharan weather anomalies: 1967 and 1968. *Mon. Wea. Rev*., **106**, 482-491.

Lamb, P., 1983: West African water vapour variations between recent contrasting Subsaharan rainy seasons. *Tellus*, **A35**, 198-212.

Lamb, P. and R. A. Peppler, 1991: Teleconnections linking worldwide climate anomalies. M. Glantz, R. Katz and N. Nicholls, Eds., Cambridge University Press, 121-189.

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

Lau, K.-H., and Lau N.-C., 1990: Observed structure and propagation characteristics of tropical summertime synoptic scale disturbances. Mon. Wea. Rev., 118, 1888–1913.

Lau, K. M., S. S. P. Shen, K.-M. Kim, and H. Wang, 2006: A multimodel study of the twentieth-century simulations of Sahel drought from the 1970s to 1990s. *J. Geophys. Res*., **111**(D07111).

Lavaysse C., C. Flamant, S. Janicot, J.-P. Lafore, D. Parker, and J. Pelon, 2008: Seasonal cycle of the West African heat low: a climatological perspective. *Clim. Dyn.,* DOI: 10.1007/s00382-009-0553-4.

Lavender, S.L. and A. J. Matthews, 2009: Response of the African monsoon to the Madden-Julian oscillation. *J. Climate*, doi:10.1175/2009JCLI2773.1.

Le Barbe, T. Lebel, and D. Tapsoba (2002) Rainfall variability in West Africa during the years 1950–90. *J. Climate*, **15**, 187–202.

Leroux S. and N. M. J. Hall 2009: On the relationship between African Easterly Waves and the African Easterly Jet. *J. Atmos. Sci.,* 66, 14 pages. DOI: 10.1175/2009JAS2988.1.

Liebmann B. and C.A. Smith, 1996. Description of a complete (interpolated) outgoing longwave radiation dataset. *Bull. of Amer. Met. Soc.,* **77**, 1275-1277.

Louvet, S., Fontaine B. and P. Roucou (2003) Active phases and pauses during the installation of the West African monsoon through 5-day CMAP rainfall data (1979-2001). *Geophys. Res. Lett*., **30**, 10.1029/2003GL018058.

Lothon, M., F. Saïd, F. Lohou and B. Campistron, 2008. Observation of the diurnal cycle of the low troposphere of West Africa. *Mon. Wea. Rev*., 136, 3477-3500.

Maloney, E.D. and J. Shaman, 2008. Intraseasonal variability of the West African monsoon and Atlantic ITCZ. *J. Climate*, **21**, 2898-2918.

Mathon V., H. Laurent and T. Lebel, 2002. Mesoscale convective systems rainfall in the Sahel. *J. Applied Meteor.,* 4**1**, 1081-1092.

Mathon, V. and H. Laurent, 2001. Life cycle of Sahelian mesoscale convective cloud systems. *Quart. J. Roy. Met. Soc.,* **127**, 377-406.
Matsuno T., 1966. Quasi-geostrophic motion in the equatorial area. *J. Met. Soc. Japan*, **44**, 25-42.
Matthews M., 2004. Intraseasonal variability over tropical Africa during northern summer. *J. Climate*, **17**, 2427-2440.
Mekonnen, A., C.D. Thorncroft, A.R. Aiyyer and G.N. Kiladis, 2009. Convectively coupled Kelvin waves over tropical Africa during the boreal summer: structure and variability. *J. Climate*, doi: 10.1175/2008JCLI2008.1
Milliff, R.F. and R.A. Madden, 1996. The existence and vertical structure of fast, eastward moving disturbances in the equatorial troposphere. *J. Atmos. Sci.,* 53, 586-597.
Moncrieff, M.W. and So, D.W.K 1989. A hydrodynamical theory of conservative bounded density currents. J. Fluid Mech., 198, 177-197.
Mounier, F., and S. Janicot, 2004. Evidence of two independent modes of convection at intraseasonal timescale in the West African summer monsoon. *Geophys. Res. Lett.,* **31**, L16116.
Mounier, F., G. Kiladis et S. Janicot, 2007 : Analysis of the dominant mode of convectively coupled Kelvin waves in the West African monsoon. *J. Climate***, 20***,* 1487-1503.
Mounier, F. S. Janicot and G. Kiladis, 2008: The West African monsoon dynamics. Part III: The quasi-biweekly zonal dipole. *J. Climate,* 21, 1911-1928.
Nicholson, S.E. 1978: Climatic variations in the Sahel and other African regions during the past five centuries. *J. Arid Environ*., 1, 3-34.
Nicholson, S. E., 1980: The nature of rainfall fluctuations in subtropical West Africa. *Mon. Wea. Rev*., **108**, 473-487.
Okumura and Xie (2004) Interaction of the Atlantic Equatorial Cold Tongue and the African Monsoon. *Journal of Climat*e, **17**, 3589-3602.
Palmer, T. N., 1986: Influence of the Atlantic, Pacific and Indian oceans on Sahel rainfall. *Nature*, **322**, 251-253.
Palmer, T., C. Brankovic, P. Viterbo, and M. J. Miller, 1992: Modeling interannual variations of summer monsoons. *J. Climate*., **5**, 399-417.
Parker, D. J, C. D. Thorncroft, R. R. Burton, and A. Diongue-Niang., 2005: Analysis of the African Easterly Jet using aircraft observations from the JET2000 experiment, *Q. J. R. Meteorol. Soc.,* 131, 1461-1482
Parsons, D. B., K. Yoneyama, and J. L. Redelsperger, 2000: The evolution of the tropical western Pacific atmosphere–ocean system following the arrival of a dry intrusion, QJRMS, 126, 517-548.
Peyrillé Ph., J.-P. Lafore, and J.-L. Redelsperger, 2007: An Idealized Two-Dimensional Framework to Study the West African Monsoon, Part I: Validation and Key Controlling Factors. *J. Atmos. Sci*., Vol. 64, No. 8, p2765-2782.
Peyrillé Ph. and J.-P. Lafore, 2007: An Idealized Two-Dimensional Framework to Study the West African Monsoon, Part II: Role of Large Scale Forcings and Characterization of the Diurnal Cycle. J. Atmos. Sci., Vol. 64, No. 8, 2783-2803.
Peyrillé Ph. 2006: Etude idéalisée de la mousson de l'Afrique de l'Ouest à partir d'un modèle numérique bidimensionnel. PhD dissertation, January 2006, Paul Sabatier University – Toulouse – France.
Philippon N and Fontaine B, 2002: The relationship between the Sahelian and previous second Guinean rainy seasons: a monsoon regulation by soil wetness? *Ann. Geoph*. **20**, 575-582.
Pohl, B., S. Janicot, B. Fontaine and R. Marteau, 2009 : Implication of the Madden-Julian Oscillation in the 40-day variability of the West African monsoon, and associated rainfall anomalies. *J. Climate*, doi :10.1175/2009JCLI2805.1
Pospichal B. and S. Crewell, 2007: Boundary Layer Observations in West Africa using a novel microwave radiometer, Meteorologische Zeitschrift, Vol. 16, No. 5, 513-523.
Pytharoulis, I. and Thorncroft C., 1999. The Low-Level Structure of African Easterly Waves in 1995. *Mon Wea Rev* **127** 2266-2280.
Ramel, R., H. Galle, and C. Messager, 2006. On the northward shift of the west african monsoon. *Climate Dynamics*, 26(4):429–440.
Redelsperger, J-L., C. D. Thorncroft, A. Diedhiou, T. Lebel, D. J. Parker, and J. Polcher, 2006: African Monsoon Multidisciplinary Analysis: An International Research Project and Field

Campaign', *Bull. Amer. Meteor. Soc.,* 87, 1739-1746.
Reed, R.J., D.C. Norquist and E.E. Recker, 1977. The structure and properties of African wave disturbances as observed during Phase III of GATE. *Mon. Wea. Rev.,* 105, 317-333.
Reed, R.J., A. Hollingsworth, A. Heckley, and F. Delsol, 1988a: An evaluation of the performance of the ECMWF operational system in analyzing and forecasting easterly wave disturbances over Africa and the tropical Atlantic Ocean. Mon. Wea. Rev., 116, 824–865.
Reed, R.J., E. Klinker, and A. Hollingsworth, 1988b: The structure and characteristics of African easterly waves disturbances as determined from ECMWF operational analysis/forecast. Meteor. Atmos. Phys., 38, 22–33.
Reiner, A., 1999: Ausgewählte Eigenschaften von "African Easterly Waves" im Sahel-Feuchtjahr 1988 und Sahel-Trockenjahr 1990. Selbstverlag des Instituts für Geophysik und Meteorologie der Universität zu Köln, 88pp.
Reiner, A., 2003: Synoptische Wellenstörungen und meso-skalige Konvektionsniederschläge in der westafrikanischen Sahel- und Sudanzone - Analyse mittels satellitengestützter Mikrowellen-Niederschlagsschätzung. Mitteilungen aus dem Institut für Geophysik und Meteorologie der Universität zu Köln (Hrsg. M. Kerschgens, F. M. Neubauer, M. Pätzold, P. Speth, B. Tezkan), No. 154, 167pp.
Roca, R., J.-P. Lafore, C. Piriou, et J.-L. Redelsperger, 2005 : Extratropical dry-air intrusions into the west african monsoon midtroposphere : An important factor for the convective activity over the sahel., *J. Atmos. Sci*., 62, 390-407.
Rotunno, R. J.B. Klemp, and M.E. Weisman 1988. A theory of long-lived squall lines. J. Atmos. Sci., 45, 463-485.
Rotstayn, L. D. and U. Lohmann, 2002: Tropical rainfall trends and the indirect aerosol effect. *J. Climate*, **15**, 2103–2116.
Rowell, D. P., 2001: Teleconnections between the tropical Pacific and the Sahel. *Q. J. R. Met. Soc*., **127**, 16831706.
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. *Q. J. R. Meteorol. Soc*., **121**, 669-704.
Rowell, 2003: The impact of Mediterranean SSTs on the Sahelian rainfall season. *J. Climate*, 16, 849–862.
Sadler, J.C., and L.K. Oda, 1978: The synoptic (A) scale circulations during the third phase of GATE, 20 August 23 August 1974. Dept. of Meteorology, University of Hawaii, Rep. UHMET 78-02, 41pp.
Sadler, J.C., and L.K. Oda, 1979: The synoptic (A) scale circulations during the second phase of GATE, 17 July 19 August 1974. Dept. of Meteorology, Univer¬sity of Hawaii, Rep. UHMET 79-14, 36pp.
Sadler, J.C., and L.K. Oda, 1980: GATE Analyses. I. The synoptic (A) scale circulations during Phase I, 26 July 16 July 1974. II. Means for phases I, II, and III. Dept. of Meteorology, University of Hawaii, Rep. UHMET 80-01, 32pp.
Schubert, W.H., Ciesielski, P.E., Stevens, D.E., and Kuo, H.C. 1991: Potential vorticity modelling of the ITCZ and Hadley Circulation, J.Atmos. Sci., 48 1493-1509.
Sherwood, S. C., 1999: Convective precursors and predictability in the tropical western Pacific. *Mon. Wea. Rev.,* **127,** 2977–2991.
Slingo A., N.A. Bharmal, G.J. Robinson, J.J. Settle, R.P. Allan, H.E. White, Environmental, P.J. Lamb, M. Issa Lélé, and D.D. Turner, 2008: Overview of observations from the RADAGAST experiment in Niamey, Niger. Part 1: Meteorology and thermodynamic variables. *J. G. R.-Atmos.*, 113, D00E01.
Slingo A, White HE, Bharmal NA, et al., 2009: Overview of observations from the RADAGAST experiment in Niamey, Niger. Part 2: Radiative fluxes and divergences. *J. G. R.-Atmos.*, 114, D00E04.
Stott, P. A., S. B. F. Tett, G. S. Jones, M. R. Allen, J. F. B. Mitchell and G. J. Jenkins, 2000: External control of 20th century temperature by natural and anthropogenic forcings. *Science*, **290**, 2133–21–36.
Straub K. H. and G.N. Kiladis, 2002. Observations of convectively coupled Kelvin waves in the eastern Pacific ITCZ. *J. Atmos. Sci.,* **59**, 30-53.

Sultan B. and S. Janicot, 2003. The West African monsoon dynamics. Part II: the “preonset” and “onset” of the summer monsoon. *J. Climate,* **16**, 3389-3406

Sultan B., S. Janicot and A. Diedhiou, 2003. The West African monsoon dynamics. Part I: Documentation of intraseasonal variability. *J. Climate,* **16**, 3407-3427

Sultan, B., S. Janicot and S. Correia 2009 : Medium-lead prediction of intraseasonal oscillations in West Africa. Accepted *Weather and Forecasting*, Special Issue « West African weather prediction and predictability ». doi :10.1175/2008WAF2222155.1

Sutton, R.T. and D.L.R. Hodson, 2005: Atlantic ocean forcing of north American and European summer climate. *Science*, **309**, 115-118.

Takayabu, Y., 1994. Large-scale cloud disturbances associated with equatorial waves. Part I: Spectral features of the cloud disturbances. *J. Met. Soc. Japan,* **72**, 43.

Taylor C. M., and T. Lebel, 1998: Observational evidence of persistent convective-scale rainfall patterns. *Mon.Wea. Rev.*, **126**, 1597-1607.

Taylor, C.M., 2008: Intraseasonal land-atmosphere coupling in the West African monsoon. *J. Climate*, **21**, 6636-6648.

Thomas, R.A. and P.J. webster, 1997 : The role of inertial instability in determining the location and strength of near-equatorial convection. *Quart. J. Roy. Met. Soc.*, **123**, 1445-1482.

Thorncroft, C. D., and M. Blackburn, 1999: Maintenance of the African easterly jet. *Q. J. R. Meteorol. Soc.*, 125, 763-786.

______, D. J. Parker, R. R. Burton, M. Diop, J. H. Ayers, H. Barjat, S. Devereau, A. Diongue, R. Dumelow, D. Kindred, N. M. Price, M. Saloum., C. M. Taylor, and A. M. Tompkins, 2003: The JET2000 Project: Aircraft Observations of the African Easterly Jet and African Easterly Waves. BAMS 2003, **84**, 337-351.

Thorncroft, C.D., J-P Lafore, G. Berry, R. Roca, F. Guichard, M. Tomasini, N. Asencio, 2007: Overview of African weather systems during the summer 2006. AMMA special issue for *CLIVAR Newsletter Exchanges*.

Thorncroft, C.D., Hall, N.M.J., and Kiladis, G.N. 2008 Three Dimensional Structure and Dynamics of African Easterly Waves Part III: Genesis. *J. Atmos. Sci.,* 65, 3596–3607 DOI: 10.1175/2008JAS2575.1

Vizy, E.K. and K.H. Cook, 2009: A mechanism for African monsoon breaks: Mediterranean cold air surges. *J. Geophys. Res.*, **114**, D01104, doi:10.1029/2008JD010654.

Ward, M. N., 1998: Diagnosis and short-lead time prediction of summer rainfall in tropical North Africa at interannual and multi-decadal time scales. *J. Climate*, **11**, 3167-3191.

Wheeler, M. and Kiladis G. N., 1999. Convectively coupled equatorial waves: analysis of clouds and temperature in the wavenumber-frequency domain. *J. of Atmos. Sci.,* **56**, 374-399.

Zheng, X., A. B. Eltahir, and K. A. Emanuel, 1999: A mechanism relating tropical Atlantic spring sea surface temperature and west African rainfall. *Q. J. R. Meteorol. Soc.*, **125**, 1129-1163.