QUARTERLY JOURNAL OF THE ROYAL METEOROLOGICAL SOCIETY
Q. J. R. Meteorol. Soc. **133**: 2065–2084 (2007)
Published online 20 November 2007 in Wiley InterScience
(www.interscience.wiley.com) DOI: 10.1002/qj.104

A physical basis for the interannual variability of rainfall in the Sahel

Sharon E. Nicholson[a]* and Peter J. Webster[b]
[a] *Department of Meteorology, Florida State University*
[b] *School of Earth & Atmospheric Sciences & Civil and Environmental Engineering, Georgia Institute of Technology*

ABSTRACT: A major factor in rainfall variability over Sahelian West Africa is the latitudinal location of the tropical rainbelt. When it is displaced abnormally far northward, the Sahel experiences a wet year. An anomalous southward displacement results in drought. In this paper we examine the question of what controls the location during the boreal summer, hypothesizing that inertial instability plays a role. An analysis of surface pressure and temperature fields, wind fields, divergence and vertical motion show that the criteria for inertial instability are satisfied in wet Augusts but not in dry ones. The key determinant appears to be the surface pressure gradient between the continent and the equatorial Atlantic. When this is large, inertial instability results in the development of a low-level westerly jet. The presence of this jet enhances the horizontal and vertical shear, and displaces the African Easterly Jet northwestward. Associated with this situation is strong vertical motion over the Sahel and subsidence over the Guinea Coast, producing dry conditions over the latter. The result is a rainfall dipole, one of two major modes of variability over West Africa. Important factors include sea surface temperatures (SSTs) in the equatorial Atlantic and pressure in the South Atlantic. The first of these factors suggests a link with the Atlantic Niño mode of tropical Atlantic variability; while the second suggests a possible link with the Pacific and the extratropical South Atlantic. Overall, our study relates the well-known SST influence on Sahel rainfall to atmospheric dynamics over the continent. Copyright © 2007 Royal Meteorological Society

KEY WORDS rainfall; interannual variability; Sahel; Africa; tropical dynamics

Received 11 October 2005; Revised 25 April 2007; Accepted 30 April 2007

1. Introduction

One of the most dramatic recent cases of abrupt climate change occurred in the West African Sahel in the late 1960s (Figure 1). Rainfall has been below the century-long mean almost every year since 1968. This dry period is evident throughout the region (10 °N–18 °N, 15 °W–30 °E), which we will call the 'Sahel'. Over much of this area, the 30-year means, which provide the standard 'climatological normal', have declined by 30%–40%.

The causes of this change have not yet been unequivocally determined. A considerable body of evidence identifies changes in sea surface temperatures (SSTs) as a major factor. Changes in the global oceans clearly promote the long-term trends in the region, and probably govern some of the year-to-year changes as well (e.g. Lamb, 1978a, 1978b; Hastenrath, 1990; Ward, 1998; Rowell *et al.*, 1992, 1995; Rowell, 2001; Giannini *et al.*, 2003). Studies of wind and moisture fields, surface circulation and the intertropical convergence zone (ITCZ) have identified the atmospheric circulation features associated with wet and dry conditions over the Sahel (e.g. Kanamitsu and Krishnamurti, 1978; Newell and Kidson, 1984; Janicot, 1992b; Fontaine and Janicot, 1992; Fontaine *et al.*, 1995; Grist and Nicholson, 2001; Nicholson and Grist, 2001; Grist, 2002).

Most studies have concentrated on evaluating either the impact of 'remote' marine forcing and associated global teleconnections, such as the El Niño Southern Oscillation (ENSO), or 'local' forcing induced by these circulation features. Few studies have attempted to relate the two effects and to determine the specific ways in which the ocean's boundary forcing modifies atmospheric circulation over the continent. This study attempts to couple the remote and local forcing via a mechanism proposed by Tomas and Webster (1997) to account for interannual variability of off-equatorial convection, such as occurs in the Sahel during the boreal summer.

The tropical Atlantic is highly sensitive to meridional, especially cross-equatorial, SST gradients (e.g. Wagner, 1996; Carton *et al.*, 1996; Chang *et al.*, 2000). By influencing surface pressure gradients, the SST gradients modulate the location of the marine ITCZ and thereby influence rainfall in Africa and South America (Hastenrath and Greischar, 1993a, 1993b). Tomas and Webster (1997) argue that cross-equatorial pressure gradients in the lower troposphere render the system inertially unstable. In this paper, we postulate specifically that SST gradients determine, via inertial instability, the presence or absence of

* Correspondence to: Sharon E. Nicholson, Department of Meteorology, Florida State University, Tallahassee, FL 32306, USA.
E-mail: sen@met.fsu.edu

Copyright © 2007 Royal Meteorological Society

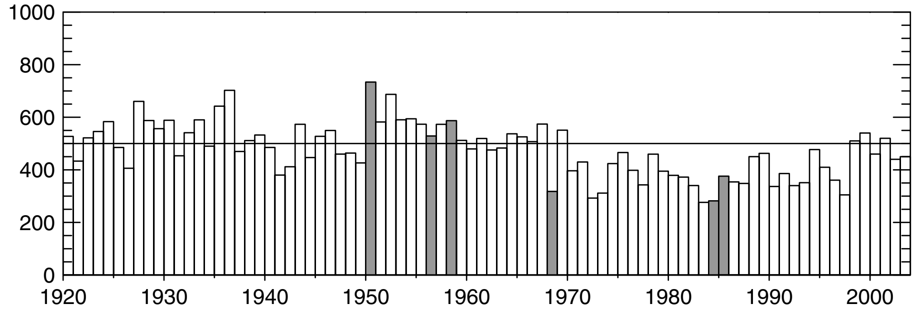

Figure 1. Long-term rainfall variability in the West African Sahel: annual rainfall in mm, from (Nicholson, 2005). The years selected for analysis in the present study are shaded.

a strong low-level westerly equatorial jet just north of the Equator. We further argue that the development of such a westerly jet displaces the African Easterly Jet (AEJ) northward into the Sahel. The AEJ overlies West Africa in the boreal summer and has a core in the mid-troposphere, with speeds of the order of 10 ms^{-1}. This feature is associated with the African Easterly Waves that produce rainfall (e.g. Burpee, 1972; Reed *et al.*, 1977). The effect of its northward displacement is a simultaneous displacement of the rainbelt into the Sahel (Nicholson, 2007b). Although this produces wet conditions in the Sahel, abnormal dryness develops to the south along the Guinea Coast.

Our analysis will show that this mechanism is consistent with the spatio-temporal patterns of rainfall variability in the region (Section 2.1) and with observed circulation changes associated with these patterns (Section 2.2). The opposition between the Sahel and Guinea Coast is one of two major modes of rainfall variability in the region. The second mode, ubiquitous dry or wet conditions, is the subject of a companion paper (Nicholson, 2007a), which examines the factors that control the intensification and weakening of the rainbelt.

Our hypothesis is tested in the context of a new paradigm for Sahel rainfall variability that connects the impact of marine forcing and the forcing via local circulation features (see Section 3.1). It provides a framework that can readily incorporate the rapidly-expanding body of knowledge about the role played by the oceans and the African land surface.

The paper is structured as follows. Section 2 reviews the literature on the relationships between Sahel rainfall, atmospheric circulation, and SSTs. Section 3 summarizes the new paradigm and describes theoretical work on off-equatorial convection, focusing on the mechanism described by Tomas and Webster (1997). Section 4 describes the hypothesis in greater detail, including the past observations that have led to its formulation. Section 5 discusses data and methodology. In Section 6 we test the hypothesis, including the links to SSTs, by contrasting wet and dry years in the Sahel. Section 7 discusses the generation of the low-level westerly wind maximum, and relates this to the tropical Atlantic variability. In Section 8 we summarize our results and their implications, including the potential importance of this mechanism in producing the protracted dry conditions of the past three decades.

2. Observational background

2.1. Nature of rainfall variability in the region

The most relevant aspects of rainfall variability for the present study are the spatial patterns and spatial scale of rainfall anomalies in the boreal summer. It has long been recognized (e.g. Nicholson, 1986) that the spatial extent of the anomalies in the most extreme years is roughly equivalent to the area between the Sahara and the Guinea Coast of the Atlantic, extending inland from the West Coast at least as far as Chad and the western Sudan around 25 °E. There are few areas of the planet in which interannual variability is so spatially coherent.

Because of this spatial coherence, two basic spatial patterns (see Figure 2) describe most of the rainfall variability (Nicholson, 1980, 1981, 1986; Janowiak, 1988; Janicot, 1992a; Nicholson and Palao, 1993; Moron, 1994; Fontaine and Janicot, 1996; Giannini *et al.*, 2003). The most common is a dipole (The appropriateness of the term 'dipole' in this situation has sometimes been challenged. It is retained here because its use has become conventional in the literature dealing with African rainfall variability, and because its use facilitates and abbreviates the discussion.), in which rainfall anomalies of opposite signs prevail in equatorial and subtropical sectors (i.e. the sub-humid Guinea Coast and the semi-arid Sahel zone further north). These regions can be identified in Figure 3. The node of the dipole is rather consistently near 10 °N, and the transition is so abrupt that, in many years, station anomalies either north or south of that latitude are almost all of the same sign (Nicholson, 1980). The second pattern is characterized by anomalies of the same sign across these regions, and can be dry or wet. The net result is four distinct configurations of rainfall anomalies. Different patterns of surface conditions and atmospheric circulation are associated with different configurations (Janicot 1992b; Fontaine and Janicot 1996).

Nicholson and Grist (2001) have proposed that the causes of the dipole and non-dipole patterns are fundamentally different. The dipole is associated with factors that influence the position of the tropical rainbelt, while

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

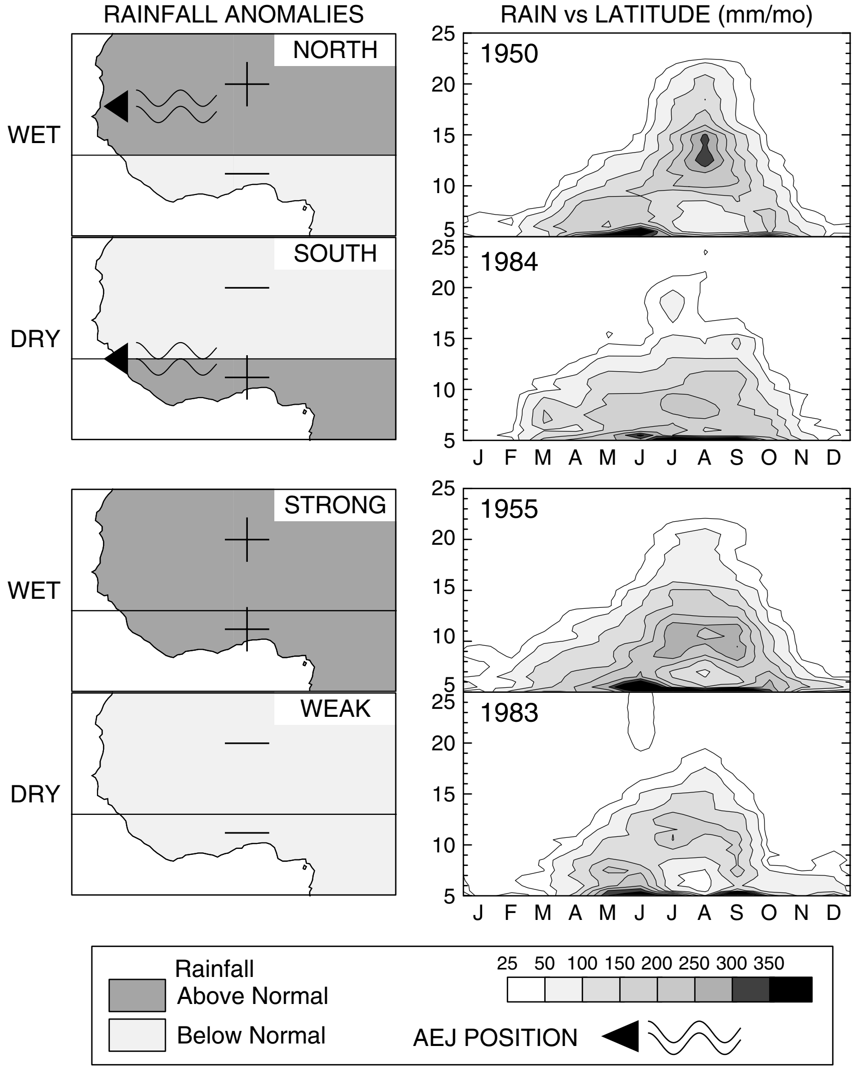

Figure 2. Schematic of the conceptual model proposed in (Nicholson and Grist, 2001) and (Nicholson, 2007b) for the boreal rainy seaon. Rainfall anomaly types (left): dark and light shading indicate above- and below-normal rainfall, respectively. The typical latitudinal location of the AEJ in August is also noted for the wet-dipole year (1950) and the dry-dipole year (1984). Structure of the rainbelt and its seasonal movement (right): rainfall as a function of month and latitude is indicated for four years corresponding to the anomaly patterns to the left. Rainfall variability in the Sahel is a result of a change in the intensity or location of the tropical rainbelt, the latter being related to a change in the location of the AEJ.

the non-dipole pattern is associated with factors that influence its intensity. Nicholson (2007b) tests this hypothesis for the four multi-year composites corresponding to the spatial configurations shown in Figure 2. These are based on rainfall anomalies in the boreal summer months of July through September. The results validate the dichotomy proposed by Nicholson and Grist (2001), and show, as hypothesized, that a latitudinal displacement of the tropical rainbelt is a major factor in the dipole patterns, but not in the non-dipole patterns.

2.2. Atmospheric circulation changes in wet and dry years

Grist and Nicholson (2001) identify the changes in the general circulation over Africa associated with wet and dry years in the Sahel, by contrasting four wet years (1958–1961) and four dry years (1982–1985), emphasizing zonal wind cross-sections that depict the upper-tropospheric Tropical Easterly Jet (TEJ), the AEJ, and the low-level equatorial westerlies. The TEJ, with a core at roughly 150 mb, extends across Asia and Africa, and results from the temperature contrast between the Himalayan plateau and the Indian Ocean to the south. The AEJ, forced by the temperature gradient between the Sahara and the Atlantic, is generally around 650 mb. The equatorial westerlies prevail just above the surface, and result in part from the Southern-Hemisphere trades as they cross the Equator.

The most pronounced circulation contrast between wet and dry years in the Sahel is the development, in wet years, of a strong westerly jet-like flow in the lower troposphere just north of the Equator. Other notable changes include a strengthening (weakening) of the TEJ, a weakening (strengthening) of the AEJ, and a northward (southward) displacement of the AEJ, in wet (dry)

Copyright © 2007 Royal Meteorological Society
DOI: 10.1002/qj

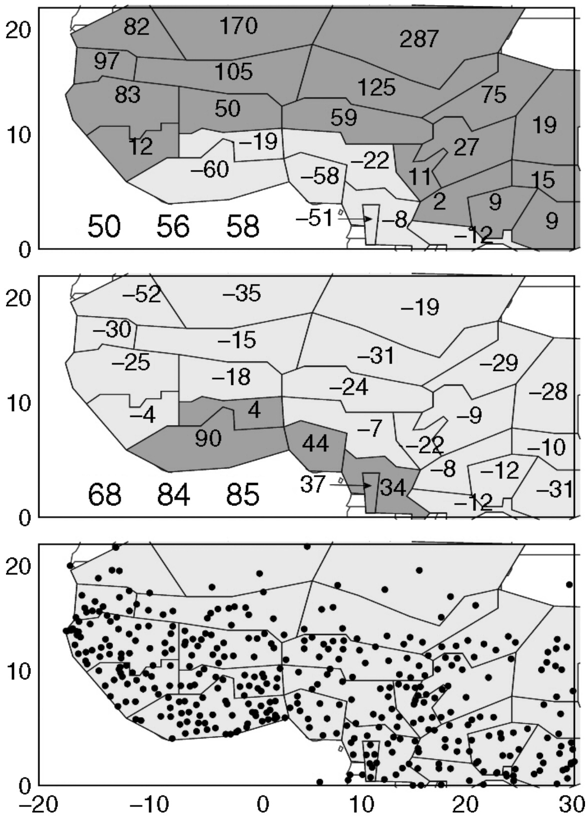

Figure 3. Mean August rainfall over West Africa for the wet-dipole (top map) and dry-dipole (middle-map) composites (expressed as a regionally-averaged percentage standard departure, based on means for the period 1968–1997). The dots indicate rainfall stations.

years. Although these results are obtained from observational analyses of the NCEP/NCAR Reanalysis dataset, an examination of radiosonde and pibal data shows qualitative agreement with each of the aforementioned points. Nicholson and Grist (2001) obtain similar results by analysing multi-decadal variability in the Sahel.

The studies cited above support the conclusions of many earlier studies on rainfall variability in West Africa (e.g. Kidson, 1977; Newell and Kidson, 1984; Kanamitsu and Krishnamurti, 1978; Fontaine and Janicot, 1992; Fontaine *et al*., 1995). However, they also draw several new conclusions concerning the causes of this variability. Perhaps the most important is that the latitude of the AEJ, rather than its strength, is of critical importance in the northern Sahel. Baroclinic instability is strong in the Sahel when the jet is displaced northward in the wet years, but it is absent in this region in the dry years. The studies mentioned above show a consistent association between an anomalously strong jet in dry years and an anomalously weak jet in wet years. Newell and Kidson (1984), however, raise the possibility that the stronger jet is an effect, rather than a cause, of the reduced rainfall. Their observation that the meridional temperature gradient is stronger in dry years is consistent with this interpretation.

Our results likewise suggest that the stronger AEJ in dry years is a consequence of other changes. However, our explanation is quite different, suggesting an additional mechanism governing the AEJ's strength. It involves the equatorial westerlies, a system that our research identifies as a critical factor in rainfall variability. Our interpretation is that the strength and location of the AEJ are dictated not only by the temperature gradient, which produces a vertical shear of a particular magnitude, but also by the westerlies, upon which the shear is imposed.

The westerlies appear to be pivotal in determining the latitude of the AEJ, and hence of the tropical rainbelt. The development of a strong westerly jet in the West African monsoon flow displaces the AEJ poleward, into a more northern track over the Sahel (see Section 3.1). This is the configuration associated with the 'Sahel wet' dipole pattern, and to a lesser extent with the ubiquitous wet conditions in the Sahel and Guinea Coast. By implication, the westerly jet in the monsoon flow is a key factor in the development of wet conditions in the Sahel.

2.3. Links with SST distributions

Observational studies have shown that consistent patterns of SST variations are associated with years of contrasting rainfall conditions in the Sahel (e.g. Lamb, 1978a, 1978b; Hastenrath, 1984; Lough, 1986; Palmer, 1986; Wolter, 1989; Folland *et al*., 1986, 1991; Lamb and Peppler, 1992; Janicot, 1992b; Fontaine and Bigot, 1993; Fontaine and Janicot, 1996; Fontaine *et al*., 1998). These relationships hold for both annual and interdecadal time-scales (Hastenrath, 1990; Ward, 1998). Modelling studies provide further support for this (e.g. Folland *et al*., 1986, 1991; Rowell *et al*., 1992, 1995; Semazzi *et al*., 1996; Giannini *et al*., 2003).

Some of the earliest papers dealing with causes of rainfall variability in the Sahel focused on SSTs in the Atlantic alone. The so-called 'SST dipole' (e.g. Lamb, 1978a; Fontaine *et al*., 1998; Janicot *et al*., 2001) has stood the test of time as a major influence on this variability. This dipole consists of anomalously warm (cold) conditions north of 10 °N and cold (warm) anomalies further south, in the equatorial Atlantic. Drought in the Sahel is associated with abnormally low temperatures in the north and abnormally high temperatures to the south during the boreal summer. Nicholson (2007a) emphasizes the more direct role of SSTs specifically in the Gulf of Guinea and the equatorial Atlantic. This is essentially the southern half of Lamb's SST dipole.

Other studies take a broader view. Some emphasize the more general inter-hemispheric temperature contrast (Folland *et al*., 1986, 1991; Servain, 1991). This contrast is consistent with the SST dipole described by Lamb, and it provides the basis of one of the two major modes of tropical Atlantic variability (Mehta, 1998; Ruiz-Barradas *et al*., 2000). However, the inter-hemispheric mode is strongest on decadal time-scales, and is most developed in the boreal spring, well before the Sahel rainy season. Ward (1998) and Rowell (2001) suggest that Pacific Ocean temperatures are likewise important: a conclusion also supported by the studies of Fontaine *et al*. (1998)

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

and Janicot *et al*. (2001). Earlier studies by Folland *et al*. (1986) and Palmer (1986) similarly concluded that the Pacific played a role, but found much weaker statistical links than those established by Ward and Rowell.

3. Understanding rainfall variability in the Sahel

3.1. A new paradigm

A new paradigm, based on the results of (Grist and Nicholson, 2001; Grist *et al*., 2002), is developed, in which 'local' factors such as the location of the AEJ and associated shears are deemed to be very important. The position of the AEJ helps to distinguish between a wet and a dry regime in the Sahel. Other factors influencing the intensity of the rainbelt determine whether the pattern is a dipole or one with ubiquitous negative or positive anomalies in the Sahel and Guinea regions. Past attempts to explain the long-term variability of Sahel rainfall have emphasized other factors, such as the position and intensity of the ITCZ, the Hadley and Walker circulations, and the strength of the AEJ. Our paradigm provides an alternative, but complementary, framework for investigating and understanding rainfall variability in this region.

The essence of our paradigm might be viewed as a dichotomy related to the mean position of the AEJ. In one mode, which will be termed the 'Sahel' mode to facilitate discussion, the jet lies in its poleward 'track' over the Sahel, with a core around 14°–18 °N during the month of August (see Figure 2). This location most commonly produces the configuration of rainfall anomalies represented by a dipole in which the Sahel is anomalously wet and the Guinea Coast is anomalously dry. As noted in Section 2.2, strong, deep and extensive equatorial westerlies and a strong vertical shear zone over the Sahel characterize the Sahel mode (Grist and Nicholson, 2001; Nicholson *et al*., 2007). The shear indicates that baroclinic instability is probably important in the Sahel in the wet years. In the opposite phase of the dipole – termed the 'Guinea Coast' mode – the jet lies further equatorward (with its core around 10°–12 °N) and to the south of the Sahel. Barotropic instability is dominant over the Sahel. In this mode, either the 'dry' rainfall pattern or the 'dry dipole' pattern develops, depending on the overall intensity of the rainbelt.

This paradigm is appealing because it can serve to reconcile contradictory conclusions on some issues, such as the relationships between Sahel rainfall and the ITCZ or the moist layer (e.g. Nicholson, 1981; Long *et al*., 2000; Lamb, 1978a, 1983; Lamb and Peppler, 1992). If the paradigm is correct, a key question is what produces a shift in the position of the AEJ. We hypothesize that this shift is related to the development of a strong equatorial westerly wind maximum. The fundamental question then becomes the cause of this westerly maximum.

Low-level westerlies (often called the 'southwest monsoon flow') are a well-known feature of West African climate. They are associated with the humid air south of the ITCZ, and arise as the southeasterly trades cross the Equator and take on a westerly course. However, neither this effect nor thermal wind considerations accounts for the relative strength of the westerly maximum during the wet years compared to the dry years. Tomas and Webster (1997) provide an explanation for the existence of a westerly maximum in the vicinity of off-equatorial convection. This mechanism, and our reasons for suggesting it as a factor in interannual variability, are discussed in Section 3.3.

3.2. Off-equatorial convection

Despite the observed fact that the SST dipole in the Atlantic Ocean appears to influence the distribution of precipitation over West Africa (e.g. Lamb, 1978a), there have been few attempts to explain the physical processes connecting the two. It is clear that SSTs influence the ITCZ (and hence precipitation) over the ocean, but it is less clear how the dipole influences rainfall over the continent. An explanation of this requires an understanding of the physical processes that determine the position of the ITCZ.

Over West Africa, the ITCZ lies well to the north of the Equator, so that the relevant process is off-equatorial convection. Wang and Wang (1999) point out that meteorological explanations for this phenomenon, which results in a latitudinally-asymmetric climate, are generally based on atmospheric internal dynamics. Therefore the role of planetary-boundary-layer processes in organizing deep convection is emphasized (e.g. Charney, 1971; Wang and Li, 1994). One aspect of this is the reason why the ITCZ tends to form away from the Equator (Philander *et al*., 1996). Kraus (1977a, 1977b) argues that the ITCZ is a 'thermal equator', separating integral areas of equal heat receipt. Thus it moves in response to the heat imbalance between the hemispheres. Emanuel (1995) hypothesizes that the thermally-direct monsoon is generated via the breakdown of a radiative–convective equilibrium state, with sub-cloud-layer processes playing a major role.

Murakami *et al*. (1992) and Mitchell and Wallace (1992) argue that over the oceans of the western hemisphere the ITCZ is not determined by atmospheric internal dynamics alone. They point to the coupled variation of the ITCZ and the equatorial cold tongue (ECT): in particular that the annual march of the ITCZ varies in tandem with the variation of the underlying ECT. Thus, ocean–atmosphere interactions are a determining factor in the location of the ITCZ (Wang and Wang, 1999).

3.3. Inertial instability

The jet streams over West Africa are the source of African Easterly Waves (e.g. Burpee, 1972). If the locations of these jets are linked to changes in the SST distribution, then one might expect different loci of wave activity from year to year. Tomas and Webster (1997) note that the (generally co-located) maxima of low-level convergence and precipitation are not necessarily at the

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

same latitude as the (co-located) SST maximum and sea-level-pressure (SLP) minimum. In fact, they find that all four of these locations coincide only in regions of small cross-equatorial pressure gradient. In cases with a substantial cross-equatorial pressure gradient, the SST maximum and surface-pressure minimum are poleward of the maxima of low-level convergence and precipitation. Tomas and Webster attribute this to the instability of a substantial cross-equatorial flow, and the generation of a stabilizing secondary circulation in a conditionally-unstable atmosphere where absolute-vorticity advection balances the shear of the zonal wind.

Tomas and Webster (1997) thus offer a framework of physical processes that tie together the changes of SST in the Atlantic associated with the SST dipole and the variability of rainfall over Africa. Furthermore, the Tomas–Webster theory suggests changes to circulation patterns over West Africa: specifically, the generation of a low-level westerly wind maximum that is important in local dynamics. The theory shows that when the cross-equatorial pressure gradient is strong, such a westerly maximum can arise in the vicinity of off-equatorial convection, poleward of the region of zero absolute vorticity.

A point in support of the inertial-instability mechanism of Tomas and Webster (1997) is that it readily explains the latitude of the convection over West Africa and its interannual variability. In this region, convection is linked to African Easterly Waves that are triggered by instabilities associated with the AEJ (e.g. Burpee, 1971). The shallow surface wind convergence lies well to the north of the convective maximum produced by the waves. In the view of Holton *et al.* (1971), the ITCZ is a composite of such zonally-propagating synoptic disturbances, which serve to maximize boundary-layer convergence at a preferred latitude. In this view, the location of the ITCZ over West Africa is closely coupled to the location of the AEJ. The low-level westerly jet, which is predicted by the inertial-instability mechanism, dictates the location of the AEJ, in tandem with the latitudinal temperature gradient over the continent. Some of the other proposed mechanisms involving internal dynamics are consistent with this hypothesis. However, the thermodynamic hypotheses, which neglect the role of the AEJ (e.g. Eltahir and Gong, 1996; Zheng *et al.*, 1999), do not seem to be applicable over West Africa.

Tomas and Webster's preliminary analysis of ECMWF model data suggests that the prerequisites for the formation of such a westerly wind maximum are found over West Africa in the boreal summer. These prerequisites are:

- a strong cross-equatorial surface pressure gradient;
- a displacement of the zero absolute-vorticity contour into the summer hemisphere;
- a divergent wind maximum bisected by the zero absolute-vorticity contour;
- divergence equatorward of the contour;
- convergence poleward of the contour; and
- strong subsidence over the Equator, but rising motion and enhanced convection to the north of the zero absolute-vorticity contour.

In essence, the rising motion (and hence the region of convergence and precipitation) is a result of the secondary circulation, where cyclonic vorticity is generated through vortex-tube stretching to ameliorate the importation across the Equator of anticyclonic vorticity.

4. Hypothesis and goals

The basic question addressed here is what causes a shift between the Sahel mode and the Guinea Coast mode. Our hypothesis is that the development of a westerly equatorial jet forces this shift. The reason for the development of the jet may be that offered by Tomas and Webster (1997). If the cross-equatorial pressure gradient is strong, a westerly wind maximum is required to achieve a balance between the advection of absolute vorticity and the shear of the zonal wind.

Accordingly, the difference in the strength of the westerly maximum in the boundary layer in wet and dry years would relate to the difference in the advection of absolute vorticity between the two sets of years. Acceleration of the divergent wind in the region where the absolute vorticity is locally anticyclonic results in a convergence centre in the boundary layer that gives rise to convection.

We hypothesize further that the westerlies result, at least in part, from the SST patterns prevailing over the equatorial Atlantic in the 1950s. Studies of Lamb (e.g. 1978a) and others describe, for that period, a dipole, with anomalously warm conditions north of about 5 °N and cold anomalies further south, in the equatorial Atlantic. Lower SSTs to the south would presumably enhance the pressure gradient by increasing the thermal contrast between the Atlantic and the Sahara, with its relatively fixed temperatures.

This hypothesis is appealing because it can account for numerous characteristics of Sahel rainfall variability. It is consistent with the observation that the wet regime in the Sahel is related to an enhanced inter-hemispheric SST gradient that is particularly pronounced in the Gulf of Guinea. Such a situation would enhance the cross-equatorial pressure gradient south of the Sahel. Furthermore, the mechanism described by Tomas and Webster (1997) can explain the frequent occurrence of the dipole pattern with wet conditions in the Sahel and dry conditions in the equatorial region. This pattern represents simultaneous enhancement of off-equatorial convection and reduction of convection in the equatorial region, as predicted. The dry conditions of the Guinea Coast lie in the area where divergence would prevail, equatorward of the zero absolute-vorticity contour. Another observation that this mechanism could explain is the apparent decoupling in wet years between the westerly maximum aloft and the moist layer (Grist and Nicholson, 2001).

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

5. Data and methodology

The datasets used in this study consist of the author's African precipitation archive (Nicholson, 1986, 1993), Reynold's reconstructed SST dataset (Smith *et al.*, 1996), and the NCEP Reanalysis dataset (Kalnay *et al.*, 1996; Kistler *et al.*, 2001). A map of rainfall stations in the archive is given in Figure 3; most records extend from around 1920 to 1997. For each station, long-term means are calculated as an average over all years in the dataset. Thus, in most cases the mean represents 70–80 years. The NCEP dataset includes daily and monthly values of various dynamical parameters. For most variables, the dataset covers the years 1948–2005. The parameters to be utilized include wind speed, zonal wind, and vertical motion.

The use of NCEP Reanalysis data for historical analyses has been questioned by some. It particular, it has been suggested that a discontinuity occurred around 1968 or 1970 (Poccard *et al.*, 2000; Janicot *et al.*, 2001; Kinter *et al.*, 2004; Chelliah and Bell, 2004). This coincided with a discontinuity in the Sahel rainfall record: the onset of the multidecadal dry conditions (Nicholson, 1993). There appears to be a consensus that NCEP estimates of wind fields are relatively reliable, but that there are difficulties in tropical divergent circulations and rainfall (Poccard *et al.*, 2000; Kinter *et al.*, 2004). Nevertheless these data are routinely used in long-term studies. In such a study, Grist and Nicholson (2001) used West African pibal and rawinsonde reports to verify conclusions based on NCEP. Moreover, in other analyses utilizing NCEP (e.g. Grist and Nicholson, 2001; Nicholson and Grist, 2001, 2002), coherent and physically reasonable results have emerged from a mixture of data sources.

In this study, we primarily use what are called 'A variables': those strongly influenced by observational data, and hence most reliable (Kalnay *et al.*, 1996). These include, for example, wind and pressure fields. Less reliable are the B variables, whose derivation is about equally dependent on observations and modelling. We have reduced the possible error by calculating the B variables (divergence, divergent wind components, and vorticity) directly from NCEP winds. Calculating these variables offline allows us to examine them with greater vertical resolution and to extend the analysis back to 1950. Rainfall is a C variable (largely model-dependent), and is known to be unreliable over Africa (Poccard *et al.*, 2000). This problem was avoided through use of the first author's gauge archive.

To check whether the prerequisites of Tomas and Webster, listed above, are met during the boreal summer over West Africa, we examine surface pressure, 925 mb absolute vorticity, the divergent wind at 850 mb and 925 mb, divergence, and vertical motion. To determine whether or not this mechanism plays a role in interannual variability in the Sahel, wet and dry years over West Africa are compared and contrasted. In addition to the variables mentioned, meridional and zonal winds, convection, relative vorticity and SST are also evaluated.

The three months of July, August and September represent the core of the rainfall season, but this study focuses on August only. August is particularly interesting because rainfall variations in this month are responsible for the greatest proportion of the interannual variability. Our preliminary analyses further justify the focus on August alone. These show markedly persistent circulation anomalies in the three rainy-season months. This may be a consequence of the relatively low resolution of the data ($2.5^\circ \times 2.5^\circ$). Nevertheless, the preliminary analysis suggests that the use of August to represent the rainy season is an efficient approach that does not compromise much detail.

Calculations are performed for a three-year wet composite (1950, 1956, 1958) and a three-year dry composite (1968, 1984, 1985). The rainfall 'dipole' discussed in Section 2.1 develops in all six years. The first – wet Sahel and dry Guinea Coast – is henceforth referred to as the 'wet dipole', and the second – dry Sahel and wet Guinea Coast – as the 'dry dipole'. The six years were chosen subjectively from regional rainfall anomalies, according to how developed the dipole is and how large the anomalies are. Typical of the years comprising the composites are 1950, which was exceedingly wet in the Sahel, and 1984, one of the driest on record in the Sahel (Figure 1). Figures 3 and 4, respectively, depict regional rainfall anomalies and rainfall as a function of latitude for the two composites during the month of August.

The contrast between the two composites is striking. In both, the August anomalies show a well-developed dipole pattern (Figure 3), but the polarity is reversed between the wet and dry cases. Over the Sahel, regionally-averaged anomalies are of the order of 0.5 standard deviations in the dry-dipole case, but up to 2 standard deviations in the wet-dipole case. Those in the Guinea Coast are of the order of 1 standard deviation. In the wet dipole, the 'rainbelt' has a strong core from 12 °N to 16 °N (Figure 4). Rainfall reaches 280 mm/mo. In the dry-dipole composite, the rainbelt is only somewhat weaker, but it lies considerably further south. Maximum rainfall is about 230 mm/mo. The core of the rainbelt lies between 7 °N and 11 °N in August, prolonging the rainy season south of the Sahel.

Figure 5 confirms that the rainfall anomalies in these composites are linked to anomalous northward and southward displacements of the AEJ, as our hypothesis suggests. The jet core over West Africa lies roughly 5° further south in the dry years than in the wet years, commensurate with the 5° southward displacement of the rainbelt. Thus, the overriding question concerning the variability is what produces this shift, especially during August.

6. Results

6.1. Surface pressure and air temperature

Figures 6 and 7 show SLP and SST for August, averaged over the three years in each composite. The permanent geographical features – the Sahara to the north

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

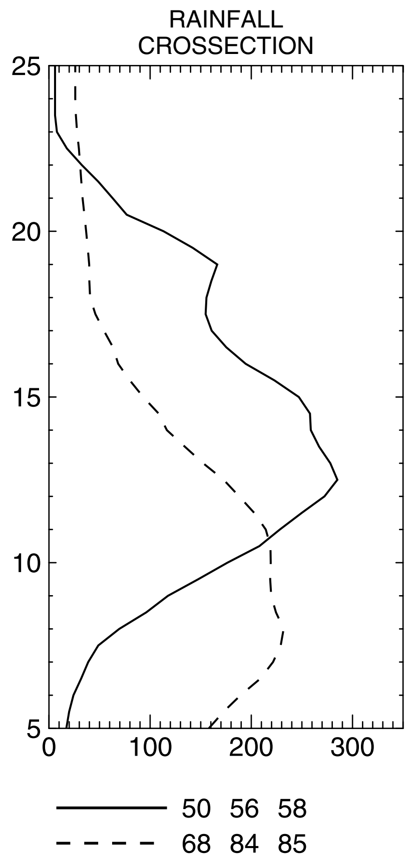

Figure 4. August rainfall (mm/mo) as a function of latitude (°N) over West Africa for the wet- and dry-dipole composites (averaged over 5 °W–5 °E).

and the Atlantic Ocean and tropical rainforest to the south – prescribe the meridional temperature and pressure gradients. Clearly, a strong cross-equatorial pressure gradient, a major prerequisite, is present in the West African sector (Figure 6). It largely reflects the surface temperature gradient between the Sahara and the Atlantic (Figure 7).

A further consistency with our hypothesis is that the surface pressure gradient is stronger in the wet-dipole years than in the dry-dipole years. The difference map in Figure 6 shows a strengthening of the gradient in wet-dipole years. Examination of the contours in the wet- and dry-dipole composite maps in Figure 6 shows that at 0 °W the pressure change from 20 °N to 20 °S is +15.5 mb in the wet composite, compared to +12 mb in the dry composite. The change from 15 °N to 15 °S is 10 mb in the wet-dipole composite versus 7.5 mb in the dry-dipole composite. Similar contrasts are apparent at 20 °W, 0 °W, and 10 °E (Table I). The greater pressure gradient in the wet years (Figure 6) is also apparent for each individual year of the composite. The contrast between the most extreme wet and dry years (1958 and 1984) is particularly strong, with a pressure gradient of 18 mb between 20 °N and 20 °S in 1958 versus 11.5 mb in 1984.

The contrast in pressure gradients is largely due to low pressure over the western Sahara, suggesting a possible link with the SST distribution in the North Atlantic. However, the contrast is even greater between the latitudes 30 °S and 20 °N, because of intense high pressure in the South Atlantic in the wet case. Over these latitudes, the change is +20 mb in the wet case and +14 mb in the dry case. This suggests the importance of inter-hemispheric temperature contrast, as suggested by Folland *et al.* (1986, 1991) and others.

August SST distributions are shown in Figure 7. Ideally, a temperature gradient along 0 °W would be evaluated. However, accurate surface temperatures are not available over the desert. Thus, the gradient is calculated over the marine portion, along a transect running from (20 °N, 20 °W) to (20 °S, 0 °W). It is clearly apparent that the SST gradients are higher in the wet-dipole composite than in the dry-dipole composite, consistent with the stronger pressure gradient in the wet case. The temperature change from 15 °N to 15 °S is −6.5 °C in the dry case, compared to −8 °C in the wet case. The change from 10 °N to 10 °S is −4.5 °C in the dry case versus −6.5 °C in the wet case, corresponding to a gradient that is nearly 50% larger. Since surface temperatures over the Sahara are relatively constant from year to year, this is representative of the contrast in the surface temperature gradient from the Atlantic to the Sahara between the wet-dipole and dry-dipole composites. The contrast is largely due to temperatures in the Gulf of Guinea, which numerous studies (e.g. Lamb, 1978a; Fontaine and Janicot, 1996; Fontaine *et al.*, 1998) have shown to be a factor in rainfall variability in the Sahel.

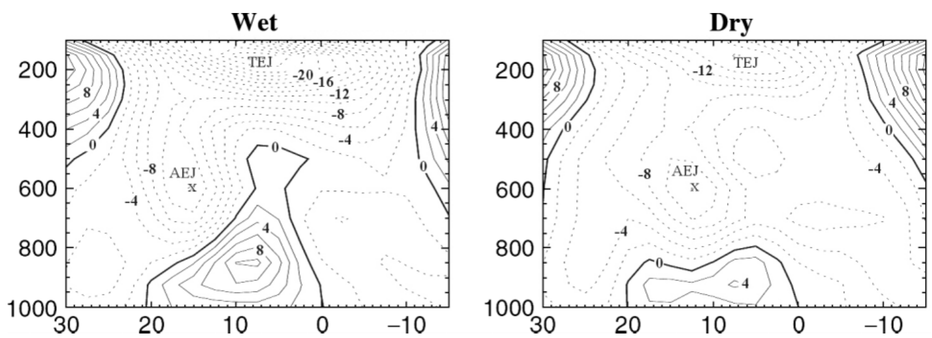

Figure 5. Vertical cross-section of August mean zonal wind (ms^{-1}) at 0 °W for the wet-dipole and dry-dipole composites. Isotachs correspond to increments of 2 ms^{-1}, with dashed lines indicating easterlies and solid lines indicating westerlies. Data are from the NCEP Reanalysis. The core of the AEJ is marked with an 'X'.

Copyright © 2007 Royal Meteorological Society

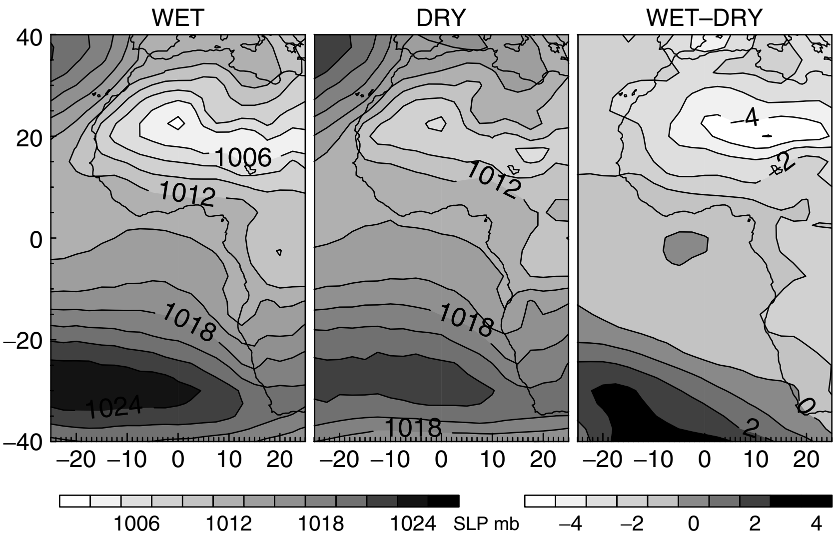

Figure 6. Mean August SLP for the wet-dipole and dry-dipole composites, and the difference between the two composites.

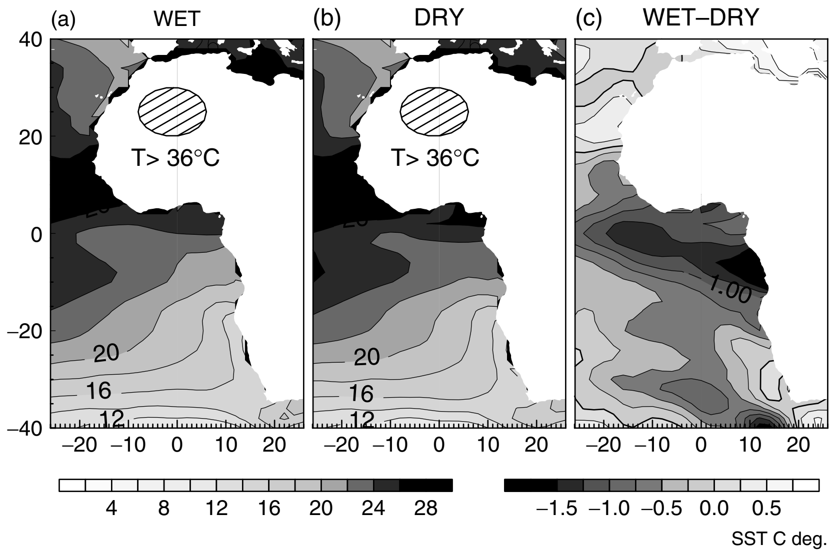

Figure 7. Mean August SST for the wet-dipole and dry-dipole composites, and the difference between the two composites. The stippled area indicates the core of the Sahara, where temperatures exceed 36 °C.

Table I. Meridional pressure gradient and latitude of zero absolute vorticity at four longitudes over West Africa for the wet-dipole and dry-dipole composites. The gradient is expressed as the pressure difference between the indicated latitudes.

	Wet composite				Dry composite			
Longitude	20 °W	10 °W	0 °W	10 °E	20 °W	10 °W	0 °W	10 °E
ΔP (20 °N–20 °S) (mb)	9.5	14	15.5	9.5	7.5	11.5	12	7.5
ΔP (15 °N–15 °S) (mb)	6.5	8.5	10	7.5	6	7	7.5	5
Latitude of zero absolute vorticity	5 °N	3 °N	4 °N	5 °N	4.5 °N	4 °N	3.5 °N	3 °N

Furthermore, temperatures in the Gulf of Guinea and the southern equatorial Atlantic are as much as 2 °C higher in the dry-dipole composite (Figure 7). This is consistent with the suggestion of Lamb (1978a) that the higher SSTs in the equatorial Atlantic promote convection in the equatorial latitudes to the south of the Sahel. In contrast, the low temperatures of the wet-dipole years push the SST maximum further north. This may be a contributing factor to the stronger convection in the Sahelian latitudes in the wet-dipole years.

6.2. Absolute vorticity and the divergent wind

Figure 8 shows the magnitude of the meridional component of the divergent wind in August at 1000 mb

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

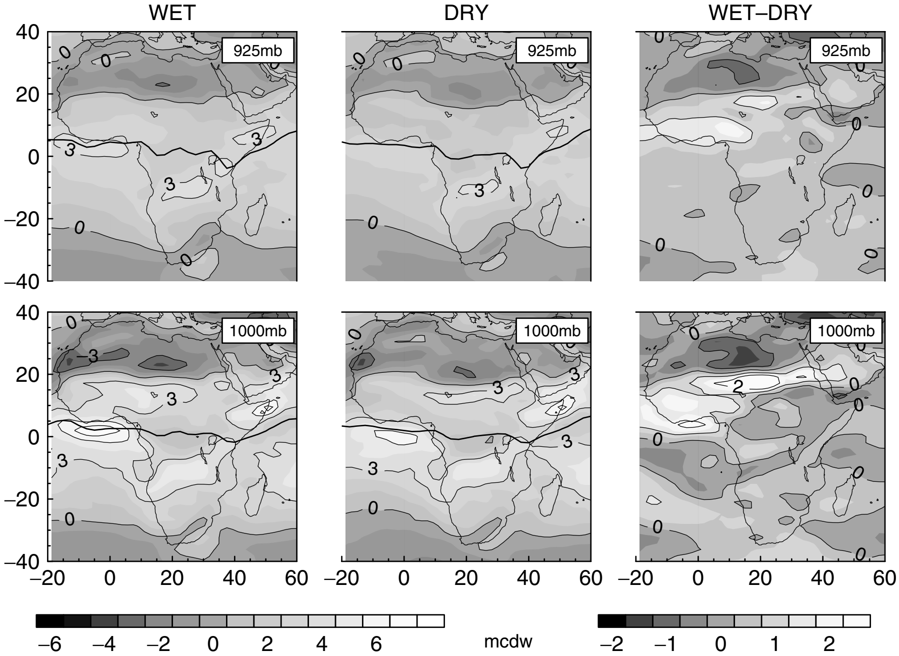

Figure 8. Meridional component of the divergent wind (10^{-6} ms^{-1}) in August, at 1000 mb and 925 mb, for wet-dipole and dry-dipole composites, and difference fields. The zero absolute-vorticity contour at 925 mb and 1000 mb is indicated by a thick solid line.

and 925 mb for the wet and dry composites. The data for 925 mb are calculated as the average of 1000 mb and 850 mb, using the velocity potential from the NCEP Reanalysis dataset. The zero absolute-vorticity contour is also shown. In all four cases, this contour is displaced into the summer hemisphere. At 925 mb it bisects the meridional divergent wind maximum in the wet composite, but lies well to the north of the maximum in the dry composite. The situation is similar at 850 mb.

A more obvious contrast is that at 1000 mb and 925 mb over West Africa the meridional component is much stronger in the wet-dipole composite than in the dry-dipole composite, and at 1000 mb it is also further north. The differences are apparent in the Gulf of Guinea and equatorial Atlantic and over the Sahel and Guinea Coast regions of the continent. At 1000 mb, the wet–dry contrast is particularly strong in the Gulf of Guinea and in the central Sahel north of Lake Chad.

Figure 9 plots the latitude of the zero absolute-vorticity contour as a function of pressure gradient in August. The data cover the years 1948–2004, and represent an average over the sector 5 °W–5 °E. Although the data are highly scattered, there is a clear linear relationship between the pressure gradient and the latitude of the zero absolute-vorticity contour, as established by Tomas and Webster (1997) (see also Figure 5). The correlation between the variables is 0.65. A relationship with rainfall is also apparent. The pressure gradient and the latitude of the zero contour are larger for the wet decades of the 1940s through 1960s than for the dry decades that follow. This supports our contention that the inertial-instability mechanism plays a role in rainfall variability over West Africa.

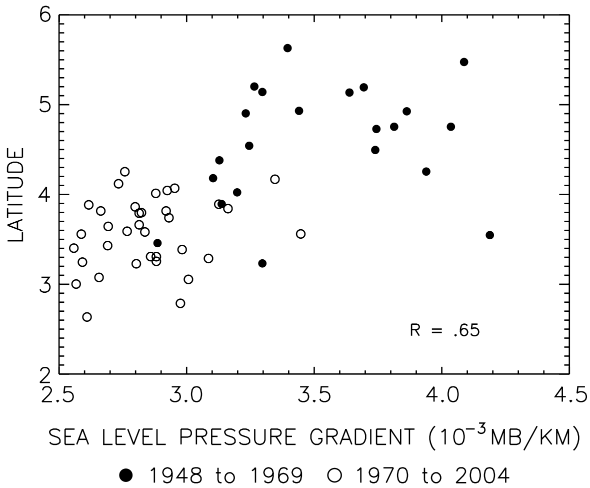

Figure 9. Latitude (°N) of the zero absolute-vorticity contour at 925 mb, plotted against surface pressure gradient, in August for the years 1948–2004. Data are averaged over 5 °W–5 °E. The solid circles represent the years 1948–1969 and the open circles represent the years 1970–2004.

6.3. Horizontal wind field and relative vorticity

Figures 10 and 11 depict the total wind vector and the mean zonal wind in the wet-dipole and dry-dipole composites. The wind vector shows a weaker (stronger) AEJ, a northward (southward) displacement of the AEJ, and a stronger (weaker) TEJ, in the wet (dry) cases. Consistent with earlier analyses, a strongly-developed low-level westerly jet appears in the wet-dipole years, but is absent in the dry-dipole years. The core of the jet is at 850 mb, but there is some indication of westerlies as high as 600 mb. The elevated westerly maximum and jet core suggest that this flow is independent of the low-level

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

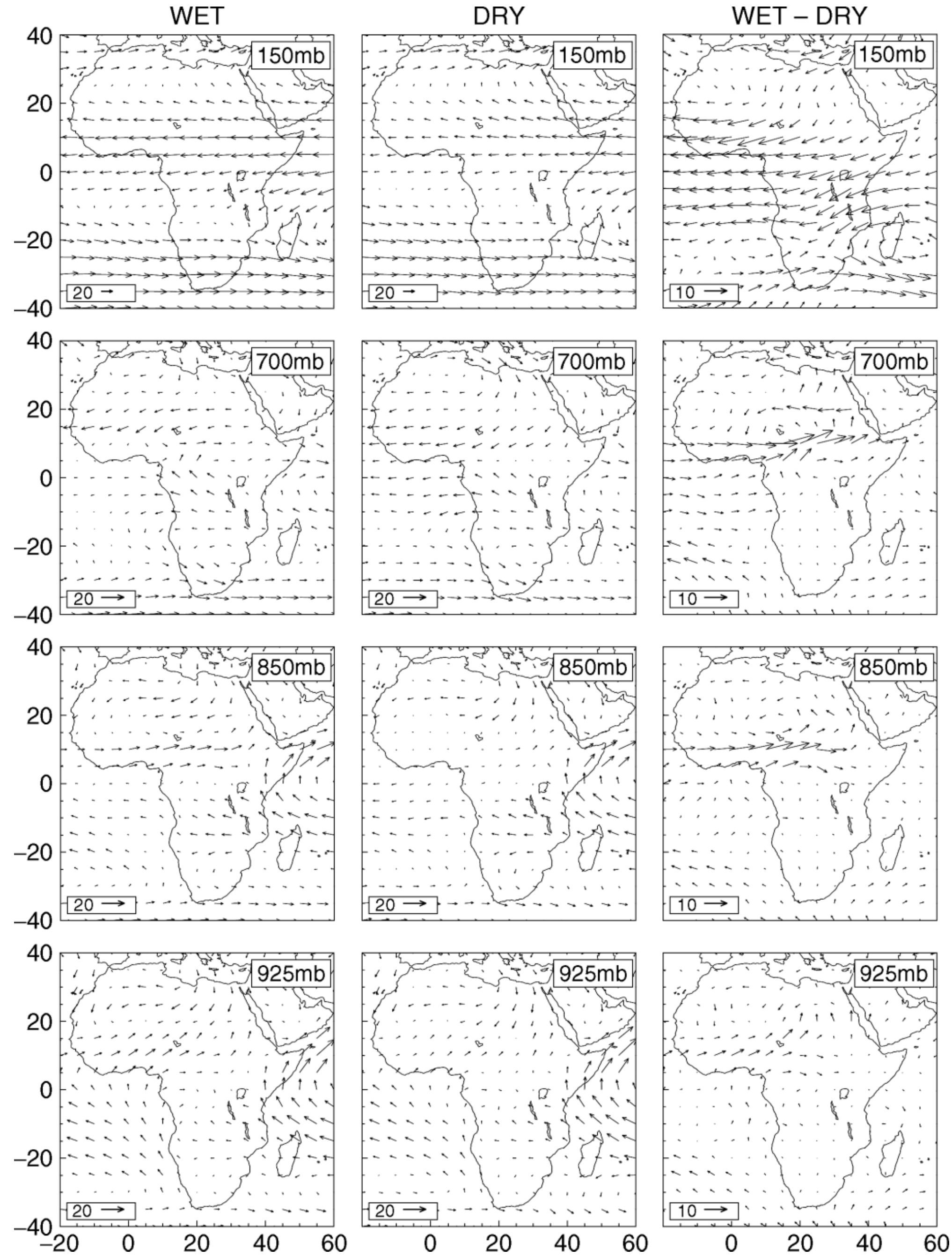

Figure 10. August mean wind vector (ms^{-1}) at four levels, for wet-dipole and dry-dipole composites, and the difference fields.

'southwest monsoon' flow that prevails during the boreal summer. Additional analyses of upper-level winds and humidity confirm this (Grist and Nicholson, 2001).

The wet–dry difference field (Figures 10 and 11) underscores the importance of the westerly jet, and the contrast in its strength and vertical development between the two cases. The presence of the westerly jet not only displaces the easterly maximum (i.e. the AEJ) northward, but also displaces the AEJ core westward. Thus, in the wet-dipole years the core of the AEJ lies over the western Sahel and the Atlantic Ocean, whereas in the dry-dipole years it lies over the central Sahel. Even stronger contrasts are apparent in the TEJ. In the wet-dipole years it is displaced equatorward and its core is strong over West Africa. In the dry-dipole years the TEJ is much weaker, especially over Africa, and its influence is primarily over the far eastern Sahel.

The relative vorticity at 700 mb (Figure 12) clearly shows the intensity and location of the AEJ and its contrasting character in the wet- and dry-dipole composites. A strong positive vorticity maximum indicates the position of the AEJ and corresponds to the cyclonic shear on its equatorward flank. In the wet-dipole composite, positive vorticity prevails from roughly 5 °N to 14 °N: thus in latitudes including much of the Sahel. The positive vorticity maximum overlies the region of lower-tropospheric

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

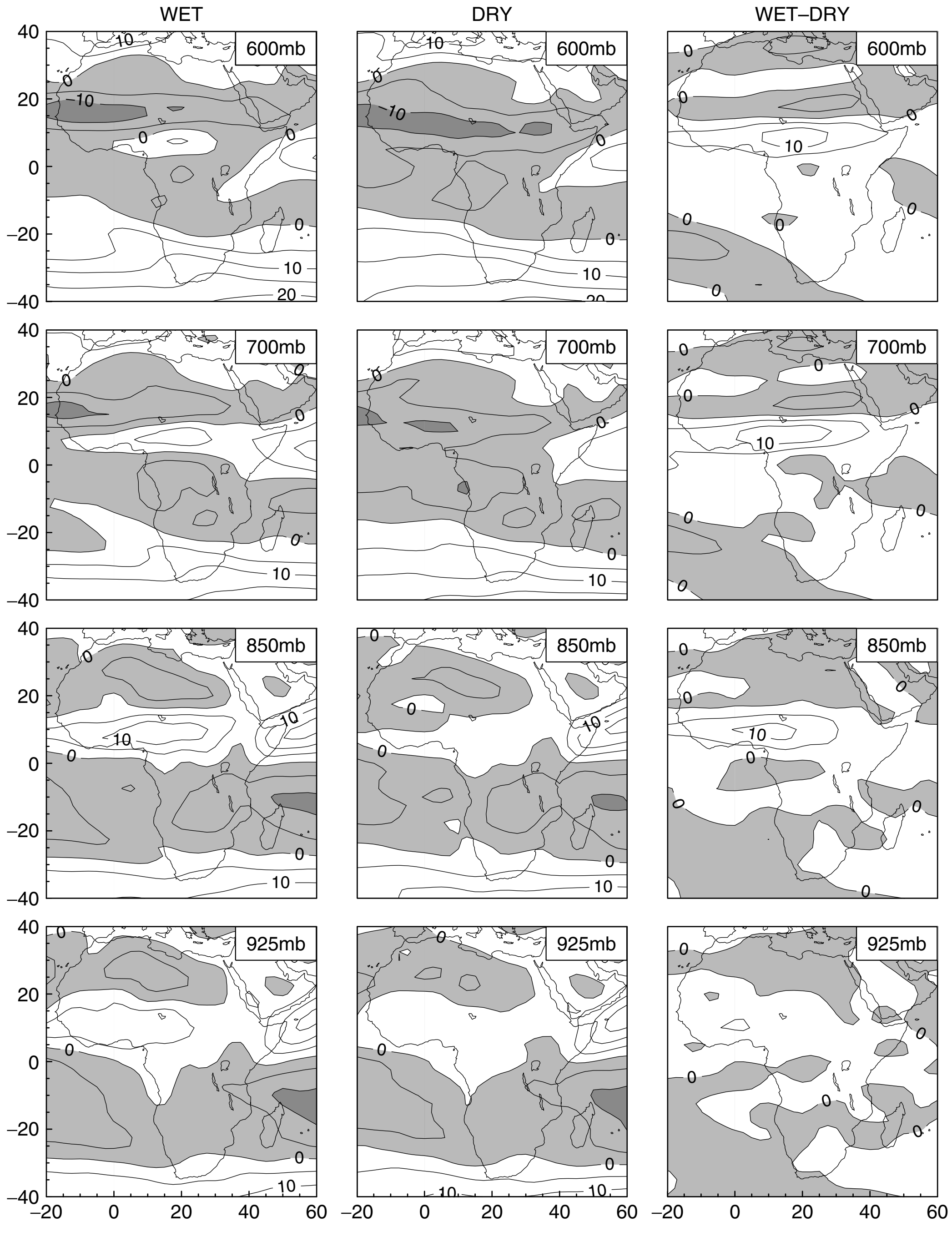

Figure 11. August mean zonal wind (ms^{-1}) at four levels, for wet-dipole and dry-dipole composites, and the difference fields.

westerlies, and reaches 1.8×10^{-6} ms^{-1}. In the dry-dipole composite, the region of positive vorticity lies south of the Sahel, extending from roughly 3 °S to 10 °N, and the maximum is 0.9×10^{-6} ms^{-1}.

Figure 13 shows the mean meridional flow at 925 mb and 850 mb. The figure mainly illustrates the influence of the low-level westerly jet. This is generally considered to be the humid 'monsoon' flow, and hence the mechanism of moisture transport into the lower troposphere over West Africa. In the dry-dipole composite, the maximum lies over the southeastern Atlantic, more or less over the cold Benguela current. Northward flow is weak over the Gulf of Guinea and over the continent. In the wet-dipole composite, the meridional flow is much stronger, has a maximum near the Guinea Coast, and extends well across the Sahel to 20 °N. In the wet-dipole composite the northward flow over the continent is strongly evident even at 850 mb; this is not the case in the dry-dipole composite.

6.4. Divergence and vertical motion

Figure 14 shows the divergence field at three levels for the wet-dipole and dry-dipole composites. The 150 mb level corresponds to the TEJ, and the core of the AEJ lies near the 700 mb level. The relative positions of the AEJ

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

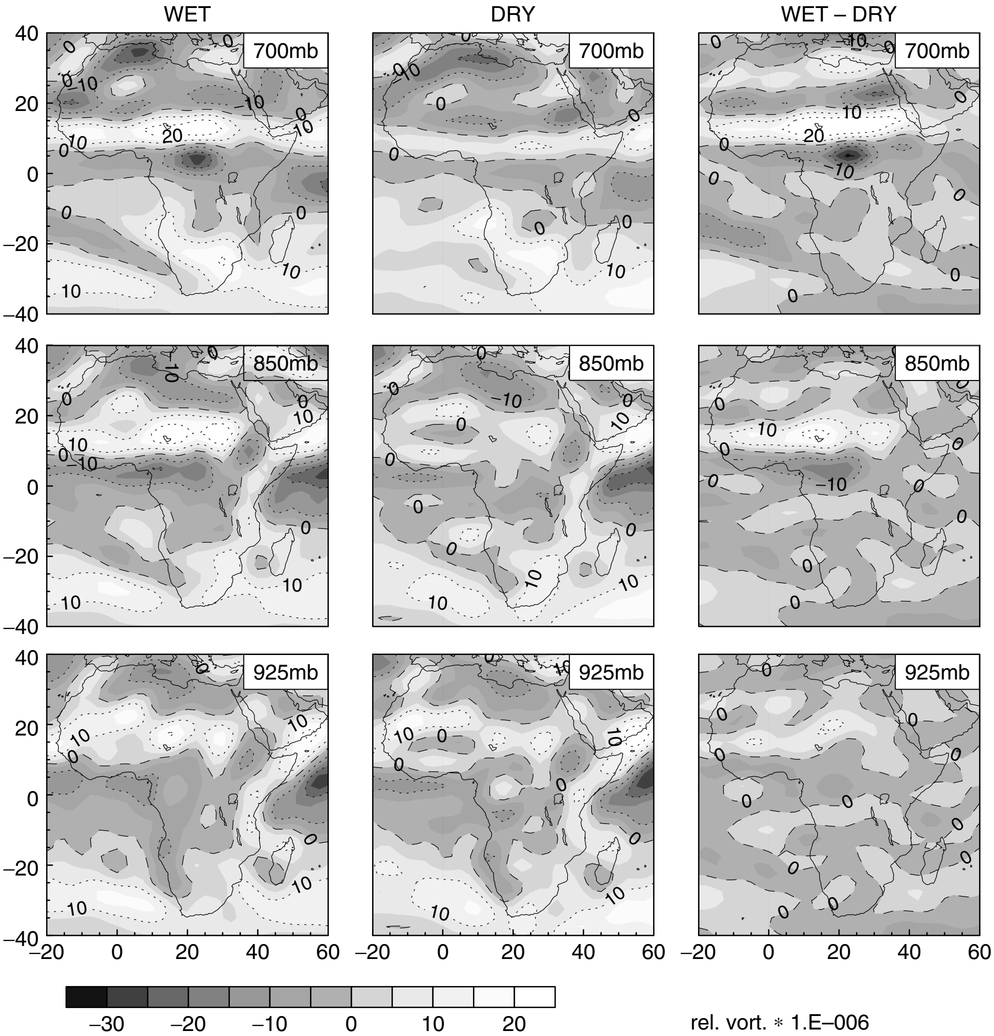

Figure 12. August relative vorticity (10^{-6} ms^{-1}) at 700 mb, 850 mb and 925 mb, for the wet-dipole and dry-dipole composites, and the difference fields.

and TEJ largely control the divergence field in the two composites and modulate the evolution of the divergence field with height.

Near the surface (925 mb), in both composites there are two areas of convergence over West Africa. A quasi-continuous band of convergence associated with the ITCZ lies over the central and northern Sahel. This band extends from roughly 15 °N to 25 °N. In the dry-dipole composite, it is notably weaker, further south, and more constricted than in the wet-dipole composite. For the most part, the atmosphere is divergent over the southern Sahel in the dry-dipole composite. A second area of convergence straddles the coastline, and is probably related to the uplift and increased friction created when the southwest monsoon reaches the coast. In the wet-dipole composite it is weaker and displaced slightly northward.

Consistent with the Tomas–Webster hypothesis, the wet-dipole composite shows strong convergence in the boundary layer (e.g. 925 mb) in a broad zone commencing immediately poleward of the zero absolute-vorticity contour and extending across the central and northern Sahel. Divergence prevails equatorward of this contour. The prerequisites for the mechanism do not appear to be met in the dry-dipole case. There is only weak convergence poleward of the zero absolute-vorticity contour, and it covers a relatively narrow latitudinal strip lying west of 0 °W. Divergence extends over much of the Sahel.

In the middle and upper troposphere, the contrast between the wet- and dry-dipole composites appears to diminish. The spatial patterns are similar in the two cases at both the AEJ level (700 mb) and the TEJ level (150 mb). For example, in both cases a divergence–convergence couplet is apparent at 700 mb, with off-equatorial convergence north of the zero absolute-vorticity contour (see Figure 8) and divergence in the equatorial region to the south. A closer examination shows that the two cases contrast strongly in terms of the magnitude and latitudinal extent of the convergence.

At 700 mb, the convergence extends from the Equator to roughly 10°–13 °N in the dry-dipole composite, and lies well to the south of Lake Chad. Its magnitude exceeds 1.5×10^{-6} ms^{-1} only in a relatively small area

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

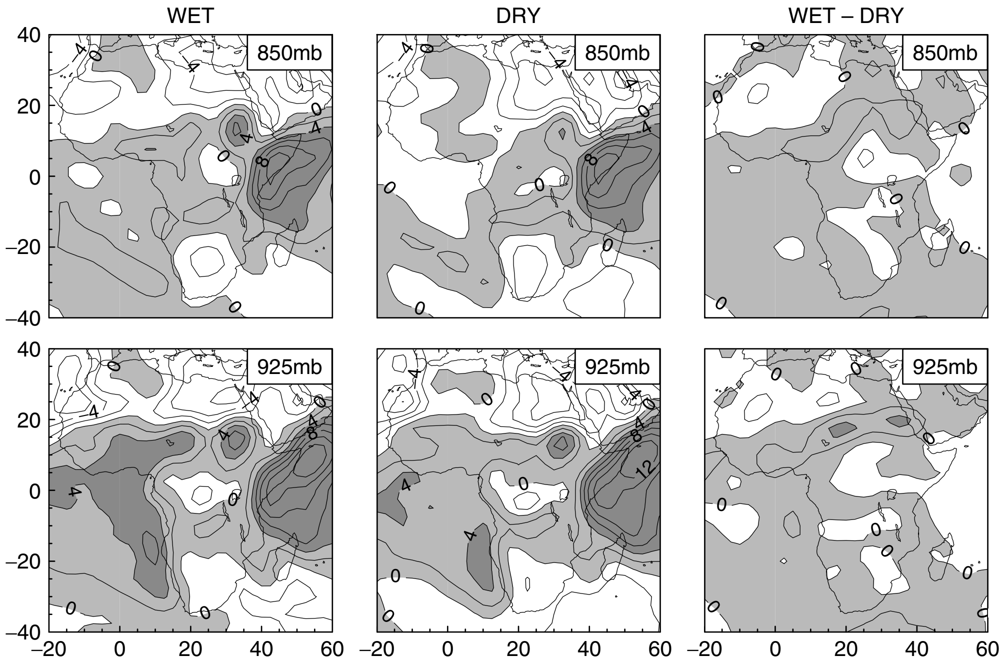

Figure 13. August mean meridional wind (ms^{-1}) at 925 mb and 850 mb, for the wet-dipole and dry-dipole composites, and the difference fields.

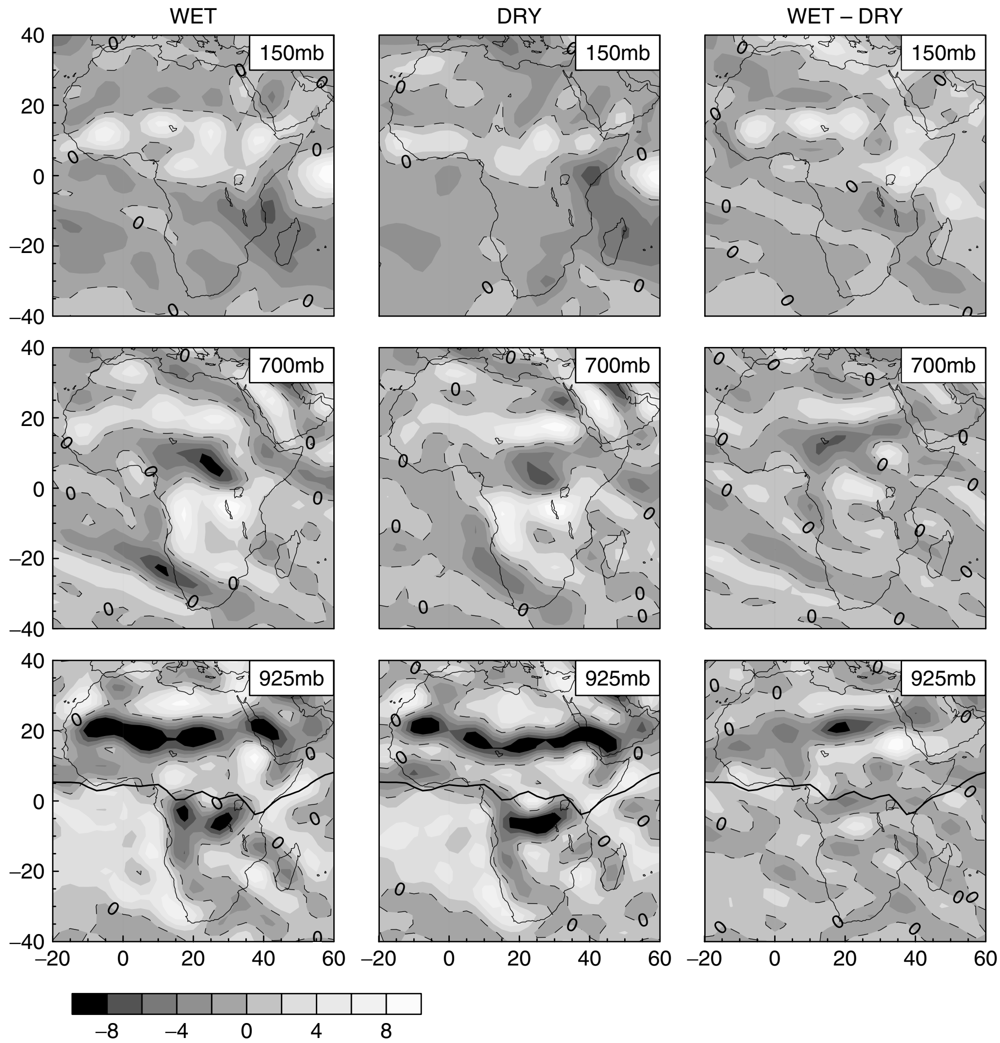

Figure 14. August mean divergence (10^{-6} ms^{-1}) at 150 mb, 700 mb and 925 mb, for the wet-dipole and dry-dipole composites, and the difference fields.

in the central equatorial latitudes, and divergence overlies the Sahel. In the wet-dipole composite, the convergence at 700 mb extends from roughly the Equator to 17 °N, well to the north of Lake Chad. Its magnitude exceeds 1.5×10^{-6} ms^{-1} throughout most of this region. The result is a continuous column of convergence extending from the surface to at least 700 mb. This will entrain equatorial moisture into the lower and middle troposphere and into the AEJ. This characteristic of the wet-dipole composite may be particularly important because, in order for the AEJ to organize convection into larger and more intense mesoscale systems, the moist layer must reach the jet level (Miller and Lindzen, 1992).

Starker contrast is apparent at the TEJ level. In the wet-dipole composite, strong upper-level divergence lies in the anticyclonic shear on the poleward flank of the TEJ. This overlies the convergence associated with the mid-level AEJ. Such a circulation exists over the Sahel, and implies strong ascent over the Sahel. In the dry-dipole composite, the upper-level divergence and mid-level convergence are markedly weaker, implying weaker ascent. This situation prevails south of the Sahel. In the latitudes of the Sahel, convergence prevails at the TEJ level, and overlies a region of mid-level divergence. This implies subsidence over the Sahel.

These conclusions are confirmed by the analysis of vertical motion. Figure 15 shows the vertical motion fields for north–south transects centred over the central Sahel at 0 °W. Transects have also been derived at 10 °W and 10 °E; these yield similar results and are not shown. In both the dry-dipole case and the wet-dipole case, the vertical motion is strongest in the layers between the AEJ and the TEJ, with a maximum at 400 to 500 mb. In the wet-dipole composite, a strong core of vertical motion extends over roughly 15° of latitude at middle levels and over 20° of latitude in the boundary layer. There is strong coupling between the surface and middle levels. Maximum vertical motion is at roughly 12°–15 °N in the upper atmosphere, but closer to 20 °N near the surface. Furthermore, consistent with our hypothesis, a subsidence maximum lies near the Equator. In the dry-dipole composite, the vertical motion is of similar magnitude, but extends over only some 8°–10° of latitude. Near the surface, there is upward motion north of the Equator and subsidence to the south.

Convection shows a consistent association with the vertical motion fields. In the wet-dipole composite, a strong rainfall maximum lies over the Sahel at roughly 13 °N. In the dry-dipole composite, the resultant convection has a maximum at about 8 °N, well to the south of the Sahel. The difference field readily explains the anomaly patterns in Figure 3. There is increased vertical motion over the Sahel and increased subsidence over the Guinea Coast in the wet-dipole case, but the reverse is true in the dry-dipole case. Consistent with the similar magnitudes of maximum vertical motion, rainfall in the core of the rainbelt is fairly similar in the two composites: of the order of 230 mm/mo in the dry-dipole case and 280 mm/mo in the wet-dipole case.

Figure 15 shows another interesting feature: a second intense core of vertical motion centred at 20 °N over the western Sahara. This core is confined to the lower atmosphere, never extending above about 750 mb. The ascent is strong between 950 mb and 800 mb. This feature is obviously associated with the heat low over the western Sahara. In the wet-dipole composite, this low-level core is merged with the extensive vertical-motion maximum over the Sahel, and a secondary convective maximum corresponds to the low-level core. This secondary maximum is absent in the dry-dipole composite, although the vertical motion over the western Sahara is relatively strong. An explanation for this contrast may lie in the meridional wind fields shown in Figure 13. In the wet-dipole case, a continuous stream of southerly flow extends from the Benguela coast to the southern Sahara. This provides moisture transport into the heat low over the western Sahara, where the vertical motion is strong enough to induce some rainfall. In the

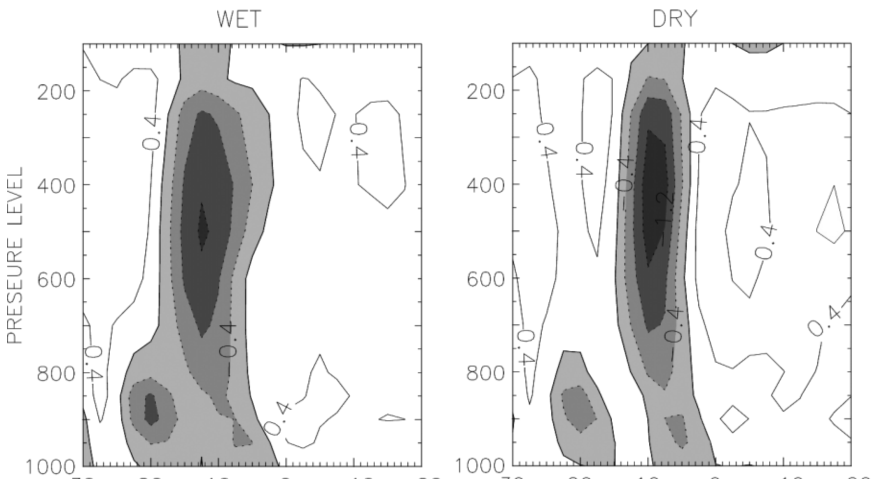

Figure 15. Mean vertical motion (omega) (10^{-3} hPa/s) in August at 0 °W. Left: wet-dipole composite. Right: dry-dipole composite.

Copyright © 2007 Royal Meteorological Society
DOI: 10.1002/qj

dry-dipole composite, both the moisture transport and the secondary maximum of rainfall and convection near 20 °N are lacking. Also notable in the dry-dipole case is that the low-level vertical motion over the Sahara is decoupled from the vertical motion at middle and upper levels. The result is strong subsidence over the Sahelian latitudes.

7. Discussion

7.1. The low-level westerly monsoon wind maximum

The well-developed westerly wind maximum at 850 mb strongly supports our contention that inertial instability plays a role in interannual variability over West Africa. For a detailed discussion of this the reader is referred to (Tomas and Webster, 1997, pp. 1477–1478). Their derivation begins with the fact that near the latitude of zero absolute vorticity ($\eta = 0$), the advection of absolute vorticity must be zero. At that latitude the advection by the rotational component of the wind $V\psi$ is also zero, so that any advection of η by the divergent component $V\chi$ must be balanced by dissipation of relative vorticity ζ. Thus:

$$-V\chi \cdot \nabla\eta = \alpha\zeta|_{\eta=0},$$

where α is a coefficient representing dissipation. As $V\chi$ has a local maximum at $\eta = 0$, ζ must also have a local maximum at that latitude to satisfy the balance condition.

In the zonally-axisymmetric case of West Africa, the maximum in relative vorticity ζ must result from a maximum in the zonal shear. The wind in this region is typically easterly above a shallow low-level surface monsoon flow. However, when the cross-equatorial pressure gradient is strong, the shear required for the balance is sufficient to turn the wind to westerly. Moreover, the westerly winds must rapidly increase in strength to the north of the $\eta = 0$ contour.

This rapid increase just to the north is certainly evident in the wet composite (Figure 11). The relationship between the latitude of $\eta = 0$ and the surface pressure gradient (Figure 9) further supports the hypothesis that the inertial-instability mechanism is operative. The high linear correlations between the speed of the westerlies at 850 mb and the latitude of $\eta = 0$ (Figure 16, upper panel, $r = 0.70$), and between the speed of the westerlies at 850 mb and the surface pressure gradient (Figure 16, lower panel, $r = 0.84$) provide final confirmation (In mid-latitudes, geostrophy predicts a linear relationship between pressure gradient and wind speed. The relationship shown in Figure 16 (lower panel) is not a result of geostrophy. First, westerlies are at 850 mb but the pressure gradient is at the surface. Westerly flow is not strongly evident at the surface. Secondly, the westerlies are at 5°–10 °N, too far south for geostrophic balance; at these latitudes the prevailing surface pressure gradient would produce more southerly flow). Moreover, both the 850 mb westerlies and the surface pressure gradient are strong during the wet period prior to 1970, but weak during the dry decades that follow. Thus we conclude that inertial instability plays a role in interannual and interdecadal variability (see also Nicholson, 2007c).

Although we have focused on the role of the low-level westerlies in displacing the AEJ northward, this flow can influence the West African rainfall regime in other ways. Flohn (1964), for example, points out that westerly flow in the Tropics is more unstable than easterly flow, as it promotes vertical motion (see also (Lettau, 1956)). He notes that rainfall occurs much more frequently in the case of equatorial westerlies than when easterlies prevail. Rowell and Milford (1993) have shown that above-average low-level westerly flow increases the probability of squall-line development.

7.2. Inertial instability in the context of tropical Atlantic variability

Three geographic regions contribute to the contrasting SST and SLP gradients in wet and dry Sahel years: the eastern equatorial region and Benguela coast (where the

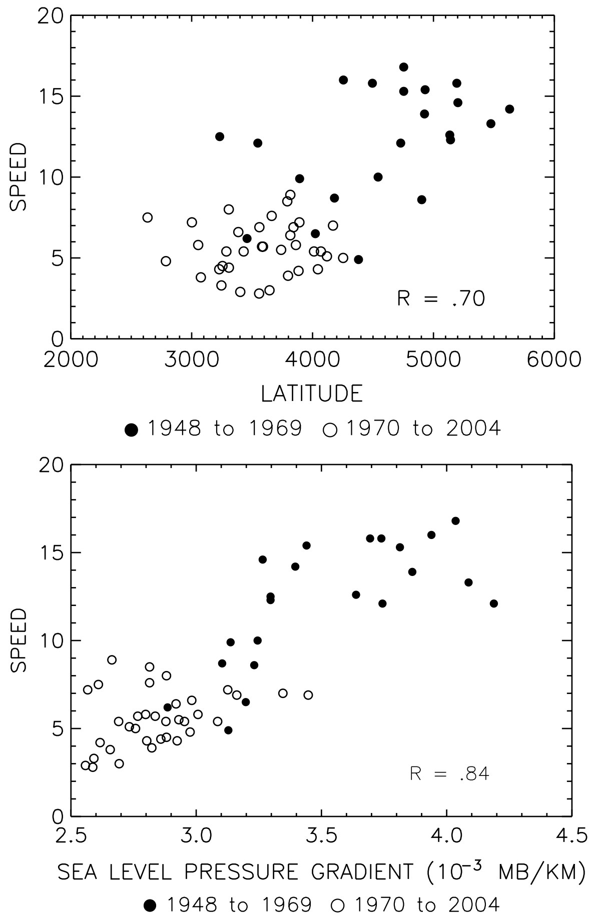

Figure 16. Upper panel: speed of the westerlies (ms^{-1}) at 850 mb plotted against latitude (°N) of the zero absolute-vorticity contour at 925 mb, for August of the years 1948–2004. Data are averaged over 5 °W–5 °E. The solid circles represent the years 1948–1969 and the open circles represent the years 1970–2004. Lower panel: speed of the westerlies (ms^{-1}) at 850 mb plotted against surface pressure gradient (10^{-3} mb/km), for August of the years 1948–2004. Data are averaged over 5 °W–5 °E. The solid circles represent the years 1948–1969 and the open circles represent the years 1970–2004.

Copyright © 2007 Royal Meteorological Society

contrast is in SST); the tropical North Atlantic; and the South Atlantic (where the contrast is in SLP). The SST anomaly pattern is reminiscent of the Atlantic Niño mode, one of the two most common patterns of variability in the tropical Atlantic (Covey and Hastenrath, 1978; Philander *et al*., 1996; Chang *et al*., 2000; Latif and Grötzner, 2000; Wang and Carton, 2003). This mode is marked by the disappearance of the equatorial cold tongue in the eastern Atlantic, and it is most developed in the boreal summer, coincident with the Sahel rainy season. This is an interannual mode, with peak variance on time-scales of 2–5 years (Zebiak, 1993; Carton *et al*., 1996; Mo and Häkkinen, 2001).

The disappearance of the cold tongue is linked to the relaxation of the trades, a situation comparable to the Pacific El Niño. The role of the trades implies that the dynamics of the ITCZ–equatorial-cold-tongue complex might act in tandem with the inertial-instability mechanism to displace the ITCZ and related convection northward into the Sahelian latitudes of West Africa. The link is via the pressure changes in the South Atlantic that enhance the pressure gradient in the wet years (see Figure 6). The higher pressure implies a strengthening of the South Atlantic High, which is ultimately the origin of the equatorial trades. Stronger trade winds imply a stronger southwest monsoon – because the monsoon flow arises as the trades cross the Equator – and stronger moisture transport from the Atlantic to the interior of West Africa.

The strong pressure anomalies in the tropical North Atlantic and the South Atlantic in the wet and dry composites suggest links to more global phenomena. The former sector is involved in the North Atlantic Oscillation (NAO) (Czaja *et al*., 2002). However, the NAO is linked to the tropical Atlantic mainly during the boreal winter, and in fact it may be forced by the tropical Atlantic (Robertson *et al*., 2003). Nevertheless, the strength of the anomalies suggests that possible relationships with the NAO should be further investigated.

The South Atlantic sector that stands out in Figure 6 is recognized as a region of strong interannual variability (Palastanga *et al*., 2002; Huang and Shukla, 2005). Interestingly, fluctuations in this sector appear to be controlled by the Antarctic Oscillation (Gong and Wang, 1999). The South Atlantic has been identified as a possible source of remote forcing of tropical Atlantic variability (Rajagopalan *et al*., 1998; Venegas *et al*., 1997; Tourre *et al*., 1999; Mo, 2000; Mo and Häkkinen, 2001; Hickey and Weaver, 2004) and African rainfall variability (Camberlin *et al*., 2001). It appears prominently in our reconstruction of links to the Pacific ENSO (Nicholson, 1997; Nicholson and Kim, 1997) and in our contrast between wet and dry years in western Central Africa (Balas *et al*., 2007) and Southern Africa (Nicholson, 1989). Warming in the South Atlantic precedes warming events in the tropical Atlantic (e.g. Nicholson, 1997). This is interesting in view of the conclusion of Mo (2000) (see also (Mo and Häkkinen, 2001)) that a wave train from the eastern Pacific, generally occurring during ENSO events, can influence variability over the tropical Atlantic via the South Atlantic. Because the South Atlantic appears to be a major factor in the interannual variations in the cross-equatorial pressure gradient, possible roles of the Pacific and extratropical South Atlantic in Sahel rainfall variability merit further study.

7.3. Interannual persistence of rainfall anomalies

One of the most distinctive features of the Sahel rainfall regime is the interannual, and even interdecadal, persistence of anomalously wet or anomalous dry conditions (Nicholson and Palao, 1993) that is strongly evident in Figure 1. A very open question is what physical processes lead to this persistence. In particular, what maintains the anomalous position of the AEJ over several years? Nicholson (e.g. 2000) and others have suggested land–atmosphere feedback, particularly via soil moisture. Some studies have even provided observational evidence of such effects (e.g. Taylor and Lebel, 1998). It is difficult to reconcile this concept with the ideas proposed here, although surface temperature or moisture gradients might in some way promote the persistence of the requisite surface pressure gradients. A recent paper by Rowell (2003) suggests that four positive atmospheric feedback loops come into play in this region. These involve rainfall, the strength of the AEJ, latent heat release near the AEJ, and moisture convergence near the AEJ. As with land–atmosphere feedback, these cannot easily be extrapolated to interannual or interdecadal time-scales.

The time-scales of persistence are consistent with those seen in tropical Atlantic variability. Both the interhemispheric mode and the NAO exhibit low-frequency variability. The latter dominates ENSO on decadal and longer time-scales (Czaja *et al*., 2002). Recent studies have suggested that numerous other low-frequency modes affect the Atlantic. One is an Atlantic multidecadal mode (e.g. Kushnir, 1994; Mehta, 1998; Enfield and Mestas-Nuñez, 1999), reflecting lower SSTs in the North Atlantic in the 1980s and 1990s and a warm phase in the 1950s and 1960s. Another is the Antarctic Oscillation (Huang, 2003; Huang and Shukla, 2005), which appears to affect the South Atlantic. Chelliah and Bell (2004) have also defined a tropical multidecadal mode based on 200 mb velocity potential and surface temperature in the region 30 °N–30 °S. These authors conclude that it may be linked to the downward trend in West African rainfall.

8. Summary and conclusions

Our analysis has shown that the various 'prerequisites' for inertial instability in the eastern Tropical Atlantic are satisfied in August for the wet-dipole composite. These include a strong cross-equatorial surface pressure gradient, a displacement of the zero absolute-vorticity contour into the summer hemisphere, a meridional divergent wind maximum bisected by the zero absolute-vorticity contour, divergence equatorward of the contour, and convergence poleward of the contour. These conditions are not satisfied for the dry-dipole composite. This suggests that

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

inertial instability does play a role in producing the wetter conditions in the Sahel.

Further evidence is that signatures of the instability are apparent in the region. These include an off-equatorial convective maximum in the Northern Hemisphere, strong ascent and enhanced convection to the north of the zero absolute-vorticity contour, an elevated westerly wind maximum just north of the Equator, and subsidence above the Equator. The westerly wind maximum accounts for the northward displacement of the AEJ and associated disturbances. The ascent and enhanced convection lies over the Sahel, thus producing the anomalously wet conditions. The subsidence accounts for the commonly-occurring rainfall dipole: abnormally dry conditions over the Guinea Coast when the Sahel is 'wet'.

The development of the westerly wind maximum suggests possible teleconnections to other regions of Africa, most notably equatorial Africa. A well-known feature of the climate of equatorial East Africa is the association between equatorial westerlies and anomalously high rainfall (Nicholson, 1997). Nicholson and Entekhabi (1987) have found statistical evidence that the Atlantic plays a more important role in governing rainfall in East Africa than does the Indian Ocean. However, the well-documented association with El Niño appears to involve both the Atlantic and the Indian Oceans (Nicholson, 1997; Nicholson and Kim, 1997). The development of strong and deep equatorial westerlies, and the link to Southern-Hemisphere pressure, may provide the mechanism for the statistical associations.

A link between rainfall variability in the Sahel and SSTs has been well established by numerous studies, beginning with the work of Lamb (e.g. 1978a, 1978b) in the 1970s. Many of these studies – particularly the modelling studies – have suggested that global SSTs play a decisive role (e.g. Folland *et al.*, 1986; Palmer, 1986; Rowell *et al.*, 1992; Ward, 1998; Giannini *et al.*, 2003). Our work confirms the importance of SSTs, but suggests that one relevant mechanism – perhaps the most important one – is more regional in scale. In particular, we have argued that the SST variability leads to changes in the position of the summer rainbelt over West Africa. Specifically, we have related the changes in SST to changes in cross-equatorial lower-tropospheric pressure gradients. With a cross-equatorial pressure gradient that changes from year to year, the location of the rainbelt also changes. During periods of reduced SST gradient, it will be located further south; during periods of enhanced gradient, it will be located further north. These arguments are consistent with the observed characteristics of wet and dry years in the Sahel.

Although we have evaluated the contribution of inertial instability to Sahel rainfall variability only on interannual time-scales, it probably plays a role on decadal time-scales as well. The contrast in pressure gradient between the wet and dry decades, seen in Figure 9, supports this idea, as do the contrasts in the speed of the westerlies and the latitude of zero absolute vorticity evident in Figure 16. Furthermore, the interhemispheric mode – the strong mode of variability in the tropical Atlantic – involves an SST dipole with opposite anomalies in the two hemispheres (Ruiz-Barradas *et al.*, 2000). This mode acts on decadal time-scales, and the phase of the SST dipole linked to wet conditions in the Sahel includes cold anomalies near to, and south of, the Equator. A preliminary analysis also shows the persistent presence of the low-level westerly jet in wet decades, and its absence in the recent dry decades (Nicholson, 2007c). Thus, the mechanism we have described appears to enhance rainfall variability in the Sahel on decadal time-scales as well.

Acknowledgements

We would like to acknowledge the assistance of Doug Klotter of Florida State University, and of Hai-Ru Chang of Georgia Tech. This research was supported by a grant from the National Science Foundation (ATM-0004479). We would like to thank the associate editor Richard Seager and two anonymous reviewers for their outstanding critical comments and suggestions for additional discussion. We would also like to thank Prof. Robert Hart, of FSU, and Dr Hai-Ru Chang, of the Georgia Institute of Technology, for their assistance in graphical production and valuable comments concerning the data sets utilized in our work.

References

Balas N, Nicholson SE, Klotter D. 2007. The relationship of rainfall variability in West Central Africa to sea-surface temperature fluctuations. *Int. J. Climatol.* **27**: 1335–1349.

Burpee RW. 1971. The origin and structure of easterly waves in the lower troposphere. *J. Atmos. Sci.* **29**: 77–90.

Camberlin P, Janicot S, Poccard I. 2001. Seasonality and atmospheric dynamics of the teleconnection between African rainfall and tropical SST: Atlantic vs. ENSO. *Int. J. Climatol.* **21**: 973–1005.

Carton JA, Cao X, Giese BS, da Silva AM. 1996. Decadal and interannual SST variability in the tropical Atlantic. *J. Phys. Oceanogr.* **26**: 1165–1175.

Chang P, Saravanan R, Ji L, Hegerl GC. 2000. The effect of local sea surface temperatures on atmospheric circulation over the tropical Atlantic sector. *J. Climate* **13**: 2195–2216.

Charney JG. 1971. Tropical cyclogenesis and the formation of the intertropical convergence zone. In: *Mathematical Problems of Geophysical Fluid Dynamics*, Reid WH (ed), pp. 355–368. *Lectures in Applied Mathematics* 14, American Mathematical Society.

Chelliah M, Bell GD. 2004. Tropical multidecadal and interannual climate variability in the NCEP-NCAR Reanalysis. *J. Climate* **17**: 1777–1803.

Covey D, Hastenrath S. 1978. The Pacific El Nino phenomenon in the Atlantic sector. *Mon. Weather Rev.* **106**: 1280–1287.

Czaja A, van der Vaart P, Marshall J. 2002. A diagnostic study of the role of remote forcing in tropical Atlantic variability. *J. Climate* **15**: 3280–3290.

Eltahir EAB, Gong C. 1996. Dynamics of wet and dry years in West Africa. *J. Climate* **9**: 1030–1042.

Emanuel K. 1995. On thermally direct circulations in moist atmosphere. *J. Atmos. Sci.* **52**: 1529–1534.

Enfield DB, and Mestas-Nuñez AM. 1999. Multiscale variabilities in global sea surface temperatures and their relationships with tropospheric climate patterns. *J. Climate* **12**: 2719–2733.

Flohn H. 1964. 'Intertropical convergence zone and meteorological equator'. WMO Technical Notes 64, 21 pp.

Folland CK, Palmer TN, Parker DE. 1986. Sahel rainfall and worldwide sea temperatures, 1901–1985. *Nature* **320**: 602–607.

Folland CK, Owen J, Ward MN, Colman A. 1991. Prediction of seasonal rainfall in the Sahel Region using empirical and dynamical methods. *J. Forecasting* **10**: 21–56.

Copyright © 2007 Royal Meteorological Society

Fontaine B, Bigot S. 1993. West African rainfall deficits and sea-surface temperatures. *Int. J. Climatol.* **13**: 271–285.
Fontaine B, Janicot S. 1992. Wind-field coherence and its variations over West Africa. *J. Climate* **5**: 512–524.
Fontaine B, Janicot S. 1996. Sea surface temperature fields associated with West African anomaly types. *J. Climate* **9**: 2935–2940.
Fontaine B, Janicot S, Moron V. 1995. Rainfall anomaly patterns and wind field signals over West Africa in August (1958–89). *J. Climate* **8**: 1503–1510.
Fontaine B, Trzaska S, Janicot S. 1998. Evolution of the relationship between near global and Atlantic SST modes and the rainy season in West Africa: statistical analyses and sensitivity experiments. *Clim. Dyn.* **414**: 353–368.
Giannini A, Saravanan R, Chang P. 2003. Oceanic forcing of Sahel rainfall on interannual to interdecadal time scales. *Science* **302**: 1027–1030.
Gong D, Wang S. 1999. Definition of Antarctic oscillation index. *Geophys. Res. Lett.* **26**: 459–462.
Grist JP. 2002. Easterly waves over Africa. Part I: The seasonal cycle and contrasts between wet and dry years. *Mon. Weather Rev.* **130**: 1337–1359.
Grist JP, Nicholson SE. 2001. A study of the dynamic factors influencing the interannual variability of rainfall in the West African Sahel. *J. Climate* **14**: 1337–1359.
Grist JP, Nicholson SE, Barcilon AI. 2002. Easterly waves over West Africa II. Observed and modelled contrasts between wet and dry years. *Mon. Weather Rev.* **130**: 212–225.
Hastenrath S. 1984. Interannual variability and the annual cycle: Mechanisms of circulation and climate in the tropical Atlantic sector. *Mon. Weather Rev.* **112**: 1097–1107.
Hastenrath S. 1990. Decadal-scale changes of the circulation in the tropical Atlantic sector associated with Sahel drought. *Int. J. Climatol.* **10**: 459–472.
Hastenrath SL, Greischar L. 1993a. Circulation mechanisms related to northeast Brazil rainfall anomalies. *J. Geophys. Res.* **98D**: 5093–5102.
Hastenrath SL, Greischar L. 1993b. Further work on prediction of northeast Brazil rainfall anomalies. *J. Climate* **6**: 743–758.
Hickey H, Weaver AJ. 2004. The Southern Ocean as a source region for Tropical Atlantic variability. *J. Climate* **17**: 3960–3972.
Holton JR, Wallace JM, Young JA. 1971. On boundary layer dynamics and the ITCZ. *J. Atmos. Sci.* **28**: 275–280.
Huang B. 2003. 'Remotely forced variability in the tropical Atlantic Ocean'. COLA Technical Rep. 147, 52 pp.
Huang B, Shukla J. 2005. Ocean-atmosphere interactions in the tropical and subtropical Atlantic Ocean. *J. Climate* **18**: 1652–1672.
Janicot A. 1992a. Spatiotemporal variability of West Africa rainfall. Part I: Regionalizations and typings. *J. Climate* **5**: 489–497.
Janicot A. 1992b. Spatiotemporal variability of West Africa rainfall. Part II: Associated surface and airmass characteristics. *J. Climate* **5**: 499–511.
Janicot S, Trzaska S, Poccard I. 2001. Summer Sahel-ENSO teleconnection and decadal time scale SST variations. *Clim. Dyn.* **18**: 303–320.
Janowiak J. 1988. An investigation of interannual variability in Africa. *J. Climate* **1**: 240–255.
Kalnay E, Kanamitsu M, Kistler R, Collins W, Deaven D, Gandin L, Iredell M, Saha S, White G, Woollen J, Zhu Y, Leetmaa A, Reynolds B, Chelliah M, Ebisuzaki W, Higgins W, Janowiak J, Mo KC, Ropelewski C, Wang J, Jenne R, Joseph D. 1996. The NCEP/NCAR 40-year reanalysis project. *Bull. Am. Meteorol. Soc.* **77**: 437–471.
Kanamitsu M, Krishnamurti TN. 1978. Northern summer tropical circulations during drought and normal rainfall months. *Mon. Weather Rev.* **106**: 331–347.
Kidson JW. 1977. African rainfall and its relation to the upper air circulation. *Q. J. R. Meteorol. Soc.* **103**: 441–456.
Kinter JL, Fennessy MJ, Krishnamurthy V, Marx L. 2004. An evaluation of the apparent interdecadal shift in the tropical divergent circulation in the NCEP-NCAR Reanalysis. *J. Climate* **17**: 349–361.
Kistler R Kalnay E, Collins W, Saha S, White G, Woollen J, Chelliah M, Ebisuzaki W, Kanamitsu M, Kousky V, van den Dool H, Jenne R, Fiorino M. 2001. The NCEP-NCAR 50-year Reanalysis Monthly means CD-ROM and Documentation. *Bull. Am. Meteorol. Soc.* **82**: 247–267.
Kraus EB. 1977a. Subtropical drought and cross equatorial energy transports. The seasonal excursion of the ITCZ. *Mon. Weather Rev.* **105**: 1009–1018.
Kraus EB. 1977b. The seasonal excursion of the Intertropical Convergence Zone. *Mon. Weather Rev.* **105**: 1052–1055.
Kushnir Y. 1994. Interdecadal variations in North Atlantic sea surface temperature and associated atmospheric conditions. *J. Climate* **7**: 2156–2159.
Lamb PJ. 1978a. Large-scale Tropical Atlantic surface circulation patterns associated with Subsaharan weather anomalies. *Tellus* **30**: 240–251.
Lamb PJ. 1978b. Case studies of tropical Atlantic surface circulation patterns during recent Sub-Saharan weather anomalies 1967 and 1968. *Mon. Weather Rev.* **106**: 482–491.
Lamb PJ. 1983. West African water vapor variations between recent contrasting Subsaharan rainy seasons. *Tellus* **A35**: 198–212.
Lamb P, Peppler R. 1992. Further case studies of tropical Atlantic surface atmospheric and oceanic patterns associated with Sub-Saharan drought. *J. Climate* **5**: 476–488.
Latif M, Grötzner A. 2000. The equatorial Atlantic oscillation and its response to ENSO. *Clim. Dyn.* **16**: 213–218.
Lettau H. 1956. Theoretical notes on the dynamics of the equatorial atmosphere. *Beitr. Phys. Atm.* **29**: 107.
Long M, Entekhabi D, Nicholson SE. 2000. Interannual variability in rainfall, water vapor flux, and vertical motion over Africa. *Journal of Climate* **13**: 3827–3841.
Lough JM. 1986. Tropical Atlantic sea surface temperatures and rainfall variations in Subsaharan Africa. *Mon. Weather Rev.* **114**: 561–570.
Mehta VM. 1998. Variability of the tropical ocean surface temperatures at decadal–multidecadal timescales. Part 1. The Atlantic Ocean. *J. Climate* **11**: 2351–2375.
Miller RL, Lindzen R. 1992. Organization of rainfall by an unstable jet with an application to African waves. *J. Atmos. Sci.* **49**: 1523–1540.
Mitchell TP, Wallace JM. 1992. On the annual cycle in equatorial convection and sea surface temperature. *J. Climate* **5**: 1140–1152.
Mo KC. 2000. Relationships between low-frequency variability in the Southern Hemisphere and sea surface temperature anomalies. *J. Climate* **13**: 3599–3610.
Mo KC, Häkkinen S. 2001. Interannual variability in the tropical Atlantic and linkages to the Pacific. *J. Climate* **14**: 2740–2762.
Moron V. 1994. Guinean and Sahelian rainfall anomaly indices at annual and monthly time scales (1930–1990). *Int. J. Climatol.* **14**: 325–341.
Murakami T, Wang B, Lyons SW. 1992. Summer monsoon over the Bay of Bengal and the eastern North Pacific. *J. Meteorol. Soc. Jpn* **70**: 191–210.
Newell RE, Kidson JW. 1984. African mean wind changes between Sahelian wet and dry periods. *J. Climatol.* **4**: 1–7.
Nicholson SE. 1980. The nature of rainfall fluctuations in subtropical West Africa. *Mon. Weather Rev.* **108**: 473–487.
Nicholson SE. 1981. Rainfall and atmospheric circulation during drought periods and wetter years in West Africa. *Mon. Weather Rev.* **109**: 2191–2208.
Nicholson SE. 1986. The spatial coherence of African rainfall anomalies: Interhemispheric teleconnections. *J. Clim. Appl. Meteorol.* **25**: 1365–1381.
Nicholson SE. 1989. 'Rainfall variability in Southern Africa and its relationship to ENSO and the Atlantic and Indian Oceans'. In: *Preprints of the Third International Conference on Southern Hemisphere Meteorology and Oceanography* (Buenos Aires, Argentina), pp. 366–367. American Meteorological Society.
Nicholson SE. 1993. An overview of African rainfall fluctuations of the last decade. *J. Climate* **6**: 1463–1466.
Nicholson SE. 1997. An analysis of the ENSO signal in the tropical Atlantic and western Indian Oceans. *Int. J. Climatol.* **17**: 345–375.
Nicholson SE. 2000. Land surface processes and Sahel climate. *Rev. Geophys.* **38**: 117–139.
Nicholson SE. 2007a. On the factors modulating the intensity of the tropical rainbelt over West Africa. *J. Climate* (in press).
Nicholson SE. 2007b. The intensity, location and structure of the tropical rainbelt over West Africa as a factor in interannual variability. *Int. J. Climatol.* (in press).
Nicholson SE. 2007c. A new picture of the structure of the "monsoon" and land ITCZ over West Africa. *Bull. Am. Meteorol. Soc.* (submitted).
Nicholson SE, Entekhabi D. 1987. Rainfall variability in Equatorial and Southern Africa: relationships with sea-surface temperatures along the southwestern coast of Africa. *J. Clim. Appl. Meteorol.* **26**: 561–578.
Nicholson SE, Grist JP. 2001. A simple conceptual model for understanding rainfall variability in the West African Sahel on

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj

interannual and interdecadal time scales. *Int. J. Climatol.* **21**: 1733–1757.
Nicholson SE, Grist JP. 2002. On the seasonal evolution of atmospheric circulation over West Africa and Equatorial Africa. *J. Climate* **16**: 1013–1030.
Nicholson SE, Kim J. 1997. The relationship of the El-Nino Southern Oscillation to African rainfall. *Int. J. Climatol.* **17**: 117–135.
Nicholson SE, Palao IM. 1993. A re-evaluation of rainfall variability in the Sahel. Part I: Characteristics of rainfall fluctuations. *Int. J. Climatol.* **13**: 371–389.
Nicholson SE. 2005. On the Question of the "Recovery" of the Ravis in the West African Sahel. *Journal of Arid Environments* **63**: 615–641.
Nicholson SE, Barcilon AI, Challa M. 2007. An analysis of West African dynamics using a linearized GCM. *J. Atmos. Sci.* (to appear).
Palastanga V, Vera CS, Piola AR. 2002. On the leading modes of sea surface temperature variability in the South Atlantic Ocean. *CLIVAR Exchanges* **7**(3–4): 12–16.
Palmer TN. 1986. The influence of the Atlantic, Pacific, and Indian Oceans on Sahel rainfall. *Nature* **322**: 251–253.
Philander G, Gu D, Lamert G, Li T, Halpern D, Lau N.-C, Pacanowski RC. 1996. Why the ITCZ is mostly north of the equator. *J. Climate* **9**: 2958–2972.
Poccard I, Janicot S, Camberlin P. 2000. Comparison of rainfall structures between NCEP/NCAR reanalysis and observed data over tropical Africa. *Clim. Dyn.* **16**: 897–915.
Rajagopalan B, Kushnir Y, Tourre YM. 1998. Observed decadal midlatitude and tropical Atlantic climate variability. *Geophys. Res. Lett.* **25**: 3967–3970.
Reed RJ, Norquist DC, Recker EE. 1977. The structure and characteristics of African easterly wave disturbances as observed during Phase III of GATE. *Mon. Weather Rev.* **105**: 317–333.
Robertson AW, Farrara JD, Mechoso CR. 2003. Simulations of the atmospheric response to South Atlantic sea surface temperature anomalies. *J. Climate* **16**: 2540–2551.
Rowell DP. 2001. Teleconnections between the tropical Pacific and the Sahel. *Q. J. R. Meteorol. Soc.* **127**: 1683–1706.
Rowell DP. 2003. The impact of Mediterranean SSTs on the Sahelian rainfall season. *J. Climate* **16**: 849–862.
Rowell DP, Folland CK, Maskell K, Owen JA, Ward MN. 1992. Modelling the influence of sea-surface temperatures on the variability and predictability of seasonal Sahel rainfall. *Geophys. Res. Lett.* **19**: 905–908.
Rowell DP, Milford JR. 1993. On the generation of African squall lines. *J. Climate* **6**: 1181–1193.
Rowell DP, Folland CK, Maskell K, Ward MN. 1995. Variability of summer rainfall over tropical north Africa (1906–92): observations and modelling. *Q. J. R. Meteorol. Soc.* **121**: 669–704.
Ruiz-Barradas A, Carton JA, Nigam S. 2000. Structure of interannual-to-decadal climate variability in the tropical Atlantic. *J. Climate* **13**: 3285–3297.
Semazzi FHM, Burns B, Lin N-H, Schemm J-K. 1996. A GCM study of the teleconnections between the continental climate of Africa and global sea surface temperature anomalies. *J. Climate* **9**: 2480–2497.
Servain J. 1991. Simple climatic indices for the tropical Atlantic Ocean and some applications. *J. Geophys. Res.* **96**: 15137–15146.
Smith TM, Reynolds RW, Livezey RE, Stokes DC. 1996. Reconstruction of historical sea surface temperatures using empirical orthogonal functions. *J. Climate* **9**: 1403–1420.
Taylor CM, Lebel T. 1998. Observational evidence of persistent convective scale rainfall patterns. *Mon. Weather Rev.* **126**: 1597–1607.
Tomas RA, Webster PJ. 1997. The role of inertial instability in determining the location and strength of near-equatorial convection. *Q. J. R. Meteorol. Soc.* **123**: 1445–1482.
Tourre YM, Rajagopalan B, Kushnir Y. 1999. Dominant patterns of climate variability in the Atlantic Ocean region during the last 136 years. *J. Climate* **12**: 2285–2299.
Venegas SA, Mysak LA, Straub DN. 1997. Atmosphere-ocean coupled variability in the South Atlantic. *J. Climate* **10**: 2904–2920.
Wagner RG. 1996. Mechanisms controlling variability of the interhemispheric sea surface temperature gradient in the tropical Atlantic. *J. Climate* **9**: 2010–2019.
Wang J, Carton JA. 2003. Modeling climate variability in the tropical Atlantic atmosphere. *J. Climate* **16**: 3876–3976.
Wang B, Li T. 1994. Convective interaction with boundary layer dynamics in the development of a tropical intraseasonal system. *J. Atmos. Sci.* **51**: 1386–1400.
Wang B, Wang Y. 1999. Dynamics of the ITCZ-equatorial cold tongue complex and causes of the latitudinal climate asymmetry. *J. Climate* **12**: 1830–1847.
Ward MN. 1998. Diagnosis and short-lead time prediction of summer rainfall in tropical North America at interannual and multidecadal timescales. *J. Climate* **11**: 3167–3191.
Wolter K. 1989. Modes of tropical circulation, Southern Oscillation, and Sahel rainfall anomalies. *J. Climate* **2**: 149–172.
Zebiak SE. 1993. Air-sea interaction in the equatorial Atlantic region. *J. Climate* **6**: 1567–1586.
Zheng X, Eltahir EAB, Emanuel KA. 1999. A mechanism relating spring sea surface semperature in the Tropical Atlantic and rainfall variability over West Africa. *Q. J. R. Meteorol. Soc.* **125**: 1129–1164.

Copyright © 2007 Royal Meteorological Society

DOI: 10.1002/qj