Large-Scale Atmospheric Dynamics and Sahelian Precipitation

KERRY H. COOK

Atmospheric Science Program, Cornell University, Ithaca, New York

(Manuscript received 14 June 1995, in final form 5 November 1996)

ABSTRACT

Observations show a broad band of precipitation across northern Africa, with maxima evident in some analyses on either side of the continent. A low-resolution GCM with simple boundary conditions produces such a band and, by producing a double-maximum structure, suggests the operation of distinct mechanisms for generating rainfall in the east and west. The precipitation, moisture convergence, and low-level wind convergence anomalies are very similar, indicating that an understanding of the low-level dynamics is essential for understanding the precipitation perturbation over the land surface. A linear model analysis shows that the anomalous low-level convergence is primarily forced by condensational heating in the middle and upper troposphere over East Africa. Low-level condensation and dry convection are also important for driving convergence in the west.

Understanding the response of the low-level flow is key for understanding how inhomogeneity at the surface is communicated into the precipitation field. Midtropospheric condensational heating stretches vortex columns and induces a positive vorticity tendency in the lower troposphere. To establish a climatology, the low-level dynamics must adjust to balance this tendency in a way that maintains moisture convergence. The balance is accomplished by the meridional advection of low absolute vorticity air from the south and frictional effects.

1. Introduction

The atmospheric dynamics at low levels serves as a connector between the precipitation field and surface conditions (on land and ocean) in the Tropics. Within 20° or 30° of the equator, depending on the season, strong condensational heating in the middle and upper troposphere (around 600–300 mb) is locally balanced by adiabatic cooling, with horizontal temperature advection of minor importance for balancing the heating. Midtropospheric vertical velocities associated with this adiabatic cooling drive horizontal wind convergence in the layers below about 800 mb by continuity. Since the lower layers are moist, atmospheric water vapor is converged and the upper-level heating is fueled and sustained.

This simple story has complications and subtleties that lead to various precipitation climatologies and variability characteristics. How the SST distribution maps into the tropical precipitation field is a topic of continued interest. In this paper, connections between land surface conditions and precipitation via the low-level atmospheric dynamics over northern Africa in summer are explored. Climate modeling is used to address two main questions. 1) Does the upper-level condensational heating force all of the low-level convergence, or can low-level diabatic heating also be important for determining the convergence field? 2) How is the response of the low-level dynamics to the forcing influenced by surface conditions, and what are the implications of that influence?

In order to understand the northern African precipitation climatology at the most fundamental level, it is important to investigate to what extent and how the rainfall distribution is determined simply by the latitudinal location and shape of the African continent. This understanding can then be used to develop a more complete picture of how the precipitation climatology is established and how its variability is generated. The influence of SST anomalies in space and time and the response to land surface conditions can then be considered in the context of a fundamental perturbation caused by the presence of the land surface.

Observations of the summertime precipitation climatology of northern Africa and results from modeling simulations are examined in the following section. Simulations with an atmospheric general circulation model (GCM) designed to investigate how this baseline perturbation comes about are described in section 3. Section 4 presents the diagnostic methods used to understand the GCM results. Section 5 contains a discussion of the role of diabatic heating, and in section 6 the dynamics of the low-level flow is analyzed. Section 7 has the conclusions.

2. The precipitation perturbation over Africa

Figure 1 shows four recently observed precipitation climatologies for Africa during boreal summer [June–
Fig. 1. June–August precipitation climatologies over Africa from different sources: (a) rain gauge compilation by Legates and Willmott (1990); (b) model reanalysis by the Goddard Earth Observing System group (Schubert et al. 1993); (c) satellite-gauge precipitation from Huffman et al. (1995); and (d) estimates from IR satellite observations from Huffman et al. (1997). Contour interval is 2 mm day$^{-1}$. 
August (JJA) average]. The ground-based observations of Legates and Willmott (1990) are shown in Fig. 1a, interpolated from the original 0.5° × 0.5° resolution to 2.25° latitude × 3.75° longitude. Figure 1b shows one of the recent reanalyses (Schubert et al. 1993), representing the 1985–93 mean precipitation. The third climatology (Fig. 1c) is the satellite-gauge precipitation product of Huffman et al. (1995); it is produced by using raingauge reports to adjust a multisatellite estimate. Precipitation rates shown in Fig. 1d are based on IR satellite observations (Huffman et al. 1997). The climatologies shown in Figs. 1c and 1d both represent the 1987–94 mean.

Despite the differences among the observed African precipitation climatologies, certain similarities emerge. Precipitation maxima in the east near the Ethiopian Plateau and on the west coast near 10°N are the most consistently reproduced features, although the locations and magnitudes vary. Each climatology adds one or two additional maxima, with three placing a rainfall maximum in the Nigeria–Cameroon region. The IR satellite observations (Fig. 1d) do not have a Nigerian maximum, but a precipitation enhancement near 20°E is shown. The model reanalysis from Goddard Earth Observing System (GEOS) (Fig. 1b) has four maxima, in agreement with the National Centers for Environmental Prediction (NCEP) reanalysis (Kalnay et al. 1996).

Figure 2 shows the JJA precipitation climatology from two GCM experiments, one with uniform land surface features and no topography (Fig. 2a; a 10-yr mean), and one with topography and realistic surface albedo and soil moisture distributions (Fig. 2b; a 12-yr mean). Both model integrations have prescribed SSTs based on observations (Shea et al. 1990) and R30 resolution (equivalent to about 2.25° lat × 3.75° long on the transform grid). Over the featureless version of Africa (Fig. 2a), two large-scale precipitation enhancements occur over the continent delineated by the, say, 10 mm day⁻¹ contour—one north of the Gulf of Guinea and one in the east. With a more realistic representation of the land surface (Fig. 2b), the basic large-scale pattern remains even though the positions of the maxima are shifted and the magnitudes modified. The East African maximum becomes focused on the high topography as in the observations. A comparison with Fig. 2a suggests that the East African precipitation maximum is modified by, but not caused by, the presence of topography. In West Africa, the presence of topography modifies the precipitation distribution as well. Semazzi and Sun (1995) also suggest that precipitation maxima over West Africa are not purely orographic.

When a square (in cylindrical equidistant projection), flat continent centered on the equator is inserted into the surface boundary conditions of an otherwise zonally uniform GCM, precipitation is always enhanced on the east coast of the continent in the summer hemisphere, even if the land surface is very dry (Cook 1994, hereafter C94). In contrast, summer hemisphere precipitation rates in the west increase slightly or decrease significantly by an amount that depends on surface conditions. This suggests that the precipitation perturbation in the east is physically different from that in the west.

The continental precipitation perturbation is closely related to the perturbation of the low-level flow in C94. Three mechanisms were identified for communicating between surface conditions and the precipitation field:

- warming at the surface generates increased dry convection, which increases (moisture) convergence over the land and enhanced precipitation;
- warming of the lower atmosphere causes reductions in relative humidity, decreased low-level condensation, and decreased (moisture) convergence and precipitation; and
enhanced midtropospheric condensation drives low-level (moisture) convergence.

The third forcing mechanism cannot be seen in a cause-and-effect context from the climatological perspective, but it is understood as a mutual adjustment among the low-level flow field, the moisture field, and the diabatic heating. This is similar to the framework of the Gill model (Gill 1980) in which anomalous (sea surface) temperatures are assumed to be associated with midtropospheric heating. It is this mechanism that drives low-level convergence and maintains strong precipitation over the eastern portion of the square continent. In the west, precipitation is largely determined by the net effect of the first two mechanisms.

Precipitation in East Africa is also distinguished from precipitation in West Africa by having different sources of precipitable water (Druyan and Koster 1989) and by having different variability signatures (Nicholson 1993).

The purpose of this paper is to understand how the large-scale structure of precipitation perturbations over Africa in boreal summer is determined and to investigate the dynamics of the low-level flow as it communicates structure at the surface into the precipitation field. The approach used involves the application of climate models to explore physical mechanisms; the goal is not to produce the best possible simulation of the observed precipitation field but to capture the large-scale pattern within a simple framework so that its basic cause is revealed.

3. GCM experiments

The climate model used here is an atmospheric GCM, a version of the model developed by the Climate Dynamics group at the National Oceanic and Atmospheric Administration’s Geophysical Fluid Dynamics Laboratory. The model is used in a somewhat simplified form. While one could argue that a GCM must be as comprehensive as possible, within the limits of our knowledge, the first priority here is to identify physical mechanisms. To generate results that can be cleanly diagnosed, simple boundary conditions are used. Since we are trying to understand how the presence of the African continent modifies the precipitation field in summer, all other sources of structure are eliminated. Sea surface temperatures, cloud amounts, and ozone concentrations are fixed and zonally uniform; sea ice is not allowed to form.

The GCM is run in “perpetual season” mode, with July conditions given by the imposed insolation and sea surface temperatures. These SSTs are zonally uniform, with a maximum value of 301 K at about 7°N. A low-resolution version of the model is used, with 15 spectral waves retained in what is equivalent to a grid resolution of about 4.5° lat × 7.5° long. Each integration begins with an isothermal atmosphere at rest; 200 days of the integration are discarded as the model finds its own quasi-equilibrium. Results presented here as GCM climatologies are averages over 1800 days of integration. As a rough measure to indicate that this integration length is sufficient, note that even the precipitation climatologies formed from each half of the full integration (i.e., the 900-day means) are nearly identical to each other and to the 1800-day mean.

The only continent included at the surface is Africa, so attention is restricted to the effects of the African land surface itself on African precipitation rates; the potentially significant presence of the Asian land mass is eliminated. In addition, the model’s representation of Africa is very simple, with no topography, 0.1 surface albedo everywhere (the same as the ocean), and uniform surface moisture that represents savanna conditions (5 cm of soil moisture is prescribed). Thus, land and ocean surfaces differ in three ways in this model: 1) the land surface can warm or cool in response to the surface heat budget unlike the prescribed-temperature ocean; 2) while evaporation proceeds at the potential rate over the ocean, drier conditions prescribed on land mean that evaporation is 0.44 times the potential evaporation rate; and 3) the land surface is “rounder” than the ocean surface, with vertical fluxes of momentum, sensible heat, and moisture enhanced to represent a more well-developed turbulent boundary layer (see below).

Figure 3a shows the precipitation perturbation, defined as the deviation from the zonal mean, from the low-resolution GCM with these boundary conditions. The background of the figure shows the continental outlines as resolved by the GCM. Two distinct precipitation maxima are generated over the land surface. One is located in East Africa over Ethiopia and Sudan. The second is in West Africa over parts of Nigeria, Niger, southern Mali, and the countries of the Guinean coast.

As in the square continent case studied by C94, the perturbation of the precipitation field is closely associated with the perturbation of the low-level flow over the warm land surface of the summer hemisphere. Figure 3b shows the low-level wind convergence anomaly from the GCM climatology. Here, the anomalous convergence is calculated as the deviation from the zonal mean and vertically integrated with mass weighting to account for the different thicknesses of the model layers. The vertical integral includes levels at 997, 979, 935, 866, and 770 mb since the vast majority of the water vapor converges within these layers. The structure of the wind convergence anomaly is very similar to that of the precipitation perturbation (Fig. 3a) and also the anomalous moisture convergence field (not shown). Convergence maxima are clearly distinguished in East and West Africa, at the same locations, and with the same relative strengths as the precipitation perturbations. Thus, understanding the low-level dynamics is essential for understanding the precipitation distribution.

The 935-mb winds (full fields) over Africa are also shown in Fig. 3b. Westerly flow dominates under both
precipitation enhancements, and there is enhanced cross-equatorial flow from the winter hemisphere at all continental longitudes.

4. Diagnostic methods

Two analysis methods are used to understand the response in the GCM. One is the application of a dry, steady-state, linearized model of the GCM’s dynamics (as used in C94). This linear model has the same 14 levels and the same horizontal resolution as the GCM, and is linearized about the zonal mean basic state from the GCM (Ting and Held 1990). Anomalies in the wind and temperature fields are solved for at each level as forced by diabatic heating and transient eddies, which are also taken from the GCM climatology. When structure in the precipitation field in the GCM is closely tied to the perturbation of the low-level wind, and when the linear model can reproduce the GCM’s low-level response, then the linear model is a valid tool for understanding connections between the precipitation field and the surface boundary conditions via the low-level flow. If this is the case, then the response can be decomposed to reveal the role of individual forcing functions (radiative, condensational, and sensible heating and the net effects of transient eddy heat transport) in causing low-level convergence.

The dynamics of the low-level flow is investigated to understand how the low-level flow adjusts to the forcing in a way that maintains the time mean convergence. From the primitive horizontal momentum equations on pressure surface one can easily derive

\[ \frac{\partial \zeta}{\partial t} + \mathbf{v} \cdot \nabla \zeta + \omega \frac{\partial \zeta}{\partial p} = - (\zeta + f) \nabla \cdot \mathbf{v} - \beta v + k \left( \frac{\partial v}{\partial p} \times \nabla \omega \right) + k (\nabla \times \mathbf{F}), \]

where \( \zeta \) is horizontal relative vorticity, \( \mathbf{v} \) is the horizontal (pressure surface) wind, and \( \omega \) is vertical \( p \)-velocity; \( f \) is the Coriolis parameter and \( \beta \) its meridional derivative. Also, \( \mathbf{F} = (F^x, F^y) \) is the horizontal acceleration due to the vertical transport of horizontal momentum.

Analysis of Eq. (1) can be written in terms of its climatological mean, denoted by an overbar, and a temporal perturbation, denoted by a prime. Taking the time mean and rearranging gives

\[ \frac{\partial \bar{\zeta}}{\partial t} = -(\bar{\zeta} + f) \nabla \cdot \bar{\mathbf{v}} - \bar{\mathbf{v}} \cdot \nabla (\bar{\zeta} + f) \]

\[ + \left( \frac{1}{\alpha \cos \phi} \frac{\partial F^x}{\partial \lambda} - \frac{1}{\alpha} \frac{\partial F^y}{\partial \phi} \right) \]

\[ - \nabla^2 \bar{\zeta} - \bar{\zeta} \nabla \cdot \mathbf{v} = 0. \]

The vertical advection of relative vorticity and the tilting term have been neglected; the analysis of the GCM output discussed below justifies this approximation.

Analysis of Eq. (2) is used to understand how the low-level flow adjusts to diabatic heating. The first term on the rhs is the vorticity tendency associated with the convergence forced by the diabatic heating (when structure in the relative vorticity field is relatively unimpor-
The linear model analysis (section 4) shows how this convergence is forced. To establish a quasi-equilibrium climate state, the atmosphere must move in such a way that the sum of the remaining terms on the rhs of Eq. (2) are equal and opposite to the convergence term. Understanding how this balance is achieved provides insight into how the precipitation is maintained in the time mean, including the role of and sensitivity to the boundary conditions.

The second term on the rhs of Eq. (2) is the vorticity tendency due to the horizontal advection vorticity, and the third term is the vorticity tendency due to friction. The last two terms represent the effects of transients, that is, the net effects of correlations between wind and vorticity perturbations on the time-mean relative vorticity field. Each of the transients terms is calculated for the GCM climatology from

$$\mathbf{\nabla} \cdot \mathbf{\nabla} \mathbf{v} = (\mathbf{\nabla} + f) \mathbf{\nabla} \mathbf{u} - (\mathbf{\nabla} + f) \mathbf{\nabla} \mathbf{\xi}$$

(3)

and

$$\mathbf{\nabla} \cdot \mathbf{\nabla} \mathbf{\xi} = -\mathbf{\nabla} \cdot \mathbf{\nabla} \mathbf{u} - \mathbf{\nabla} \cdot \mathbf{\nabla} \mathbf{\xi}.$$

(4)

In the GCM, $F^s$ and $F^f$ are related to vertical structure in the stresses, $\tau$, according to

$$\frac{\partial \mathbf{u}}{\partial \mathbf{t}} \sim F^s = -\frac{g}{\rho} \frac{\partial \tau^s}{\partial \mathbf{\alpha}}$$

and

$$\frac{\partial \mathbf{v}}{\partial \mathbf{t}} \sim F^f = -\frac{g}{\rho} \frac{\partial \tau^f}{\partial \mathbf{\alpha}}.$$

(5)

At the surface,

$$\tau^s = -\rho_s \mu C_D u_{14}$$

and

$$\tau^f = -\rho_s C_D v_{14}.$$

(6)

where $\rho$ is the density of air and $V$ is wind speed. The subscript 14 refers to the lowest model layer ($\sigma = 0.996$) and the subscript $s$ refers to the surface; $C_D$ is the bulk aerodynamic drag coefficient, which is a measure of the surface roughness; $C_D$ is set equal to 0.001 over ocean surfaces and 0.003 over land. The momentum flux from (or to) the surface modifies the flow when Eq. (6) is used in applying Eq. (5) at the lowest model level. The momentum flux from the surface is diffused through the lowest five layers of the model, up to $\sigma = 0.770$. At each level,

$$\tau^s = -\rho g \frac{\partial \mathbf{u}}{\partial \mathbf{\alpha}}$$

and

$$\tau^f = -\rho g \frac{\partial \mathbf{v}}{\partial \mathbf{\alpha}}.$$

(7)

where $K_v$ is a diffusion coefficient that depends on the wind shear.

This standard and relatively simple treatment has been reviewed so that it is clear where structure in the $F^s$ and $F^f$ terms can originate in the model. Contributions to the horizontal derivatives of $F^s$ and $F^f$ in Eq. (2) can arise from structure in the wind and density fields as well as variations in $C_D$, associated with the land/sea distribution.

5. Forcing of moisture convergence by diabatic heating

The linear model diagnosis is used to understand the relationship between low-level convergence and individual diabatic heating processes. Figure 4a shows the wind convergence anomaly from the linear model, vertically mass integrated over the model layers in which nearly all of the moisture convergence occurs in the GCM (as for Fig. 3b). To generate this solution, the model was linearized about the zonal mean wind and temperature fields from the GCM climatology. Anomalies were forced by the combined effects of dry and moist convection, radiative heating, sensible heating, and thermal transients. The structure of the linear solution with this full forcing is quite similar to the anomalous convergence field generated in the GCM (Fig. 3b): the magnitude is somewhat larger, but it is reduced by about 20% when the effect of heat transport by transient eddies is taken into account. The good agreement between the GCM output and the linear model solution indicates that the low-level dynamics is essentially linear, and that the linear framework is useful for distinguishing among the effects of various forcing functions.

In the middle and upper troposphere, the presence of the land surface below is felt through perturbations of the condensational heating field. Figure 5a shows a cross section of the condensational heating field from 770 to 370 mb in the vicinity of Africa at about 11°N. While radiative cooling is somewhat significant, with values of about $-1.2 \times 10^{-5}$ K s$^{-1}$, it is relatively uniform over the region; the other diabatic heating components are quite small.

When the condensational heating field above 770 mb (Fig. 5a) is the only forcing in the linear model, much of the low-level convergence is induced. Figure 4b shows the linear solution for this case. The similarity to the full-forcing case (Fig. 4a) is striking, with convergence forced over East and West Africa and anomalous divergence just south of the equator and in the northwest. In the east, almost all of the convergence from the full forcing case (Fig. 4a) is captured in Fig. 4b. This dominance of upper-level condensational heating in forcing low-level convergence in the east is expected based on C94. In the west, the result is different from the square continent case in which low-level heating dominates the forcing. With the Gulf of Guinea present, the cycle in which precipitation and the attendant condensational heating maintain low-level (moisture) convergence can become established. However, in contrast to East Africa, only a little more than half of the convergence in the west is forced in this way and the divergence to the south is also weaker than the full-forcing case. As seen below, the rest of the structure in the west is due to low-level heating.

Figure 5b shows the sum of the heating due to condensation, dry convection, and sensible heating (i.e., the vertical diffusion of heat) below 770 mb. Unlike the
upper-level heating, the diabatic heating at low levels is a little stronger in the west than in the east. Heating due to dry convection is responsible for the double maximum structure; the condensational and sensible heating fields are fairly uniform across the continent.

Results from the linear model show that this low-level heating drives low-level convergence over West Africa. Figure 4c shows the low-level convergence forced in the linear model by the low-level heating field of Fig. 5b. While there is some effect in the east, the convergence is larger in the west and makes up the difference between the full forcing (Fig. 4a) and upper-level heating (Fig. 4b) cases. The north–south dipole structure in Fig. 4c indicates that the low-level heating contributes to forcing the cross-equatorial diversion of the winter hemisphere tropical easterlies over the western half of Africa. This sensitivity to low-level diabatic heating suggests that West Africa has more ways to be sensitive to land surface temperature and, therefore, land surface conditions, than East Africa. [This is also supported by the results of C94 and Cook (1994b).]

The vertical structure of the response to full diabatic heating is more complicated over West Africa than in the east because of the importance of both low-level and upper-level heating in the west. In the west, convergence through the 997-, 979-, and 935-mb layers is diminished by divergence at 866 and 770 mb. In the east, convergence occurs at all levels included in the vertical integral.

The linear model analysis highlights an important difference between West and East Africa, suggesting ways in which variability in these two regions may be different. The analysis also points to a similarity in the two regions, namely, that the forcing of low-level convergence by midtropospheric heating is important in both regions. The heating stretches vortex tubes in the atmosphere below, inducing low-level wind convergence and a positive vorticity tendency [Eq. (2)]. To establish a stable climatology in the presence of this heating, the low-level dynamics has to respond to the forcing in a way that 1) converges enough moisture to supply the precipitation field and 2) balances the forced positive vorticity tendency. The final precipitation climatology results from the mutual adjustment of the heating field and the low-level dynamics.

In the next section, the low-level dynamics is investigated to understand how the vorticity budget is established in the presence of heating aloft. As was the case for the linear model analysis, there are differences in

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**Fig. 4.** Anomalous wind convergence in the linear model mass integrated over the lowest model levels (excluding the 997-mb level) forced by (a) full forcing (see text), (b) condensational heating at and above 680 mb, and (c) diabatic heating at and below 770 mb. Contour interval is $10^{-4} \text{s}^{-1}$ in (a) and $0.5 \times 10^{-4} \text{s}^{-1}$ in (b) and (c). Divergence is indicated by dashed contours.
6. Dynamics of the low-level response

Analysis of the GCM climatology confirms neglecting the vertical advection of relative vorticity and the twisting term in simplifying Eq. (1) into Eq. (2). In addition, the sum of the vorticity transients terms is small; although each term is significant, they tend to oppose each other ($\nabla \cdot \nabla \phi = 0$). Thus, two terms are more important for opposing the externally forced positive vorticity tendency, namely, the horizontal advection and friction terms.

Figure 6a shows the convergence term calculated from the GCM output, mass integrated over the 997-, 979-, 935-, 866-, and 770-mb levels; this is interpreted as the vorticity tendency due to diabatic heating. The structure of the precipitation and convergence perturbations (Figs. 3a and 3b, respectively) is reflected in this term since the relative vorticity perturbation is not introducing much structure.

Opposing the positive vorticity tendency in the regions of forced low-level convergence are the vorticity advection term (Fig. 6b) and the friction term (Fig. 6c). The advection term is largest on the equatorward side of the precipitation/low-level convergence/convergence term maxima; the friction term (Fig. 6c) is important on the poleward side. Both the advection and friction terms exhibit longitudinal structure over the land surface, but it is more pronounced in the advection term.

a. Role of vorticity advection

Magnitudes of the advection term are comparable to those of the convergence term, but the strongest vorticity advection occurs to the south of the convergence maxima (Fig. 6b). In West Africa, the local maximum is directly over the Guinean coast about 8° south of the precipitation maximum. In East Africa, vorticity advection is strongest at 5°N and 30°E, near where two of the observed climatologies shown (Figs. 1b and 1d) place a precipitation maximum. The positive vorticity tendency due to advection in the northwest, in the vicinity of the west coast countries of Senegal and Mauritania, offsets the negative tendency associated with decreased convergence and precipitation (Fig. 3).
The vorticity tendency due to climatological vorticity advection is

$$\frac{\partial \zeta}{\partial t}_{\text{advection}} = -\mathbf{v} \cdot \nabla (\zeta + f)$$

$$= -\mathbf{u} \frac{\partial \zeta}{\partial x} - \mathbf{v} \frac{\partial \zeta}{\partial y} - \mathbf{v} \beta,$$  \hspace{1cm} (8)

where $\beta = \frac{df}{dy}$. In the GCM climatology, zonally elongated relative vorticity maxima are associated with the regions of strong convergence (Fig. 7). Maximum relative vorticity values are about $1 \times 10^{-5}$ s$^{-1}$ at 15°–20°N; the planetary vorticity at 15°N is $4 \times 10^{-5}$ s$^{-1}$, so setting $\zeta + f = f$ is only a rough approximation near the precipitation maxima. The zonal elongation of the relative vorticity maximum results in small zonal relative vorticity gradients; that, combined with the fact that the westerly flow onto the continent is weak (Fig. 3b), results in magnitudes of the zonal advection term in Eq. (8) that are an order of magnitude smaller than the convergence term. Druyan and Koster (1989), in a GCM analysis, cite the Atlantic Ocean due west of Africa as an important source of precipitating water for the western Sahel during rainy years. This suggests that the two GCMs are behaving differently, or that rainy years are rainy because the zonal advection of moisture is more effective.

Meridional advection of relative and planetary vorticity are both important for balancing the vorticity tendency forced through the convergence term. Figure 8a shows the meridional advection of relative vorticity mass integrated over the lowest five levels of the model; Fig. 8b is the planetary vorticity advection. (A superposition of Figs. 8a and 8b essentially yields Fig. 6b.) Large meridional relative vorticity gradients (Fig. 7) and strong northward flow across the Horn of Africa and the Guinean coast (Fig. 3b) combine to create significant meridional advection of relative vorticity. The mechanism is only important over the land surface. It is especially pronounced over West Africa and, since it is essentially a nonlinear mechanism involving products of perturbation winds, it is possibly the reason that the linear model fails to capture the full magnitude of the low-level convergence over West Africa (Fig. 4b).

While the meridional advection of relative vorticity has a north/south dipole structure centered on the equator, the meridional advection of planetary vorticity is largest on the equator where $\mathbf{v}$ and $\beta$ are both greatest. It is important over the oceans as well as over the land surface, where it doubles in magnitude in east and west

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**Fig. 6.** Three terms in the vorticity budget [Eq. (2)] calculated from the GCM climatology and mass integrated over the lowest five model levels: (a) convergence term, (b) total vorticity advection, and (c) friction term. Contour intervals are $10^{-11}$ s$^{-2}$. 
equatorial Africa because of the high perturbation meridional velocity.

Assuming for the moment that the positive vorticity tendency associated with the forced convergence is balanced totally by the advection of planetary vorticity (the Sverdrup balance), and letting $\frac{\zeta + f - f}{2} = \dot{\zeta}$ (see Fig. 7), we can estimate the meridional velocity required for such a balance. Setting the convergence term equal to the planetary advection term in the vorticity tendency equation gives

$$\dot{\zeta} = a (\tan \phi) (-\nabla \cdot \vec{V}), \quad (9)$$

where $a$ is the radius of the earth. With $-\nabla \cdot \vec{V} = 5 \times 10^{-5} \text{s}^{-1}$ from the GCM climatology, and $\phi = 7^\circ$ latitude, then $\dot{\zeta} = 4 \text{ m s}^{-1}$. This value for the meridional velocity is similar to the values observed and simulated (Fig. 3). Farther north, it would be more difficult to maintain this mechanism since $\tan \phi$ increases rapidly away from the equator. For example, even at 15°N northward velocities of about 9 m s$^{-1}$ are required to balance the same amount of convergence by the meridional advection of planetary vorticity. The importance of this mechanism indicates that the position in latitude of the Guinean coast is an important factor determining the precipitation climatology of West Africa and indicates why the advection term is most important for balancing the externally induced vorticity tendency on the equatorward side of the convergence maxima.

Planetary vorticity advection also influences how far north the continental precipitation reaches, especially in the west. Negative precipitation anomalies occur in northwest Africa, especially on the coast between 10° and 20°N (Fig. 3a). This decreased precipitation is associated with low-level divergence and equatorward flow (Fig. 3b) that is forced mainly by the upper-level heating (Fig. 4b). The attendant advection of high plan-
etary vorticity air provides the positive vorticity tendency to balance the negative tendency associated with anomalous divergence. The effect is dampened somewhat by the negative vorticity tendency associated with relative vorticity advection since the air flowing southward over northeastern Africa flows up the vorticity gradient (Fig. 7). Precipitation and low-level convergence reaches farther north over East Africa where the southward flow is weaker.

The essence of the connection between the low-level wind convergence and the precipitation is that the converging winds must carry sufficient moisture to feed into the upper levels. If the low-level convergence (positive vorticity tendency) was suddenly being balanced by the meridional advection of dry, low-planetary-vorticity air, the climatological precipitation maximum and, therefore, the forced low-level convergence would not be maintained. This is the case of the square continent discussed by C94. However, structure in the moisture field is not of primary importance in determining the distribution of the moisture convergence field.

Structure in the evaporation field is also not closely related to the precipitation perturbation, as is generally the case over ocean surfaces as well. Evaporation rates are comparable to precipitation rates, with values of about 4 mm day\(^{-1}\) over a broad area of equatorial Africa but they do not reflect the structure in the precipitation field. Remember that the GCM experiment is designed to highlight the role of the low-level dynamics, so the soil wetness is uniform on the land surface and doesn’t respond to the precipitation or evaporation fields. Thus, it is likely that structure in the low-level atmospheric water vapor and evaporation fields is underestimated here. However, even in GCM experiments with more realistic soil moisture distributions (e.g., Lenters and Cook 1995), almost all of the structure in the precipitation anomaly is reflected in the convergence and not in the evaporation field. Thus, we have

\[
P^* = E^* - \int_0^{\sigma=10} (\nabla \cdot \mathbf{q}) \frac{d\sigma}{g \rho_w} \nonumber \]

\[\cong - \int_{\sigma=14}^{\sigma=10} (\nabla \cdot \mathbf{q}) \frac{d\sigma}{g \rho_w} \]

\[\cong \int_{\sigma=14}^{\sigma=10} \frac{\nabla \cdot \mathbf{q}}{\sigma} \frac{d\sigma}{g \rho_w}, \tag{10}\]

where the asterisk indicates a deviation from the zonal mean and \(q\) is the atmospheric water vapor mixing ratio.

\(b.\) Role of friction

Equation (2) indicates that zonal (meridional) gradients in the vertical flux of \(v\) (\(u\)) momentum can generate vorticity. The magnitude of the friction term (Fig. 6c) is not as large as the vorticity advection term (Fig. 6b), but friction is playing a significant role across the northern half of the continental precipitation enhancement. Thus, understanding the role of friction is important for understanding how far north the ITCZ reaches over the African continent.

The vorticity tendency due to friction is almost exclusively associated with meridional structure in the \(u\)-momentum flux in the GCM; that is,

\[
\frac{\partial \xi}{\partial t}_{\text{friction}} \approx -\frac{1}{a} \frac{\partial F^s}{\partial \phi} = \frac{1}{a} \frac{\partial}{\partial \phi} \left( g \frac{\partial \tau^s}{\partial p} \right), \tag{11}\]

where Eq. (5) has been used. In the lowest model layer,

\[
\frac{\partial \xi}{\partial t}_{\text{friction}} \approx \frac{g}{a \rho_p} \frac{\partial}{\partial \phi} \left( \tau_x - \tau_{13.5} \right), \tag{12}\]

where the subscript \(s\) refers to the surface and the subscript 13.5 refers to the half-\(\sigma\) level above the lowest model layer. In deriving Eq. (12), meridional variations in \(p\) have been neglected. Using Eq. (6),

\[
\frac{\partial \xi}{\partial t}_{\text{friction}} \propto C_D \frac{\partial (V_{14}u_{14})}{\partial \phi} + V_{14}u_{14} \frac{\partial C_D}{\partial \phi}. \tag{13}\]

According to Eq. (13), the vorticity tendency due to friction can be associated with meridional gradients in the surface roughness, as represented by the bulk aerodynamic drag coefficient, and with meridional gradients in the low-level wind. The \(u\)-momentum flux is largest near the surface; it is diffused vertically through the layers in which most of the moisture converges [Eq. (7)].

The negative vorticity tendency due to friction over Sahelian Africa (Fig. 6c) is primarily due to meridional variation in the zonal wind [first term on the rhs of Eq. (13)]. Figure 9a shows the \(u\)-momentum flux, \(F^u\), at 979 mb; the distribution is similar at each of the five lowest model levels over which momentum is diffused. Away from the land surface, there are positive values throughout the Tropics, indicating a \(u\)-momentum flux from the surface to the atmosphere under easterly flow (Fig. 9b). Negative values in the middle latitudes indicate the downward momentum flux that develops in the presence of low-level westerlies. The magnitudes of the \(u\)-momentum flux are proportional to the low-level zonal wind, which is shown in Fig. 9b.

Over northern Africa between 5\(^\circ\) and 20\(^\circ\)N, negative values of \(F^u\) occur, with magnitudes rivaling those of the summer hemisphere midlatitudes. The westerlies induced over Africa at the same latitudes are associated with the transport of \(u\)-momentum from the atmosphere to the surface (\(F^u < 0\)) and a deceleration of the zonal wind, indicated schematically by the westward-pointing arrow in Fig. 9b. Farther north, easterlies at the surface over western Africa at about 25\(^\circ\)N (the “Harmattan winds” that carry Saharan dust out over the Atlantic) are associated with upward (from the surface to the atmosphere) momentum transport (\(F^u > 0\) and an ac-
acceleration of the zonal wind, indicated by the eastward arrow. The result is a positive meridional gradient of $F^a$ that is associated with a negative vorticity tendency due to the stress differential acting on parcels. As indicated by the arrows on Fig. 9b, this results in an anticyclonic perturbation vorticity.

The effects of surface roughness and stress differentials on the low-level convergence in the Tropics has been discussed in the literature as a coastal effect (Bryson and Kuhn 1961; Hastenrath 1991). In these GCM experiments, the only region where the second term on the rhs of Eq. (13) is nonnegligible is near $15^\circ$S. Here, the strong winter hemisphere trades sense the jump in $C_D$ at the land–sea interface, but only in the two lowest levels of the model ($\sigma = 0.997, 0.979$). The response is reflected in the convergence field at these levels (not shown) but not in the precipitation field (Fig. 3a). [A weak coastal response also occurs on the African west coast (Mauritania), but it is even more highly confined in the vertical.]

This is not to say that the fact that the land surface is "rougher" than the ocean surface is not playing a role in determining the convergence and precipitation field in northern Africa. The increased roughness enhances the vertical $u$-momentum flux [first term, rhs of Eq. (13)]. This causes the mechanism described above to be a more effective way of balancing the vorticity tendency forced by the condensational heating aloft, and so stronger precipitation (i.e., greater stretching of the vortex columns in the lower troposphere) can be supported climatologically.

A GCM experiment quantifies this idea. In this integration, the GCM is given exactly the same boundary conditions as those described in section 3 except the drag coefficient $C_D$ is set equal to its ocean value (0.001) over land as well (called the "smooth" continent here, as opposed to the original "rough" continent). The precipitation perturbation over Africa from that experiment is shown in Fig. 10a; Fig. 10b shows the difference in the full precipitation fields, for the rough ($C_D = 0.003$) continent minus the smooth ($C_D = 0.001$) continent. Comparison with Fig. 3 indicates that both continental precipitation maxima are 25%–30% weaker in the smooth continent case. Rainfall is up to 1 mm day$^{-1}$ less near $10^\circ$N, $15^\circ$E (northern Cameroon/western Nigeria) over the rougher continental surface since the West African precipitation maximum is located farther west over the rough continent. The latitude of the precipitation enhancement across the continent is the same, as is the northern cutoff and the negative perturbation on the equator. However, the longitude of the West Africa precipitation maximum over the smooth is about $10^\circ$ of longitude to the east of the West African precipitation maximum for the rough case.

These differences in precipitation are reflected in the low-level flow field since the low-level convergence anomaly has the same structure as the precipitation perturbation. Figure 11a shows the 935-mb wind vectors over the smooth (thinner arrows) and rough (thicker arrows) continents. Also shown is the surface pressure from the smooth continent experiment; the surface pressure field over the rough continent is nearly identical. The wind velocities over the rough continent are generally smaller than those over the smooth continent, but they blow more directly down the pressure gradient (represented by the surface field here). In the three-way
balance of the Coriolis, friction, and pressure gradient forces, a strengthening of friction decelerates the wind, weakens the Coriolis force, and leads to flow more directly into low pressure.

The implications for the low-level convergence field are seen in the wind vector differences for the rough minus smooth case in Fig. 11b. The convergence field of these difference vectors is shown by the contours. In the continental interior, near 10°N, 15°E, where the precipitation is smaller over the rough continent (Fig. 10b), the turning of the winds into the pressure gradient causes stronger anomalous wind divergence and precipitation reductions. North of the Gulf of Guinea and in East Africa, anomalous convergence is associated with the strengthening and relocation of the precipitation maxima. It is this modification of the flow that leads to the simulated dependence of the low-level convergence and precipitation on surface roughness.

In the diffusion of $u$-momentum through the lower layers of the model, Eq. (7) indicates that a relatively strong contribution to the vorticity tendency is expected where the zonal wind shear is strong. At 935 mb, wind magnitudes are very similar over the rough and smooth continents, and the contribution of the frictional mech-
anism becomes weak above that level. Below 935 mb, however, the zonal wind speed over the rough continent is smaller than over the smooth continent (not shown), giving a stronger vertical wind shear and a stronger contribution to the vorticity tendency from the friction term.

The above discussion identifies two ways in which the low-level flow adjusts to the convergence forced by heating aloft to maintain a stable climatology, namely, through vorticity advection and frictional effects. The relative importance of these ways of balancing the vorticity budget varies with height in the atmosphere, and there are differences between East and West Africa. Figure 12a shows a vertical cross section of the convergence term [Eq. (2)] across the continental precipitation maximum at 15°N. The West African maximum at about 5°E is somewhat lower than the East African maximum near 35°E, with maximum convergence at about 965 mb in the west and 950 mb in the east (maybe not a significant difference given the model’s vertical resolution).

Figure 12b shows a vertical section of the advection term, and Fig. 12c is a vertical section of the friction term. Each is plotted at the latitude where it is largest, near 8° and 20°N, respectively. As shown in Fig. 4, the vorticity advection term is larger than the friction term in the vertical integral. However, the vertical cross section indicates the important role that the friction term plays very near the surface and emphasizes that this term is important for understanding how changes in land surface conditions affect precipitation rates.

7. Conclusions

Observational climatologies for northern Africa in summer show a broad band of precipitation that is enhanced in the east and in the west in some datasets. A low-resolution GCM reproduces strong precipitation across northern Africa. Two distinct maxima develop that suggest separate mechanisms sustain continental precipitation over East and West Africa. The presence of surface features, such as topography and surface albedo variations, cause modifications of, but are not fundamental determinants of, the large-scale precipitation pattern. This paper uses GCM and linear model analyses to understand how the fundamental perturbation comes about due to continentality.

Simple boundary conditions are used in the GCM so that the first-order precipitation perturbation can be isolated and connections between structure at the surface and structure in the precipitation field can be clearly analyzed. The model uses a flat version of the African continent, with uniform and unchanging surface albedo and soil wetness. Clouds are also prescribed, and there is no longitudinal structure in the prescribed SSTs. Thus, the double-maximum structure in the precipitation field is most fundamentally determined by the location of the African continent in latitude and its shape.
The low-level convergence anomaly mimics the structure in the precipitation field, indicating that the perturbation of the low-level flow over the land surface is instrumental for determining the precipitation anomaly. The application of a linear model shows how the low-level convergence anomalies are forced. In East Africa, and in agreement with the results of C94 for the square continent case, low-level convergence is forced almost entirely by midtropospheric condensational heating. This mechanism is also important over West Africa and accounts for a little more than half of the convergence. The rest is forced by low-level (below about 700 mb) heating due to condensation, dry convection, and sensible heating. Since surface conditions can directly influence these low-level heat fluxes, the West African region is potentially more complex in terms of its sensitivity to the land surface than East Africa.

An examination of the vorticity budget is used to further connect the land surface with the precipitation field through an understanding of the low-level dynamical response to both midtropospheric condensational heating and the surface. Middle-tropospheric heating associated with the continental precipitation maxima stretches vortex columns in the troposphere below, inducing a positive vorticity tendency. For the precipitation maxima to become established as a feature of the time-mean climate, the lower atmosphere must move in such a way that negative vorticity tendencies are generated to balance. In addition, this atmospheric motion must carry and converge sufficient moisture to maintain the condensation field.

One way that the forced positive vorticity tendency in the lower atmosphere is balanced is by the advection of low absolute vorticity air. The advection of planetary and relative vorticity are both important, especially on the equatorial half of the precipitation features. In the former, which is essentially the Sverdrup balance, the stretched vortex columns move poleward to regions with higher Coriolis parameter \( f \) to conserve potential vorticity. The contribution from the advection of relative vorticity is especially important in West Africa. This is a nonlinear effect and occurs as the diverted wintertime easterlies flow up the relative vorticity gradient.

Absolute vorticity advection also plays a role in determining how far north over Africa precipitation is maintained, particularly in the west. As air flows into the region of convergence from the north, it is also flowing up the relative vorticity gradient and generates negative vorticity tendency. However, the flow is also advecting higher planetary vorticity air from the north. The planetary vorticity advection term dominates, resulting in a positive vorticity tendency that is strongest in the northeast where the northerly flow is most pronounced.

Surface stress differentials are also important in balancing diabatically forced positive vorticity tendencies under the precipitation maxima. The effect is greatest below 900 mb and north of the strongest negative vorticity advection. Perturbations in the low-level wind over the land surface are responsible for generating the stress differentials and not differences in surface roughness along coastlines. The development of low-level westerlies near 10°N and a low-level easterly jet to the north near 20°N (the West African easterly jet) results in strong meridional gradients in the vertical flux of \( u \)-momentum. The effect of surface friction under the westerlies is easterly acceleration and under the easterlies it is westerly acceleration. Therefore, a negative vorticity tendency is generated, and this mechanism is important for balancing the forced positive vorticity tendency.

Because of the perturbation of the low-level winds and the enhanced roughness of the land surface, this frictional mechanism is stronger over the land surface. However, the difference in roughness between the land and ocean surface is not fundamentally responsible for the continental precipitation enhancements. If the land surface has the same roughness as the ocean surface in the GCM, the precipitation perturbation is about 25% smaller, but it still occurs because the meridional structure in the surface winds remains.

This study is aimed at contributing to our basic understanding of the processes involved in establishing the time-mean precipitation distribution over Africa in summer. Such an understanding is necessary to improve our ability to understand and predict variability in the region and to be able to distinguish between climate change and variability.

Acknowledgments. The author thanks Lesley Greene for integrating the GCM to produce Fig. 2. This research was supported by Grant ATM-9300311 from the National Science Foundation.

REFERENCES