Dynamics of Wet and Dry Years in West Africa

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(Manuscript received 28 April 1995, in final form 1 October 1995)

ABSTRACT

This paper proposes a theoretical framework for describing interannual climatic variability over West Africa. The dynamical theory of zonally symmetrical thermally direct circulations suggests that a meridional monsoon circulation must develop over any tropical region (off the equator) when the absolute vorticity near the tropopause reaches a threshold value of zero. However, for a moist atmosphere that satisfies a quasi-equilibrium balance between moist convection and the radiative forcing, the absolute vorticity at upper-tropospheric levels is a function of both latitude and the meridional distribution of boundary-layer entropy. Hence, the onset of a monsoon circulation depends in a nonlinear fashion on these two factors. The theory predicts that a flat distribution of entropy does not drive any circulation and that a relatively large gradient of entropy should drive a strong monsoon circulation. The location of the region of West Africa, relatively close to the equator, dictates that the dynamics of a monsoon over that region are relatively sensitive to interannual fluctuations in the meridional gradient of boundary-layer entropy. Here, we present observations on entropy and wind over West Africa during the monsoon seasons of 1958 and 1960. The following observations were consistent with the proposed relationship between boundary-layer entropy and the monsoon circulation: a large meridional gradient of boundary-layer entropy, a healthy monsoon, and wet conditions over the Sahel region were observed in 1958; and a nearly flat distribution of entropy, very weak circulation, and relatively dry conditions were observed in 1960. Moreover, the proposed theoretical relationship between the meridional gradient of boundary-layer entropy and the monsoon circulation over West Africa is consistent with the empirical observations of sea surface temperature anomalies (SSTAs) in the tropical Atlantic and rainfall in the Sahel region. Theoretically, a cold (warm) SSTA in the region located south of the West African coast should favor a large (small) meridional gradient of entropy, a strong (weak) monsoon circulation, and wet (dry) conditions in the Sahel. A large body of observations confirms that cold (warm) SSTAs off the southern coast of West Africa are associated with wet (dry) years in the Sahel region.

1. Introduction

The continuing drought in West Africa has had a significant impact on the social and economic conditions of several African nations. The rainfall levels during the last two decades have been significantly low compared to the long-term average for this century (Fig. 1). Based on instrumental records, the recent rainfall decline in West Africa is unprecedented in duration, intensity, spatial character, and seasonal expression (Farmer and Wigley 1985). In addition, the 1980s constitute the driest decade of the century over West Africa (Nicholson 1993). Understanding the causes of this African drought is the motivation for this study. However, these recent rainfall fluctuations are not the subject of this study. Instead, we focus on the natural variability in rainfall and circulation over West Africa prior to the onset of the current drought episode. Understanding this natural variability will indeed help in explaining the mechanisms behind the current African drought.

The climate over West Africa has several unique features. The annual rainfall is almost constant along each latitude. However, annual rainfall decreases sharply from the south to the north with a gradient of about 1 mm per kilometer, from about 1500 mm near the coast at 5°N to about 100 mm along the border with the Sahara Desert at about 20°N. The rainy season occupies, roughly, the months of June, July, August, and September. Hence, the length of the dry season, though variable, is generally longer than that of the wet season. The intensity of the current drought in West Africa is different for different months of the rainy season. As noted by Farmer and Wigley (1985), the recent decline in rainfall is most evident during the months of August and September. A smaller and less persistent decline in rainfall levels is observed during the months of June and July. This observation is particularly important since August and September represent the late half of the rainy season, when the monsoon circulation is likely to dominate the dynamics of the atmosphere over West Africa.

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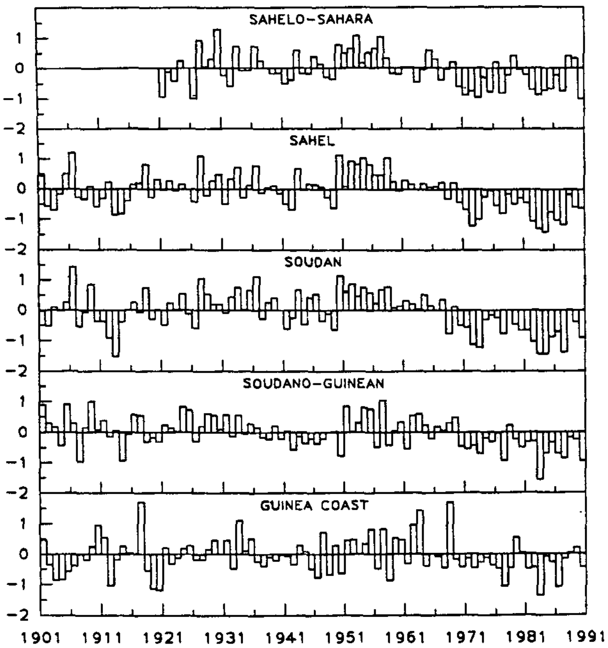

FIG. 1. Rainfall fluctuations in West Africa (1901–1990), expressed as regionally averaged standardized departures (Nicholson 1993).

The current deficit in rainfall levels is observed over most of the region of West Africa. The observed rainfall series, shown in Fig. 1, indicates a significant decline for all the subregions. In particular, we note the significant decline in rainfall even in the humid region of the Guinea coast. However, the most significant changes in rainfall occurred in the Sahel and Soudan regions. Similar persistent drought conditions are observed in other hydrologic series. For example, a significant decline is observed in the annual flows of the Senegal River and the Niger River; these observations are reported by Farmer and Wigley (1985). In addition, the evident decline in the level of Lake Chad is consistent with the drought conditions that have persisted for the last few decades. The Senegal River, the Niger River, and Lake Chad are the most significant hydrologic systems in West Africa.

Several studies have been conducted to investigate the causes and mechanisms of this African drought. Nicholson (1989) presents several hypotheses and views from the literature on the causes of the African drought. The most noted hypothesis is that of Charney (1975), which relates this drought to the changes in land cover and surface albedo along the southern border of the Sahara Desert. Another view on this problem relates the drought to anomalous large-scale patterns of atmospheric circulation and/or global patterns of SST distribution [for examples see Kidson (1977) and Folland et al. (1986)]. A third hypothesis suggests that the drought conditions in West Africa are triggered by anomalous large-scale forcing and then sustained by local feedbacks that involve land surface processes (Nicholson 1989).

An understanding of the natural variability in any climate system is a prerequisite for explaining the causes of drought. The dynamics of the monsoon circulation are intimately coupled to the natural variability of rainfall and climate over West Africa. Until recently, our understanding of atmospheric circulations in the Tropics, including monsoons, was very limited. Although some progress has been made, several theoretical issues on the dynamics of atmospheric circulations in the Tropics have not yet been resolved or verified. In this paper, the importance of a theoretical understanding of the natural variability in the monsoon circulation is emphasized as being necessary before the changes in the circulation and rainfall that have occurred over West Africa can be explained. Hence, the focus of this paper is not the climate as observed during recent years, but instead, we focus on the natural variability of the system before the onset of the current drought around the late 1960s. The dynamics of the monsoon circulation over West Africa during the late 1950s and early 1960s are studied using observations of several atmospheric variables: wind, temperature, humidity, and rainfall. The years 1958 and 1960 are selected as the wettest and the driest years in the Sahel region, respectively, during the period from 1958 to 1967, inclusively. The contrast in the dynamics of monsoons between these two wet and dry years is investigated in an attempt to understand the natural variability in the dynamics of monsoons over that region. Hopefully, the insight gained from this analysis should help to identify the potential causes of the current drought in West Africa.

In summary, this paper suggests that the dynamics of the monsoon circulation over West Africa are regulated by the meridional gradient of boundary-layer entropy. This argument is supported by theoretical and observational analysis. Moreover, observations of SST distribution, such as those reported by Lamb (1978) and Lough (1986), suggest that the variability in SST distribution in the tropical Atlantic plays a significant role in the regional climate over West Africa. The theory proposed here for describing the monsoon circulation over West Africa implies a certain relationship between boundary-layer entropy in the tropical Atlantic (which is directly related to SST) and the monsoon circulation in West Africa (which is directly related to rainfall). This theoretical relationship is compared to observations of the relationship between SST in the tropical Atlantic and rainfall in the Sahel in order to verify this aspect of the proposed relationship between entropy, circulation, and rainfall.

In the following, we first briefly review some previous studies on drought and rainfall variability in West

Africa. Then, a theoretical framework for analyzing the dynamics of monsoons is presented and discussed. The heart of this framework is the proposed relationship between entropy, monsoon circulation, and rainfall in West Africa. The main section of this paper presents analysis of atmospheric data from the region during a wet and a dry year. The roles of SST and land cover are discussed in light of the proposed relationship between entropy, circulation, and rainfall over West Africa. The last section includes the conclusions of this study.

2. Review

Rainfall variability over West Africa has been the focus of several theoretical and observational studies. The most noted among them is the study of Charney (1975), which presents a hypothesis for explaining the occurrence of droughts in this region by relating the regional circulation to the dynamics of the Sahara Desert. Charney's theory is based on the concept that deserts work as radiative sinks of heat, such that the loss of heat due to the emission of planetary radiation is balanced by adiabatic warming due to subsidence. Under these conditions, Charney (1975) suggests that removal of vegetation, such as is caused by overgrazing at the desert border, increases surface albedo, causing an additional radiative cooling. The latter can only be balanced by additional adiabatic warming and enhancement of sinking motion. Charney argues that this enhancement of the sinking motion could push the ITCZ southward and thus reduce rainfall. According to this mechanism, the drought in West Africa may be related to the changes in vegetation near the border with the Sahara Desert.

The main weakness of Charney's mechanism is the fact that vegetation at the desert border grows in response to rain and closely follows the seasonal cycle of rainfall. The vegetation that exists by the end of the short rainy season may not survive more than few months into the long dry season. Recent satellite pictures of vegetation in Africa, such as those presented by Tucker et al. (1991), illustrate clearly the close relationship between rainfall and vegetation in this region. Under these conditions, the vegetative cover lacks the memory that is needed to carry information between successive years, and thus, the Charney mechanism fails to explain the persistence of the continuing drought in West Africa.

During the last two decades, several other studies have focused on the African drought. Walker and Rowntree (1977) investigate the effect of soil moisture on the circulation and rainfall over West Africa using a modeling approach. The results of Lamb (1983) indicate that the sub-Saharan drought does not appear to be associated with the northward supply of unusually dry surface air to West Africa from the tropical Atlantic. Using rawinsondes observations, Kidson (1977) reports a weakening in the meridional circulation in August during dry years and points out that the easterly jet at 200 mb over West Africa is stronger in wet years than in dry ones. Newell and Kidson (1984) compare African mean wind changes between Sahelian wet and dry periods. Hastenrath (1984) discusses the mechanisms of atmospheric circulation and climate in the tropical Atlantic. These mechanisms are related to the observed climate variability over West Africa.

The role of soil moisture, evaporation, and albedo in the climate of West Africa has also been investigated by several modeling studies. Yeh et al. (1984) consider the effect of soil moisture on short-term climate and hydrology in several regions, including West Africa. Sud and Fennessy (1984) use a numerical model to study the influence of evaporation on the July circulation in several semiarid regions, including the Sahel. The study of Sud and Molod (1988) simulates the influence of anomalous evaporation and surface albedo on the July circulation over West Africa, using a general circulation model. The results of Cunnington and Rowntree (1986) point to the importance of initial conditions of soil moisture and atmospheric conditions in numerical simulations of climate over West Africa. Kitoh et al. (1988) use a general circulation model to investigate the influence of soil moisture and albedo changes over Africa on the summer climate. The role of soil wetness distribution in short-range rainfall forecasting is also studied by Powell and Blondin (1990). The results of all these studies point to the importance of land surface fluxes in the dynamics of the regional climate over West Africa.

The relation of rainfall over West Africa to the distribution of SST in the tropical Atlantic has been the focus of several studies. Lamb (1978) investigates the large-scale tropical Atlantic surface circulation patterns that are associated with sub-Saharan weather anomalies. This study involves analysis of rainfall records over the Sahel region and the distribution of SST over the Atlantic Ocean. The circulation and sea surface temperature anomalies (SSTAs) associated with wet and dry years are estimated based on observations. The study of Folland et al. (1986) discusses the relation between Sahel rainfall and worldwide SST distribution. They argue that the persistently wet and dry years in the Sahel region are strongly related to contrasting patterns of SSTA at the global scale. The use of SST observations for the purpose of forecasting rainfall in the Sahel is explored by Owen and Ward (1989), using a global SST dataset. Lough (1986) identifies different patterns of normalized SST departures in the tropical Atlantic, using principal component analysis. The same study correlates these different patterns to rainfall in the Sahel region.

The research on the West African drought is reviewed by Farmer and Wigley (1985) and Nicholson

(1989), but as pointed out by Peixoto and Oort (1993), there is no general accepted theory regarding the cause or the physical mechanisms responsible for this continuing geophysical phenomenon. Until recently, our understanding of the mechanisms and processes responsible for the natural variability in the monsoon circulation over West Africa (or elsewhere) was quite limited. But, indeed, an understanding of these processes is necessary for explaining the trends in regional rainfall over West Africa. In fact, this need is recognized in the same classic paper by Charney (1975), which notes that very little was understood at that time regarding the processes controlling large-scale circulation over the region.

3. Theory

Observations of several atmospheric variables such as rainfall, temperature, and humidity suggest that climate over West Africa can be approximated by a zonally symmetrical description. The dynamical theories of thermally direct, zonally symmetrical circulations in the tropical atmosphere are discussed by Held and Hou (1980), Lindzen and Hou (1991), Plumb and Hou (1992), and Emanuel et al. (1994). The dynamical theory for the response of a dry zonally symmetrical atmosphere to subtropical thermal forcing is developed by Plumb and Hou (1992). Below a threshold value of the thermal forcing, the atmosphere adopts a steady state of thermal equilibrium with no meridional flow. For supercritical forcing, the thermal equilibrium breaks down, and a strong meridional circulation develops. In the case of a dry atmosphere, Plumb and Hou (1992) describe the subtropical thermal forcing by specifying the distribution of equilibrium temperature. In a moist atmosphere that satisfies a quasi-equilibrium balance between moist convection and large-scale radiative forcing, the vertical distribution of saturation entropy is uniform. Hence, the subtropical thermal forcing is uniquely related to the meridional distribution of boundary-layer entropy. This quasi-equilibrium balance is assumed here in describing the atmosphere over West Africa.

For a zonally symmetrical circulation, the thermal wind relation is described by

$$\frac{\partial u}{\partial p} = \frac{1}{f}\frac{\partial \alpha}{\partial y}, \tag{1}$$

where u is zonal wind, p is pressure, f is the Coriolis parameter, α is specific volume, and y is the distance in the meridional direction. The Maxwell's relations, [see Emanuel (1994)], imply that

$$\left(\frac{\partial \alpha}{\partial y}\right)_p = \left(\frac{\partial \alpha}{\partial s^*}\right)_p \frac{\partial s^*}{\partial y} = \left(\frac{\partial T}{\partial p}\right)_{s^*} \frac{\partial s^*}{\partial y}, \tag{2}$$

where s^* is saturation entropy, $s^* = c_p \ln(\theta_e^*)$. Here c_p is specific heat capacity at constant pressure, and θ_e^* is the equivalent potential temperature of air if saturated at the same pressure and temperature. Assuming a moist adiabatic lapse rate and integrating Eq. (1) from the surface where wind is assumed to be zero, the following relation is developed:

$$u_t = -\frac{1}{f}(T_0 - T_t)\frac{\partial s_b}{\partial y}, \tag{3}$$

where u_t and T_t are wind and temperature at the tropopause. T_0 is surface temperature, and s_b ($\approx s^*$) is boundary-layer entropy. The absolute vorticity at the tropopause is given by

$$\eta_t = f - \left(\frac{\partial u_t}{\partial y}\right) = f + \frac{\partial}{\partial y}\left(\frac{1}{f}(T_0 - T_t)\frac{\partial s_b}{\partial y}\right). \tag{4}$$

Under the assumption of moist neutral atmospheric conditions. Equation (4) specifies the relationship between absolute vorticity at the tropopause and the meridional distribution of boundary-layer entropy (Emanuel et al. 1994). Depending on the distribution of entropy, either of two possible regimes may dominate the dynamics of the tropical atmosphere: a radiative–convective equilibrium regime or an angular momentum conserving regime (Plumb and Hou 1992). A radiative–convective equilibrium regime should exist if the absolute vorticity at the tropopause has the same sign as the Coriolis parameter f. This condition implies

$$1 + \frac{1}{f^2}\frac{\partial}{\partial y}\left((T_0 - T_t)\frac{\partial s_b}{\partial y}\right) - \frac{\beta}{f^3}(T_0 - T_t)\frac{\partial s_b}{\partial y} > 0, \tag{5a}$$

where β is $\partial f/\partial y$. The derivation of the above equations assumes a geostrophic balance. A similar set of equations can be developed assuming a gradient wind balance, following Plumb and Hou (1992) and Emanuel (1995), which results in a similar condition:

$$\left(\frac{\partial}{\partial \varphi}\left[\frac{\cos^3\varphi}{\sin\varphi}(T_0 - T_t)\frac{\partial s_b}{\partial \varphi}\right]\right) + 4\Omega^2 a^2 \cos^3\varphi \sin\varphi > 0, \tag{5b}$$

where φ is latitude, and Ω is the angular velocity of the earth.

When condition (5) is violated, the absolute vorticity is close to zero, and an angular momentum conserving regime should dominate the dynamics, resulting in a meridional circulation. Hence, the distribution of entropy in the atmospheric boundary-layer controls which of these two different regimes dominates the dynamics. Condition (5) quantifies a threshold that has to be achieved for the onset of a meridional (monsoon) circulation. The physical variables that control this threshold behavior are the gradient and the second derivative of the meridional distribution of boundary-layer entropy. Since absolute vorticity is proportional to the gradient of angular momentum, a zero absolute vorticity in the upper troposphere is a signature of an angular momentum conserving regime. The latter describes the monsoon circulation.

The region of West Africa is located at around 10°N. At this latitude, the Coriolis parameter and planetary vorticity are relatively small compared to the values of the same parameters at other tropical regions where monsoon circulations are also important. For example, the Indian monsoon occurs in a region located at around 25°N. This fact makes it relatively easy for meridional circulations to develop over this region since a relatively small magnitude of relative vorticity would achieve cancellation of planetary vorticity. However, the parameter β, which describes the rate of change in the Coriolis parameter with respect to latitude, is relatively large for West Africa compared to India. These geometrical facts are summarized in Table 1. They have several important consequences on the dynamics of monsoon circulations. First, Eq. (3) suggests that the zonal wind at upper levels in West Africa is relatively more sensitive to variability in the meridional gradient of entropy. This variability could be natural or anthropogenic in nature. Second, Table 1 and condition (5a) suggest that the occurrence of meridional circulations over West Africa is relatively sensitive to the distribution of entropy. For example, a unit change in the gradient of boundary-layer entropy over West Africa is likely to contribute to the left-hand side of condition (5a) *fifteen times* the corresponding contribution that would result from the same unit change of entropy gradient applied over India. Hence, the dynamics of monsoons over West Africa are relatively sensitive to the meridional gradient of boundary-layer entropy. This sensitivity suggests that the climate of West Africa is relatively vulnerable to the regional and global factors that may affect the meridional gradient of boundary-layer entropy.

The dynamical theory of monsoon circulations provides a general theoretical framework for understanding the natural variability in the dynamics of the atmospheric circulation over West Africa. As argued in the introduction of this paper, an understanding of these dynamics is a prerequisite for explaining any regional climate change, such as that indicated by the current drought. In the following section, we test these concepts by analyzing atmospheric data on boundary-layer entropy and wind over West Africa for wet and dry years.

TABLE 1. A comparison between the geometrical parameters relevant to monsoons [condition (5a)] for West Africa and India.

Variable	Units	West Africa	Indian region
f	$\sec^{-1}$	2.5E − 5	6.1E − 5
$\frac{1}{f^2}$	$\sec^2$	16.0E8	2.6E8
$\frac{\beta}{f^3}$	$\sec^3\ m^{-1}$	55.8E2	3.6E2

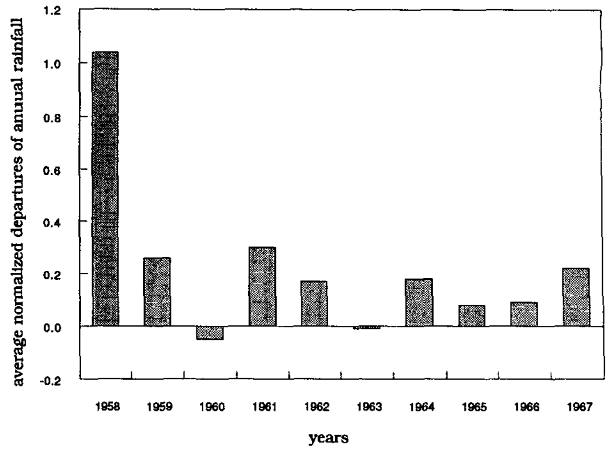

FIG. 2. Rainfall fluctuations in the Sahel region (1958–1967), expressed as regionally averaged standardized departures [based on the data of Nicholson (1994)].

4. Observations

The data used in this study are a subset of the Geophysical Fluid Dynamics Laboratory (GFDL) atmospheric circulation data (Oort 1983). This monthly dataset has a resolution of 2.5° in the meridional direction and 5° in the zonal direction at 11 pressure levels and covers the period from 1958 to 1989. The first ten years of the data, from 1958 to 1967, represent an analysis of rawinsondes records for 0000 UTC. The rest of the data are based on more than one daily observation, taken at different times of the day. Hence, although the GFDL data for the period from 1958 to 1989 is inhomogeneous, the subset of the data for the period from 1958 to 1967 is actually rather homogeneous. The years selected for this analysis are 1958, which was the wettest year in the Sahel region for the period from 1958 to 1967, and 1960, which was the driest year in the same period [see Fig. 2, which is based on the data of Nicholson (1994)]. Indeed, the year 1958 was wetter than any other year between 1959 and the present.

The observations of zonal and meridional wind at 5°E averaged for August and September of 1958 are shown in Figs. 3a and 3b, respectively. Similarly, Figs. 3c and 3d show the observations of zonal and meridional wind at the same location for August and September of 1960. The zonal wind at the equator is significantly stronger in August and September of 1958 than the zonal wind observed during the same period in

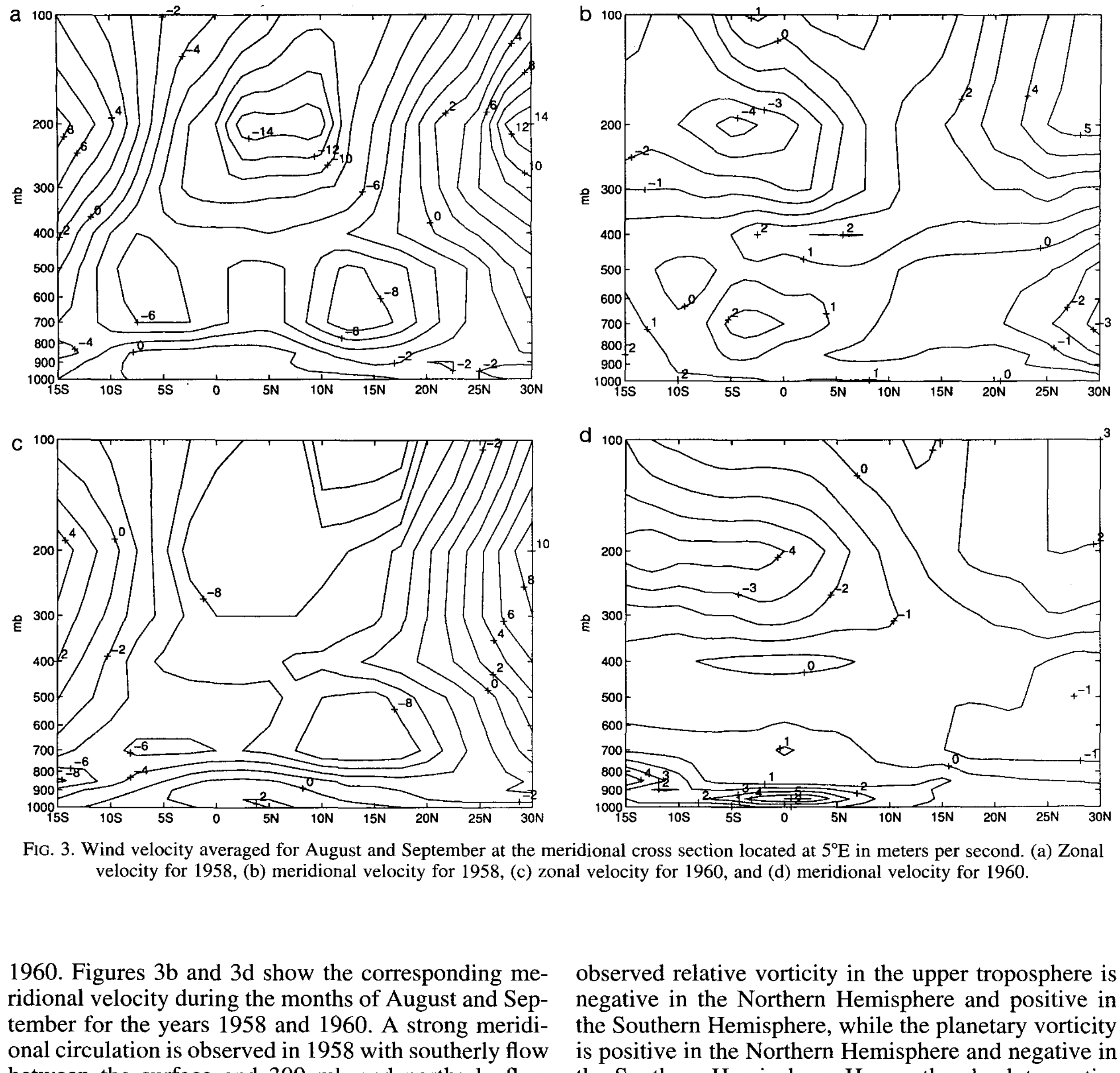

FIG. 3. Wind velocity averaged for August and September at the meridional cross section located at 5°E in meters per second. (a) Zonal velocity for 1958, (b) meridional velocity for 1958, (c) zonal velocity for 1960, and (d) meridional velocity for 1960.

1960. Figures 3b and 3d show the corresponding meridional velocity during the months of August and September for the years 1958 and 1960. A strong meridional circulation is observed in 1958 with southerly flow between the surface and 300 mb and northerly flow between 300 and 100 mb. The meridional circulation and, in particular, the southerly flow in the middle troposphere in August and September of 1960 are relatively weak. These observations present a clear contrast in the conditions of the monsoon: a healthy circulation in 1958 compared to a weak circulation in 1960.

The distribution of relative vorticity is related to the distribution of wind. The relative vorticity in the longitudinal cross section located at 2.5°E is computed from the observations of zonal and meridional wind. Figures 4 and 5 show the relative vorticity at 2.5°E in August and September of 1958 and 1960, respectively. At about 200 mb, the magnitude of the relative vorticity is much larger in 1958 compared to 1960. This is particularly true at around 5°S and 15°N. Note that the observed relative vorticity in the upper troposphere is negative in the Northern Hemisphere and positive in the Southern Hemisphere, while the planetary vorticity is positive in the Northern Hemisphere and negative in the Southern Hemisphere. Hence, the absolute vorticity, that is, the sum of the planetary vorticity and the relative vorticity, should approach zero at locations in the Northern Hemisphere, where the magnitude of relative vorticity is significant enough to cancel planetary vorticity. The cancellation between the relative vorticity and the planetary vorticity can be seen clearly at 200 mb. The relative, planetary, and absolute vorticity distributions at 200 mb averaged for the months of August and September are plotted in Figs. 6a and 6b for the years 1958 and 1960, respectively. The absolute vorticity distribution over a large region in West Africa is flatter in 1958 than the corresponding distribution for 1960. This is particularly true for the region extending between 5° and 15°N. In this region, while the absolute vorticity for 1960 increases from about 1×10^{-5} s^{-1}

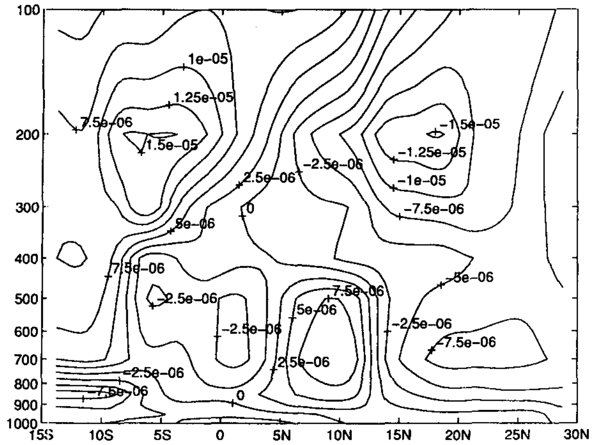

FIG. 4. Relative vorticity (in $\sec^{-1}$) averaged for August and September at the meridional cross section located at 2.5°E for 1958.

at 5°N to about $3 \times 10^{-5}\ s^{-1}$ at 15°N, the corresponding distribution of absolute vorticity for 1958 increases from about $1 \times 10^{-5}\ s^{-1}$ at 5°N to only about $2 \times 10^{-5}\ s^{-1}$ at 15°N. Overall, the magnitude of absolute vorticity at 200 mb over West Africa for 1958 is not significantly different from zero. This is indeed true if we consider the limited accuracy of wind observations and the fact that averaging took place over the long period of two months. Since absolute vorticity is proportional to the gradient of angular momentum, the observation of small value of absolute vorticity during August and September of 1958 over West Africa is consistent with dominance of an angular momentum conserving regime. The latter is a sign of a healthy monsoon circulation.

The theory presented in section 3 suggests that the dynamics of monsoons are related to the distribution of boundary-layer entropy. According to the theory, we would expect a flatter distribution of boundary-layer entropy in 1960 than in 1958. The data on temperature and humidity for the pressure levels 1000 and 950 mb, are used to estimate boundary-layer entropy. The latter is averaged in the zonal direction between 10°E and 10°W. Figures 7a and 7b show the entropy distributions between 15°S and 30°N during the month of July in 1958 and 1960, respectively. These observations are representative of the conditions prior to the onset of the monsoon. Similarly, Figs. 7c and 7d show the entropy distributions between 15°S and 30°N in the months of August and September, during the monsoon season, in the years 1958 and 1960, respectively. Indeed, the observations indicate a flatter distribution of entropy in 1960 than in 1958. These results are consistent with the other observations on wind and vorticity and are indeed consistent with the theory of section 3.

The threshold behavior of monsoon circulations that is described by condition (5) is tested in this section using the observations of boundary-layer entropy during July of 1958 and 1960. A critical distribution of boundary-layer entropy can be obtained by setting the left-hand side of (5b) equal to zero and then integrating the resulting equation with respect to latitude. (Emanuel 1995). The resulting distribution of boundary-layer entropy is a function of the magnitude and latitude of maximum boundary-layer entropy. Figures 8a and 8b show the theoretical critical distributions of boundary-layer entropy and the corresponding observations for July of the years 1958 and 1960, respectively. The solid line represents the theoretical distribution that matches the observations in the magnitude and location of the maximum s_b. For 1958, Fig. 8a shows a maximum s_b of about 5868 J/Kg/K located at about 15°N, while for 1960, Fig. 8b shows a maximum s_b of about 5857 J/Kg/K located at about 12.5°N. The magnitude of the gradient of s_b in July of 1958 clearly exceeds the critical gradient that marks the onset of the monsoon, while the corresponding gradient in July of 1960 is flatter than the critical gradient. Moreover, further analysis carried out on these same observations indicate that the magnitude of the term in condition (5) that includes the first derivative of entropy is large and that the contribution of the other term that includes the second derivative of entropy is relatively small. These entropy observations for 1958 and 1960 suggest that in the wet year, the angular momentum conserving regime should dominate the dynamics of the tropical atmosphere over West Africa, and in the dry year, the radiative-convective equilibrium regime should dominate the dynamics over the same region.

In summary, the observations from West Africa are consistent with the proposed relationship between the dynamics of monsoons over West Africa and the gradient of boundary-layer entropy. A healthy monsoon circulation is observed in the relatively wet year 1958.

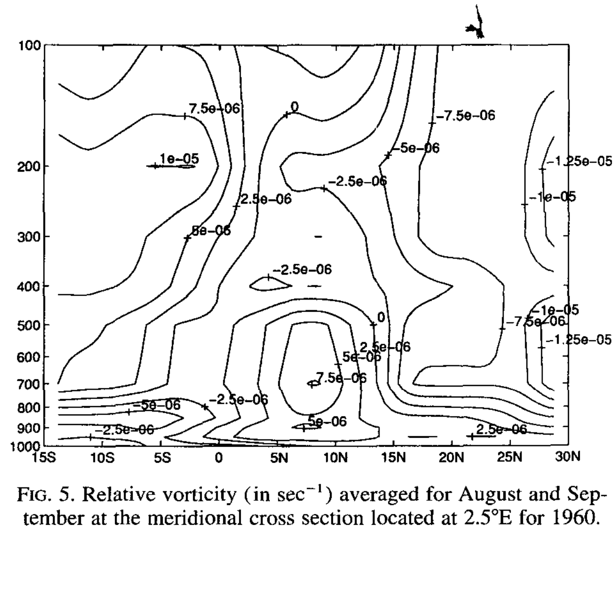

FIG. 5. Relative vorticity (in $\sec^{-1}$) averaged for August and September at the meridional cross section located at 2.5°E for 1960.

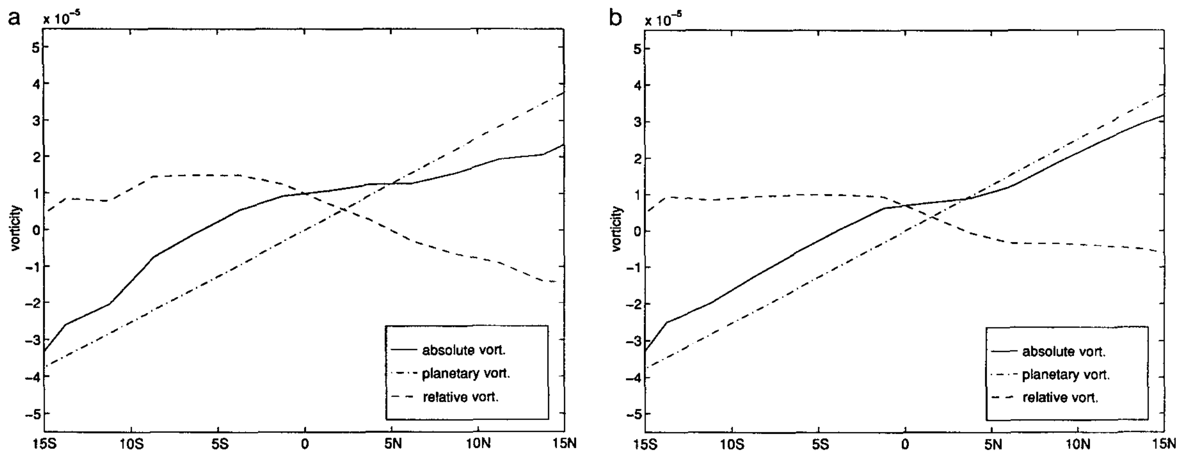

FIG. 6. Planetary, relative, and absolute vorticity (in $second^{-1}$) at 200 mb and 2.5°E for (a) 1958 and (b) 1960.

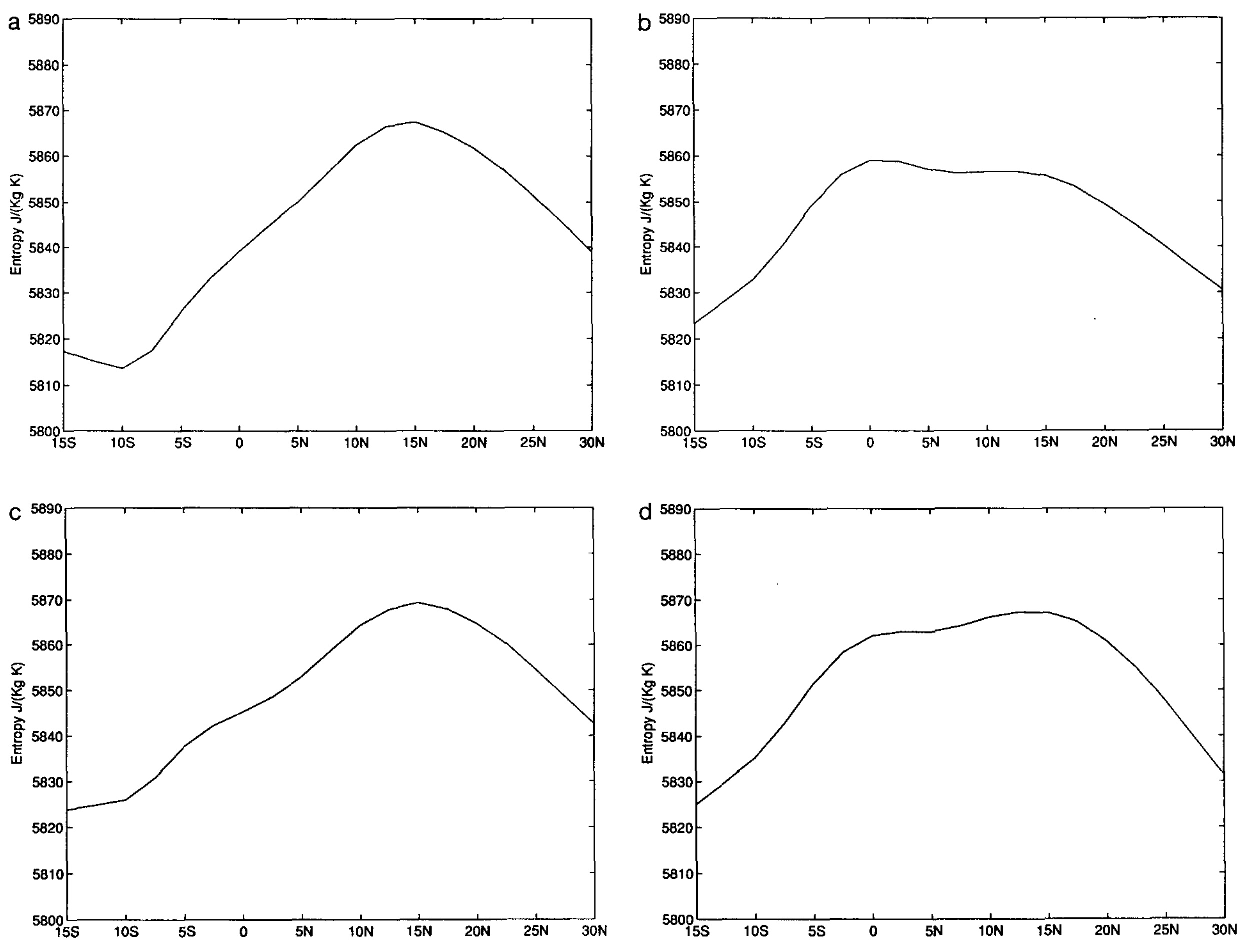

FIG. 7. The meridional distribution of entropy in the boundary-layer (1000 and 950 mb) averaged for July and averaged between 10°E and 10°W. Units are joules per kilogram per kelvin. (a) July 1958, (b) July 1960, (c) August and September 1958, and (d) August and September 1960.

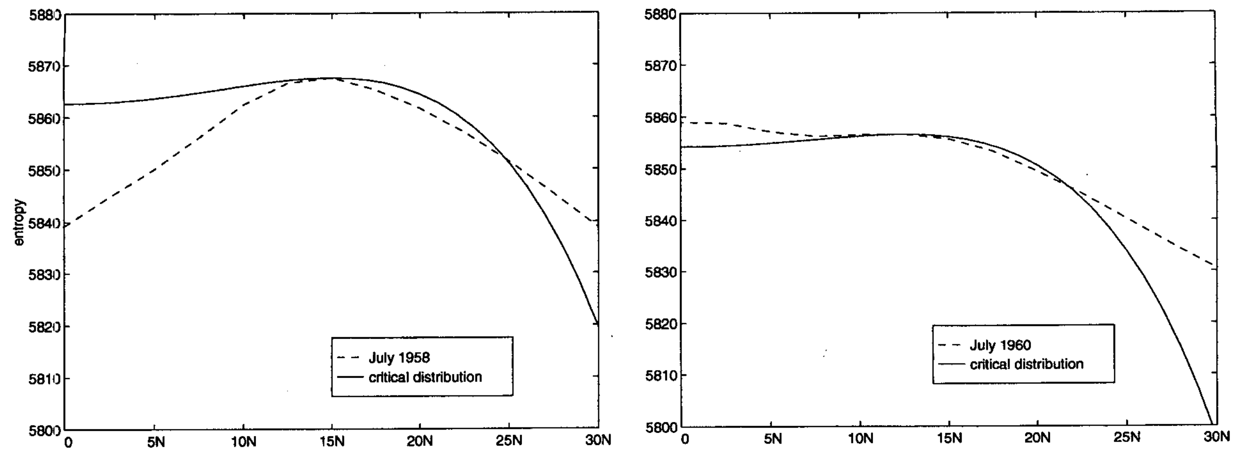

FIG. 8. The critical meridional distribution of boundary-layer entropy in comparison to the corresponding observations for the years (a) 1958 and (b) 1960. Units are joules per kilogram per kelvin.

The observations for the same year show a relatively large meridional gradient of boundary-layer entropy. The tendency of the tropical atmosphere in the rainy season toward achieving a quasi equilibrium between the radiative forcing and moist convection would favor a uniform vertical distribution of saturation entropy. Moist convection is responsible for coupling the monsoon dynamics in the upper troposphere with the boundary-layer entropy. The initial rainfall resulting from local convective storms heats the upper troposphere and results in a negative relative vorticity near the tropopause. In 1958, the observed magnitude of this negative relative vorticity is significant enough to result in near cancellation of planetary vorticity over a large area. The dynamical theory of monsoon circulations predicts that a vanishing absolute vorticity near the tropopause should lead to dominance of an angular momentum conserving regime. The latter favors a healthy monsoon circulation that provides a large-scale forcing for further rainfall.

During the relatively dry year 1960, initially, the diabatic heating due to water vapor condensation in the upper troposphere is not significant enough to cause a large negative relative vorticity near the tropopause. As a result, no cancellation of planetary vorticity is observed near the tropopause. Hence, instead of eventually developing an angular momentum conserving regime, a radiative–convective equilibrium regime seems to dominate the dynamics of the atmosphere over West Africa in that year. Under such conditions, most of the rainfall results from convective storms that are forced locally.

The results of the analysis on wind are in general agreement with some of the observations reported by Kidson (1977). The latter study includes analysis of wind over West Africa and emphasizes that the decline in rainfall over the Sahel region has evidently been paralleled by a weakening of the Northern Hemisphere circulation. The same study points to the fact that dry years over the Sahel are associated with a weaker easterly jet. However, the analysis of Kidson (1977) does not deal with entropy or upper-tropospheric vorticity. The analysis in this paper also suggests that the weakening of the easterly jet in relatively dry years is associated with the relatively weak gradient of boundary-layer entropy and that the weakening of the meridional circulation is associated with a relatively significant magnitude of absolute vorticity near the tropopause. As argued in the theory of section 3, development of a strong meridional circulation requires a vanishing absolute vorticity at upper levels.

The natural variability in the climate of West Africa is at least partly regulated by the meridional gradient of boundary-layer entropy, which is a function of several factors, including global SST distribution and land surface fluxes from within the region of West Africa. The following section considers the role of these two factors in the climatic variability over West Africa.

5. The role of SST and land cover type

The proposed relationship between entropy, circulation, and rainfall points to an important land–atmosphere–ocean interaction. As shown in Fig. 9, depending on the meridional gradient of boundary-layer entropy, a strong circulation may or may not develop. The entropy fluxes from the ocean surface and the land surface are important factors, though not the only ones, in determining the meridional distribution of boundary-layer entropy. Other processes that modify boundary-layer entropy include radiative cooling, entrainment at the top of the boundary layer, and convective downdrafts. The land–atmosphere–ocean interaction of Fig. 9 is fundamentally different from the land–atmosphere

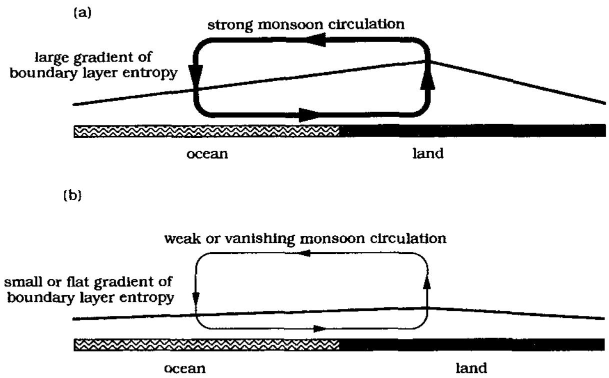

FIG. 9. A schematic of the proposed land–atmosphere–ocean interaction in West Africa.

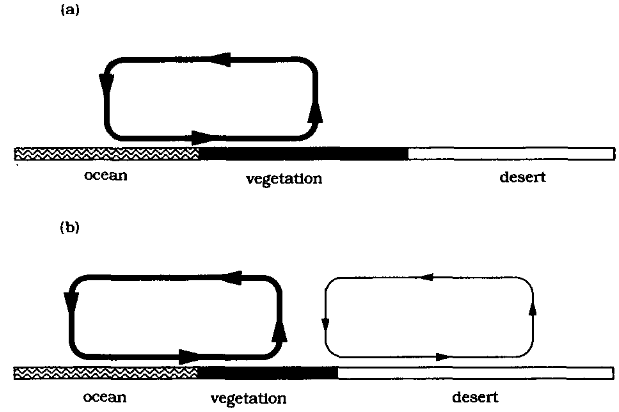

FIG. 10. A schematic of the land–atmosphere interaction, introduced by Charney (1975), to depict the response to a change in albedo near the desert border. (a) The circulation prior to the change in albedo and (b) the circulations after the change in albedo.

interaction introduced by Charney (1975), which is reviewed in section 2 and depicted schematically by Fig. 10. While the former describes the dynamics of the moist atmosphere across the land–ocean boundary, the latter emphasizes the dynamics of a dry atmosphere across the savanna–desert border. One important implication of the land–atmosphere–ocean interaction mechanism proposed in this paper is that variability in SST off the coast of West Africa and/or changes in land cover (affecting parameters such as albedo, roughness, root zone depth) in the coastal region of West Africa may play a more important role in the regional climate than do changes in land cover near the desert border.

The theory and observations presented in this paper point to some of the potential links between rainfall variability over West Africa and different regional and global factors. The meridional distribution of boundary-layer entropy south of the land–ocean border, which is located around 5°N latitude, is largely controlled by the SST over the southeastern tropical Atlantic (SETA). However, north of 5°N, the contribution to boundary-layer entropy due to land surface fluxes is significant. Hence, on one hand the interannual variability in SST distribution may influence rainfall variability over West Africa through its impact on the meridional gradient of boundary-layer entropy. On the other hand, large-scale changes in land cover over West Africa may modify local fluxes of entropy and, hence, change the meridional gradient of boundary-layer entropy. The role of SST and land cover in the variability of climate in West Africa is summarized in Fig. 11. SST and land cover control the two ends of the meridional distribution of boundary-layer entropy. The meridional gradient of entropy is a function of the prevailing conditions over the ocean and land surface. These conditions are characterized by SST and the type of land cover.

The relationship between SST in the tropical Atlantic and rainfall in West Africa is the focus of several studies. Lamb (1978) presents composites of SST distribution in the tropical Atlantic for wet and dry years in the Sahel region. The wet years were 1943, 1950, 1952, 1957, 1954, and 1955, which were the six wettest years from 1941 to 1972. The dry years were 1972, 1942, 1971, 1970, 1949, 1968, and 1941, which were the seven driest years during the same period. The distribution of SSTAs during the months of July, August, and September, relative to the long-term average for

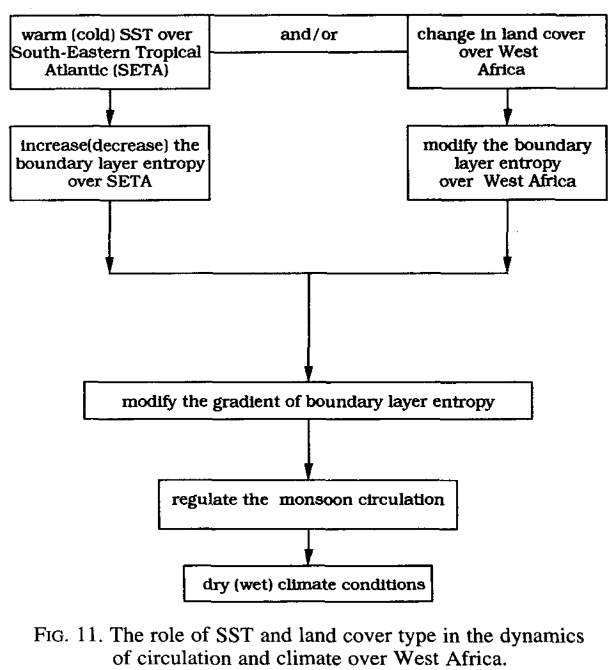

FIG. 11. The role of SST and land cover type in the dynamics of circulation and climate over West Africa.

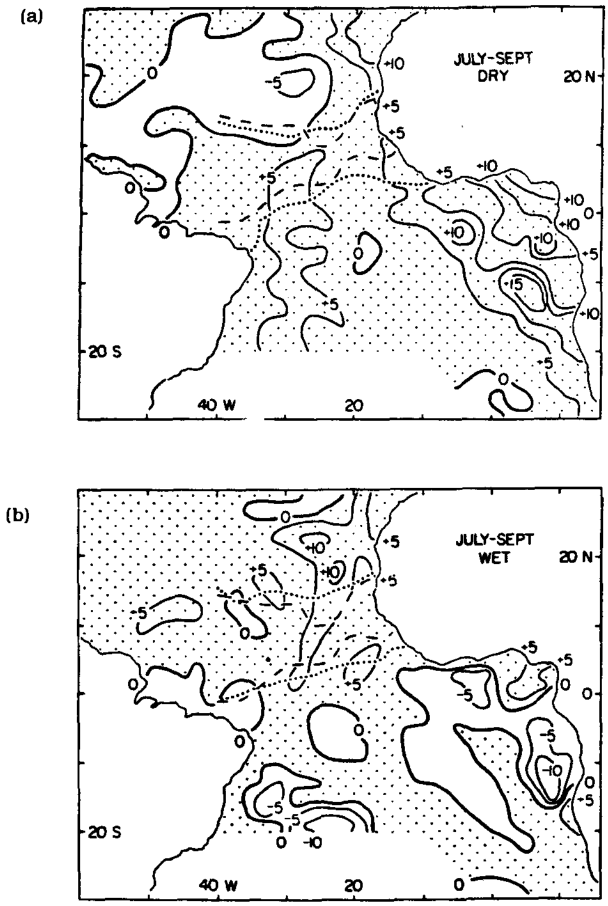

FIG. 12. Sea surface temperature anomalies in the tropical Atlantic. Units are 0.1°C. (a) Dry years in the Sahel region and (b) wet years in the Sahel region (Lamb 1978).

the period from 1911 to 1970, is shown in Fig. 12, which is taken from Lamb (1978). The distribution of SSTAs of the different years are averaged and presented separately for the wet and dry years. The observed SSTA in the region of SETA, which is defined as the area 20°S–5°N, 15°W–10°E, is significantly different during wet years than the observed SSTA in the same region during dry years. In general, warm (cold) SSTAs are associated with dry (wet) years. These findings are confirmed by further case studies that are reported by Lamb and Peppler (1992) and include observations for the years 1972 and 1984. The similar study of Lough (1986) investigates the relation of SST in the tropical Atlantic and rainfall in the Sahel region. The patterns of SSTA in the tropical Atlantic are identified using principal component analysis. For the period from 1948 to 1972, the second eigenvectors of SSTA are significant, and they correlate negatively with the Sahel rainfall. The second eigenvector of monthly and seasonal rainfall computed by Lough (1986) is quite similar to the pattern of SSTA distribution identified by Lamb (1978). Both patterns indicate anomalous SST conditions in SETA. Owen and Ward (1989) suggest that a measure of SST conditions in the region of SETA, given by the covariance between observed SSTA fields and the pattern of SSTA identified by Lamb (1978) and Lough (1986), can be used for forecasting rainfall in the Sahel region. The proposed measure explains about 36% of the variability in the observed Sahel rainfall during the period 1940–1987. The important conclusion to be drawn from all these empirical observations is that Sahel rainfall is significantly correlated to SST conditions in the region of SETA and that a warm (cold) SSTA in that region is associated with dry (wet) conditions in the Sahel region.

The theory and analysis presented in sections 3 and 4 may shed some light on the dynamics behind the observed relationship of SST in the region of SETA to the Sahel rainfall. Both theory and observations suggest that the dynamics of monsoons are regulated by the meridional gradient of boundary-layer entropy. The boundary-layer entropy over the ocean is a strong function of SST conditions. A warm (cold) SSTA would lead to large (small) fluxes of moist entropy and, hence, a large (small) magnitude of boundary-layer entropy. As illustrated schematically in Fig. 13a, a warm SSTA over the SETA would lead to an anomalously large magnitude of boundary-layer entropy over that region. Hence, every thing else being equal, a warm SSTA over SETA would favor a small gradient of boundary-layer entropy, a weak monsoon circulation, and dry conditions in the Sahel region. On the other hand, a cold SSTA is consistent with a large gradient of boundary-layer entropy, a strong monsoon circulation, and relatively wet conditions. Hence, the theory and analysis of this paper are consistent with the em-

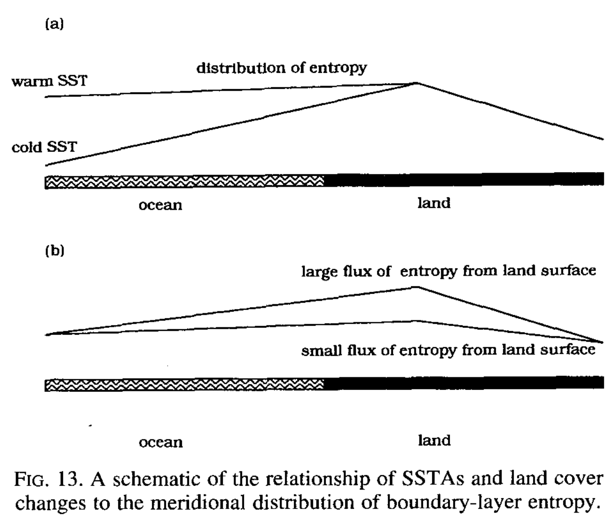

FIG. 13. A schematic of the relationship of SSTAs and land cover changes to the meridional distribution of boundary-layer entropy.

pirical observations of SST in SETA and rainfall in the Sahel.

The role of land cover over West Africa in regulating the variability of regional climate can be deduced by following a similar approach. The boundary-layer entropy over West Africa is partly supplied by local fluxes of sensible and latent heat. The energy balance and surface heat fluxes from West Africa are regulated by the fact that the prevailing type of land cover is forest, savanna, or short grass. As a result, land cover over West Africa is an important factor in determining the meridional gradient of boundary-layer entropy. Variability in surface fluxes from within West Africa is another key process in regulating the variability in the meridional gradient of entropy. As illustrated by Fig. 13b, the type of land cover that enhances the magnitude of the local boundary-layer entropy would enhance the monsoon circulation and favor wet conditions. Changes in land cover that modify the fluxes of entropy are likely to affect meridional distribution of entropy, leading to changes in the monsoon circulation and regional rainfall.

6. Conclusions

The theory and observations presented in this paper suggest that the dynamics of the monsoon circulation over West Africa are regulated by the meridional distribution of boundary-layer entropy. While a nearly flat distribution of boundary-layer entropy is associated with a rather weak circulation, a large gradient of entropy drives a strong monsoon circulation.

The dynamics of monsoon circulations over West Africa are relatively sensitive to natural and anthropogenic variability in the meridional gradient of boundary-layer entropy. This is particularly true if we compare West Africa to other tropical regions that experience monsoons, such as China or India. The reason behind this sensitivity lies in the relatively small magnitude of planetary vorticity and the relatively large magnitude of the gradient of planetary vorticity over West Africa. This sensitivity may explain some of the low-frequency variability observed in paleo climatic records from West Africa (Nicholson 1989).

The finding that the dynamics of wet and dry years in the Sahel region are related to the distribution of boundary-layer entropy has important implications regarding the relationship between SST distribution in the tropical Atlantic Ocean and rainfall in the Sahel. During the last few decades, several studies have confirmed that observed rainfall in the Sahel region is significantly correlated with certain patterns of SST distribution. In particular, wet (dry) years in the Sahel are associated with cold (warm) SST distribution in SETA. These empirical observations are consistent with the theory described in this paper. A cold SST anomaly in the SETA region results in relatively low boundary-layer entropy. Due to the location of SETA directly south of West Africa, a low boundary-layer entropy over SETA favors a large gradient of boundary-layer entropy. The theory described in this paper suggests that a large gradient of boundary-layer entropy would force a strong monsoon circulation, resulting in a relatively wet rainy season.

The proposed mechanism of land–atmosphere–ocean interactions provides a suitable framework for understanding the role of land surface processes in the regional climate of West Africa. Natural and/or anthropogenic variability in land cover may impact the large-scale circulation in this region through changes in surface fluxes and boundary-layer entropy over the land region.

The analyses presented in this paper focus on the observations in two extreme hydrologic years during the period from 1958 to 1967, inclusively. This is the only period for which we have atmospheric data prior to the onset of the drought. Some of our future research will be directed toward understanding the role of the meridional distribution of entropy in climate variability during the current drought. In addition, we plan to study the dynamics of the monsoon circulation over timescales that are shorter than the monthly timescale. Specific hypotheses regarding the causes of the drought will be developed and tested.

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