Generation of the African Easterly Jet and Its Role in Determining West African Precipitation

Kerry H. Cook

Atmospheric Science Program, Cornell University, Ithaca, New York

(Manuscript received 12 June 1997, in final form 9 February 1998)

ABSTRACT

An examination of analyses and model simulations is used to show that the African easterly jet forms over West Africa in summer as a result of strong meridional soil moisture gradients. In a series of GCM experiments, the imposition of realistic surface wetness contrasts between the Sahara and equatorial Africa leads to strong positive meridional temperature gradients at the surface and in the lower troposphere; the associated easterly shear in the atmosphere is strong enough to establish easterly flow—the African easterly jet—above the monsoon westerlies at the surface. Positive temperature gradients associated with the summertime distributions of solar radiation, SSTs, or clouds are not large enough to produce the easterly jet in the absence of soil moisture gradients. A thermally direct ageostrophic circulation is identified that can accelerate the largely geostrophic zonal flow and maintain the jet.

While moisture converges throughout the lower troposphere over East Africa, moisture divergence between 600 and 800 mb overlies low-level convergence over West Africa to the south of the African easterly jet. This moisture divergence is important for determining the total column moisture convergence. Since the moisture divergence is closely tied to the jet dynamics, and the jet’s magnitude and position are sensitive to SST and land surface conditions, a mechanism by which the West African precipitation field is sensitive to surface conditions is suggested.

1. Introduction

The African easterly jet (AEJ, also known as the West African jet) is a prominent feature of the complicated zonal wind structure that forms over northern Africa in summer. Figure 1 shows the jet from two reanalyses, plotted at the level of maximum easterly velocity. Figure 1a is the zonal wind at 600 mb from the July National Centers for Environmental Prediction (NCEP) climatology (Kalnay et al. 1996), and Fig. 1b is the 500-mb zonal wind in the European Centre for Medium-Range Weather Forecasts (ECMWF) climatology (Trenberth 1992); the jet is plotted at its maxima in each dataset. In both cases the jet is confined to a width of 5°–10° of latitude, but the NCEP version of the jet extends farther east. Both climatologies place the jet core near 15°N on the west coast; a tendency for southward displacement inland is more pronounced in the NCEP data. Maximum easterly velocities are 11 m s⁻¹ at 500 mb in the ECMWF data, and 12.5 m s⁻¹ at 600 mb in the NCEP data. The disparate levels of the jet maximum in the analyses may be partially related to the pressure levels chosen for reporting (850, 700, 600, and 500 mb in NCEP, and 850, 700, and 500 mb in ECMWF). However, Burpee (1972) shows figures based on 8 yr of observations with a 10 m s⁻¹ jet maximum near 650 mb.

Figure 2 shows a cross section of the zonal wind velocity on the Greenwich meridian from the NCEP July climatology. Low-level westerlies between the equator and about 20°N mark the African monsoon. Northward flow onto the continent across the Guinean coast curves eastward to form the equatorward section of cyclonic flow about a thermal low centered over Saharan Africa. Surface easterlies prevail north of the thermal low, forming the Harmattan winds that carry dust westward within the Saharan air layer (e.g., Karyampudi and Carlson 1988). The tropical easterly jet, which is part of the anticyclonic circulation about the Asian monsoon high, forms near the tropopause. Above the low-level westerlies, and distinct from the Harmattan easterlies and the tropical easterly jet, is the African easterly jet maximum at 600 mb and 13°N.

The jet may be instrumental in creating an environment in which African wave disturbances develop through baroclinic and barotropic instability (e.g., Renick 1976; Thorncroft and Hoskins 1994a,b) and may play a role in determining the region’s precipitation distribution through these wave disturbances (e.g., Payne and McGarry 1977; Rowell and Milford 1993) or through its role in determining the large-scale column precipitation.
Fig. 1. Observations of the zonal wind over northern Africa in July at (a) 600 mb from the NCEP climatology (Kalnay et al. 1996) and (b) 500 mb from the ECMWF climatology (Trenberth 1992). Contour interval is 1 m s\(^{-1}\).
moisture convergence (Rowell et al. 1992). In addition, the African wave disturbances have long been identified as sources of tropical cyclone activity in the Atlantic (e.g., Frank 1970). A better understanding of why the jet forms, and its sensitivity to surface conditions, will be useful for understanding the mechanics of the region’s basic climate dynamics as well as its intra- and interannual variability; such an understanding is necessary to advance our prediction capabilities.

Burpee (1972) states that the AEJ is a “response to the surface baroclinic zone and the reversal of the temperature gradient in the middle troposphere.” Other authors (e.g., Walker and Rowntree 1977; Krishnamurti et al. 1979; Chang 1982) echo a similar understanding, but the cause of the low-level baroclinicity is not demonstrated. While it seems reasonable to assume that the surface wetness gradient enhances the positive meridional temperature gradient imposed through the solstitial solar forcing, one cannot assume that this gradient will be strong enough to generate easterly flow above the surface westerlies or that it is the only (or most) important factor. Other factors that could be important are sea surface temperatures, since the Gulf of Guinea is generally cool due to upwelling and cloud distributions. Druyan (1989), for example, suggests that cloud distributions lead to the generation of the AEJ.

Several studies on synoptic timescales point to a central role for soil moisture in the generation of the baroclinic zone and its consequence, the AEJ. Most telling is the work of Cunnington and Rowntree (1986), in their study to diagnose the causes of excessive precipitation over the Sahara in an atmospheric general circulation model (AGCM). Because they were interested in operational prediction for northern Africa, the focus is on the effects of initial conditions (as opposed to the effects of boundary conditions, as in a climate study). Figures 7a and 7b in Cunnington and Rowntree (1986) show that the AEJ is well defined in their model only when the Saharan atmosphere is initially dry. They do not discuss this result, perhaps because the AEJ is essentially absent in the observations (at 20°E longitude) with which they compare.

The purpose of this study is to identify the cause or causes of the African easterly jet in the summertime climatology. In the next section, data from the NCEP reanalysis project (Kalnay et al. 1996) are analyzed to
investigate the degree to which the jet is geostrophic and, therefore, attributable to the reversed meridional temperature gradient. The structure of the ageostrophic wind is also discussed. Results from simulations with a GCM are reported in section 3, with the purpose of isolating the boundary conditions that are the cause of the baroclinicity and the jet. The ageostrophic flow that must be present to accelerate the flow and maintain the jet is discussed in section 4. Sections 5 and 6 include investigations of how and why surface conditions can impact the location and intensity of the jet. Section 7 addresses the role of the African easterly jet in determining the precipitation climatology. Conclusions are drawn in section 8.

2. Dynamics of the African easterly jet

As a first step in understanding the dynamics and generation of the AEJ, we investigate the degree to which the jet is geostrophic. Here, geostrophy is defined as a balance between the pressure gradient and the local Coriolis accelerations according to

$$v_g = \frac{1}{f} k \times \nabla \phi.$$  \hspace{1cm} (1)

This definition of geostrophy was chosen rather than geostrophy-0,

$$v_g, 0 = \frac{1}{f_0} k \times \nabla \phi,$$  \hspace{1cm} (2)

where $f_0$ is a reference latitude, so the ageostrophic wind expresses exactly any local imbalances between pressure gradient and Coriolis forces. Thus, the geostrophic wind referred to here is not nondiagnostic (see section 4).

Figure 3 shows the ageostrophic zonal wind over northern Africa at 600 mb from the NCEP climatology for July. The ageostrophic component of the wind, $v_a$, is calculated as the difference between the full wind $v$ and the geostrophic wind $v_g$ given by Eq. (1). The magnitude of the zonal component of $v_a$ (Fig. 3) is small compared with the full jet, and the geostrophic component (not shown) is very similar to the full zonal wind (Fig. 1a), even within 8° of the equator. On the northern jet boundary, the ageostrophic flow is easterly near the coast and makes the full jet 10% stronger; inland, the ageostrophic flow is westerly. On the equatorward side of the jet maximum, ageostrophic flow augments the easterly wind maximum inland, and so may contribute to a jet orientation that is slightly from southeast to northwest instead of due east to west. However, this orientation is not seen in the ECMWF analysis (Fig. 1b). The meridional component of the ageostrophic wind is more substantial and is discussed in section 4.

Figure 4a shows geopotential heights and total wind vectors at the level of the jet maximum in the NCEP reanalysis. A high develops over the land surface, centered at about 30°N and a little west of the Greenwich meridian. This is the Saharan high, described briefly by Koteswaram (1958). The Saharan high occurs at about...
FIG. 4. Geopotential height and total wind vectors from the NCEP climatology for Jul at (a) 600 mb and (b) 925 mb. Contour interval is 10 gpm; wind vector length of 10 m s$^{-1}$ is indicated.
the same latitude as the North Atlantic high to the west, but it is clearly a distinct feature. Relatively strong positive geopotential gradients to the south of the high support strong geostrophic easterly flow; this is the African easterly jet.

Low pressure covers northern Africa at very low levels (Fig. 4b), in sharp contrast to the Atlantic high to the west. North Africa’s thermal low appears as a westward extension of the immense thermal low associated with the Indian monsoon. The lowest heights associated with the African thermal low (or, more accurately, thermal trough) are located near 20°N.

Since the jet is essentially geostrophic, the easterly shear that accounts for the development of easterly flow above the monsoon westerlies (Fig. 2) must be associated with positive meridional temperature gradients below the level of the jet maximum. Previous work has associated the African easterly jet with such meridional gradients. For example, Burpee (1972) notes that the easterly wind maximum near 600 mb is “a response to the surface baroclinic zone and the reversal of the temperature gradient in the middle troposphere,” referring to the reestablishing of a negative meridional temperature gradient above the jet maximum.

The atmospheric moisture budget complicates the relationship between the geostrophic wind shear and the temperature gradient somewhat, since

$$\frac{\partial \phi}{\partial y} \approx - \frac{\partial T_v}{\partial y} = -(1 + 0.6q) \frac{\partial T}{\partial y} + 0.6T \frac{\partial q}{\partial y},$$

where $T_v$ is virtual temperature and $q$ is specific humidity. According to the NCEP reanalysis, the negative specific humidity gradient causes the virtual temperature gradient to be about 10% smaller than the temperature gradient between 10° and 20°N, where the strongest gradients occur. The effect diminishes to 5% at 700 mb.

Figure 5 shows a vertical cross section of meridional temperature gradients from NCEP at 12.5°N in July. Over Africa, which covers 15°W–50°E longitude at this latitude, positive temperature gradients extend from the surface. At 600 mb, the level of the jet core, a sign change in the gradient occurs over the continent, with the positive gradient replaced by a weak negative tem-
perature gradient associated with the insolation distribution. Positive meridional temperature gradients also occur at low levels over the eastern Pacific (120°–80°W), but they are much smaller than the gradients over Africa.

The vertical structure of the gradient implies that the origin of the lower atmosphere’s positive meridional temperature gradient over Africa is the land surface. According to the NCEP data, surface temperatures over northern Africa in summer reach over 308 K. The highest temperatures are in the west between 20° and 25°N (Fig. 6), and they are more than 10 K larger than surface temperatures of equatorial Africa. A comparison between Figs. 1a and 6 suggests the relationship between the African easterly jet and the meridional surface temperature gradient, since the latitude of the jet maximum is coincident with the strongest surface temperature gradient.

To first order, then, an understanding of the African easterly jet depends on understanding the surface meridional temperature gradient and its communication into the atmosphere. However, the associated ageostrophic circulations are also relevant, being potentially important for understanding the variability of the jet and also for understanding how the jet is maintained.

3. Cause of the African easterly jet

To determine how the jet is generated, experiments with a GCM are conducted. The GCM is a version of the Geophysical Fluid Dynamics Laboratory (GFDL) Climate Dynamics Group model, used here with R30 resolution (equivalent to about 2.25° lat × 3.75° long); there are 14 σ levels in the vertical located at σ = 0.015, 0.050, 0.101, 0.171, 0.257, 0.355, 0.460, 0.568, 0.676, 0.777, 0.866, 0.935, 0.979, and 0.997. Simple physical parameterizations are used, including moist convective adjustment and the bucket hydrological budget at the surface [see Manabe (1969) for a general description].

While a wide variety of surface features all over the globe might be influential, the climatological jet must be fundamentally related to features and structure on the African continent and/or the adjacent ocean. To focus on the most basic cause of the jet, the model was run with various idealized surface boundary conditions, with the goal of isolating the surface feature(s) responsible for the presence of the jet. Four experiments are discussed, as summarized in Table 1. In each simulation, the GCM was configured with fixed, zonally uniform clouds and zonally uniform July sea surface temperatures based on the observations of Shea et al. (1990). Solar forcing represents perpetual July and there is no
Table 1. GCM experiments.

<table>
<thead>
<tr>
<th>Name</th>
<th>Soil moisture</th>
<th>Land surface drag coefficient</th>
<th>SSTs</th>
</tr>
</thead>
<tbody>
<tr>
<td>SIM1</td>
<td>Realistic</td>
<td>0.001</td>
<td>Zonally uniform</td>
</tr>
<tr>
<td>SIM2</td>
<td>Uniform</td>
<td>0.001</td>
<td>Zonally uniform</td>
</tr>
<tr>
<td>SIM3</td>
<td>Realistic</td>
<td>0.001</td>
<td>Cold eastern Atlantic</td>
</tr>
<tr>
<td>SIM4</td>
<td>Realistic</td>
<td>0.003</td>
<td>Zonally uniform</td>
</tr>
</tbody>
</table>

diurnal cycle. The only land surface included is the African continent; it is a featureless version of Africa, with no topography and uniform surface albedo of 0.1. Each integration proceeds from a dry isothermal atmosphere at rest; the first 200 days are discarded, and the climatology is created by averaging over 1800 days. The boundary conditions specified for each of the last three simulations (SIM2, SIM3, and SIM4) differed from SIM1 in only one boundary condition each.

Figure 7 shows the African easterly jet from SIM1, which includes a realistic but prescribed soil moisture distribution from Mintz and Walker (1993) as shown in Fig. 8a. [SIM1 also has a “smooth” land surface; the coefficient used in the bulk aerodynamic formulation of momentum, heat, and moisture fluxes from the surface (see section 6) is the same over land and ocean.] The AEJ in this climatology is centered on the west coast, in a position more similar to the ECMWF representation of the jet (Fig. 1b) than the NCEP jet (Fig. 1a). The jet in the GCM is about half the magnitude of the observed jet and extends less deeply into the continental interior; the modeled jet has a latitudinal scale similar to the observations.

Calculated surface temperatures are shown in Fig. 8b; note the approximate collocation of the soil moisture (Fig. 8a) and temperature gradients. Model surface temperatures are warmer than those from the NCEP analysis (Fig. 6). Over equatorial Africa, the surface in the GCM is about 5 K warmer than in the reanalysis, and it is up to 10 K warmer over the Sahara. This is not unexpected considering the simplified boundary conditions, especially the prescribed low surface albedo. However, the resulting strong surface temperature gradients suggest that the jet in the GCM should be stronger than the jet in the reanalysis, but it is weaker. The warm surface temperatures are not communicated into the model atmosphere effectively enough, and lower atmospheric meridional temperature gradients over North Africa are about half the magnitude of those in the NCEP reanalysis.

Fig. 7. July zonal wind over northern Africa at 680 mb from a GCM with simplified surface boundary conditions. Contour interval is 1 m s⁻¹; heavy line indicates the African coast as resolved in the R30 GCM.
Figure 8. (a) Soil moisture prescription for GCM Jul simulation. Contour interval is 2 cm of water. (b) Jul surface temperature climatology from GCM. Contour interval is 2 K; heavy line denotes Africa as resolved in the model.

Figure 9a shows a cross section of the zonal wind across northern Africa at 10°W longitude from the idealized GCM simulation. The jet maximum occurs at 680 mb, somewhat lower in the troposphere than in the reanalysis (Fig. 2). However, the vertical structure is similar, with the easterly wind maximum of the African easterly jet clearly distinguished from low-level Harmattan easterlies to the north and monsoon westerlies below. Even though there are differences between the modeled and analyzed jet, it is clear from Figs. 7 and 9a that the GCM with these simple boundary conditions produces an African easterly jet.

Figure 9b shows the cross section of the zonal wind from SIM2, which has boundary conditions identical to those of SIM1 except that the soil moisture is prescribed to represent savanna conditions (5 cm) everywhere on the continent. The African easterly jet does not occur in this case. Monsoon westerlies are generated, as are Harmattan winds, only a little weaker than in the case with realistic soil moisture distribution. However, the surface winds extend deeper into the troposphere with no easterly wind maximum aloft.

A number of GCM simulations with various boundary conditions were performed to identify the conditions under which the jet forms. In these simulations differences in land and sea surface conditions influence the position and/or intensity of the jet in relatively modest
ways. However, an African easterly jet only occurs in the GCM if there is a realistic soil moisture gradient.

In order for the jet to form, the low-level temperature gradient must be positive and large enough so that the associated wind shear leads to easterly flow above the monsoon westerlies. The source of the atmospheric temperature gradient is the surface temperature gradient. Without the negative meridional soil wetness gradient, the temperature gradient would be positive but not large enough to cause easterly flow above the westerlies. With realistic soil moisture prescribed over Africa in the GCM, the surface temperature gradient is realistically strong. The communication of the gradient into the atmosphere, however, is not as complete as in the reanalysis, and the jet is weaker in the GCM. However, the fundamental relationship between the temperature gradient and the shear holds in the GCM as in the reanalysis, and the jet maximum occurs at the pressure level where the positive temperature gradient is replaced by the negative gradient of the free atmosphere (Burpee 1972) or, as formulated by Thorncroft (1998, manuscript submitted to Quart. J. Roy. Meteor. Soc.), where the temperature profile leaves the dry adiabat and follows a moist profile.

Previous work (e.g., Walker and Rowntree 1977; Krishnamurti et al. 1979) has demonstrated that the region’s time-dependent circulation is sensitive to soil moisture distributions. Here we go a step farther and state that the climatological African easterly jet owes its existence to the presence of the negative soil moisture gradient. Summer insolation and dryness over Saharan Africa in summer are the two necessary and sufficient conditions for the formation of the jet.

To support the argument that the soil moisture gradient is the crucial land surface condition required for the jet to form, consider how the land surface temperature distribution is determined over West Africa in summer through the surface heat budget. Table 2 shows the components of the surface heat budget from the NCEP reanalysis for July. Three locations are represented, each of which is on the Greenwich meridian: at 7°N (a location of strong continental precipitation), 15°N (Sahel; latitude of the jet), and 28°N (Sahara; latitude of strong surface heating). Two components of the heat budget encourage a positive meridional temperature gradient along this transect. One is the latent heat flux, which increases (becomes less negative) to the north. The other component is the incident solar radiation, which is about 50% greater over the Sahara than it is at 7°N. All of the other components have negative gradients, which would support cooler temperatures to the north. For example, the upward solar flux increases by more than a factor of 3 in association with an increase in surface albedo. However, the increase in the solar incident dominates, resulting in a net increase in solar heating of the surface with increasing latitude.

One reason that the incident solar flux increases with increasing latitude is that the summertime insolation at the top of the atmosphere increases from the equator to about 23°N. Another reason is that cloud amounts decrease to the north, and the atmospheric albedo decreases accordingly. In the GCM simulation with uniform soil moisture, summertime solar forcing results in a weak positive meridional temperature gradient over tropical West Africa that is confined below 750 mb. The geostrophic response to this positive meridional temperature gradient must cause the westerlies to weaken somewhat with height (Fig. 9b), but the effect is not strong enough to reverse the wind direction.

In addition to influencing the shortwave fluxes, clouds and water vapor also interact strongly with longwave radiation. Less clouds and water vapor in the north result in smaller downward longwave surface fluxes in the north. In addition, the upward longwave flux at the surface increases from south to north due to increased thermal emission from the warmer surface. The net result is an increase in the net longwave cooling of the surface as one goes north. The increase in net longwave cooling with increasing latitude is larger than the increase in net solar heating, so the net radiative heating rate over the Sahara is smaller than the radiative heating to the south. Thus, the reversed meridional temperature gradient that causes the African Easterly jet is not explained by meridional structure in cloud amounts or radiative heating.

The source of the strong positive gradient is found in the sensible and latent heat flux distributions (Table 2). Latent cooling of the surface is comparable to net solar heating rates over equatorial Africa but drops to zero over the Sahara. An order-of-magnitude increase in sensible heating rates, accompanied by increases in thermal emission, compensates for the loss of latent cooling. Figures 10a and 10b show the distributions of

<table>
<thead>
<tr>
<th></th>
<th>ITCZ (7°N)</th>
<th>Sahel (15°N)</th>
<th>Sahara (28°N)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Downward solar flux</td>
<td>218 W m⁻²</td>
<td>291 W m⁻²</td>
<td>383 W m⁻²</td>
</tr>
<tr>
<td>Upward solar flux</td>
<td>−42 W m⁻²</td>
<td>−83 W m⁻²</td>
<td>−124 W m⁻²</td>
</tr>
<tr>
<td>Net solar heating</td>
<td>176 W m⁻²</td>
<td>−208 W m⁻²</td>
<td>−259 W m⁻²</td>
</tr>
<tr>
<td>Downward longwave flux</td>
<td>403 W m⁻²</td>
<td>419 W m⁻²</td>
<td>363 W m⁻²</td>
</tr>
<tr>
<td>Upward longwave flux</td>
<td>−442 W m⁻²</td>
<td>−493 W m⁻²</td>
<td>−516 W m⁻²</td>
</tr>
<tr>
<td>Net longwave heating</td>
<td>−39 W m⁻²</td>
<td>−74 W m⁻²</td>
<td>−153 W m⁻²</td>
</tr>
<tr>
<td>Net radiative heating</td>
<td>137 W m⁻²</td>
<td>134 W m⁻²</td>
<td>106 W m⁻²</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>−8 W m⁻²</td>
<td>−95 W m⁻²</td>
<td>−101 W m⁻²</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>−134 W m⁻²</td>
<td>−31 W m⁻²</td>
<td>−0.1 W m⁻²</td>
</tr>
</tbody>
</table>

Table 2. Surface heating rates over northern Africa in July from NCEP.
Fig. 10. (a) Sensible and (b) latent heat fluxes from the surface over Africa in Jul from NCEP. Contour interval is 20 W m$^{-2}$.

sensible and latent heat fluxes from the surface to the atmosphere from the NCEP July climatology. The largest gradients in both fluxes are coincident with the largest soil moisture gradient (Fig. 8a) and are centered on the latitude of the jet (Fig. 1a). Going north from the Guinean coast, soil moisture values decrease and latent cooling of the surface is disabled. Surface temperatures increase and drive a stronger sensible heat flux. The larger sensible flux does not completely compensate for the loss of latent cooling, as thermal emission from the surface also increases with increasing surface temperature (Table 2).

Replacing latent cooling of the surface by sensible and radiative cooling has two consequences. First, the surface temperature is driven higher until cooling by the sensible and thermal radiation terms balance the solar heating. Sensible and thermal heat fluxes operate at higher temperatures than the latent heat flux; if evaporation cools a surface with a certain heat flux incident on it, the equilibrium temperature of the surface will be lower than if the surface is dry and cooled by sensible and thermal heat fluxes. The second consequence is that the heat flux from the surface is distributed in the lower troposphere by dry convection and diffusion, rather than being deposited in the middle and upper troposphere by condensational heating. A pronounced reversal of the meridional temperature gradient in the lower troposphere results, and the African easterly jet appears.

This heat budget analysis is based on the NCEP reanalysis, which is a model-dependent calculation not highly constrained by observations in the assimilation scheme (Kalnay et al. 1996). However, similar results were found by Wai et al. (1997) in their analysis of the Hydrological Atmospheric Pilot Experiment (HAPEX)–Sahel ground-based and radiosonde observations collected during the 1992 rainy season. They point to the critical role of soil moisture gradients for maintaining the African easterly jet in August and part of September through controls on the surface heat budget in much the same way as described above.

4. Ageostrophic flow

While the African easterly jet is a geostrophic feature, the ageostrophic flow in the vicinity of the jet is important to understand for two reasons. First is the question of the maintenance of the jet. Since the zonal wind is not accelerated in strict geostrophic balance, the ageostrophic component of the wind is responsible for maintaining the jet against dissipation. The second reason is the connection with the vertical velocity and precipitation fields; the jet is interesting for its proximity to the continental section of the intertropical convergence zone (ITCZ) and for its possible role in influencing vertical velocity and precipitation distributions and their variability.

While the zonal component of the ageostrophic flow is small compared with the geostrophic zonal wind (Fig. 3), the meridional flow is largely ageostrophic. Figure 11 shows contours of the ageostrophic meridional velocity, along with shaded contours indicating vertical velocity, from NCEP. The ageostrophic meridional velocity, defined as the difference between the geostrophic meridional velocity from Eq. (1) and the total meridional wind, has a quadrupole structure, with equatorward flow below about 800 mb north of 20°N and poleward low-level flow to the south. Relatively strong equatorward ageostrophic velocities occur between about 800 and 500 mb south of 20°N, with weak northerly flow to the north.

In combination with this ageostrophic meridional flow, the distribution of vertical velocities (filled contours in Fig. 11) suggests a vertically confined circulation system in the $y$-$z$ plane immediately north of the ITCZ. The deep ITCZ system is marked by the vertical $p$ velocity maximum between 300 and 400 mb at 7°N. To the north, between 15° and 20°N, a second and even stronger upward velocity maximum occurs between 800
and 850 mb. This upward motion connects the region of low-level ageostrophic wind convergence and the region of divergence aloft. The entire circulation system is confined below 500 mb. It can serve as a source of kinetic energy for the jet, a mechanism that Burpee (1972) speculated may be important for maintaining the jet. This thermally direct, “secondary” circulation is also discussed by Wai et al. (1997) in the context of their analysis of HAPEX–Sahel observations. While that program was not designed to capture such a circulation, Smith et al. (1994) found an analogous circulation in the American midwest that developed “in response to well organized surface gradients of moisture and sensible heating.”

The vertical velocity field (Fig. 11) in the deep Tropics is nearly identical in structure to the diabatic heating field away from the surface. In the subtropics, generously defined here as 10°–35°N, the vertical velocity maximum is not coincident with the heating maximum but occurs above and equatorward of the level of maximum heating at the surface. Therefore, the diabatic heating is not being balanced totally by adiabatic cooling as in the deep Tropics. The presence of northward low-level flow (Fig. 11) up the positive meridional temperature gradient suggests that the poleward advection of cooler air plays a role in balancing the diabatic heating. Cook (1997) discusses the role of this southerly monsoon flow across the Guinean coast in supporting the summertime precipitation in West Africa by advecting moisture and low vorticity air into the region.

As defined by Eq. (1), the geostrophic wind is divergent, so the ageostrophic wind is not precisely diagnostic of the vertical velocity field (Blackburn 1985). There are two components to the wind divergence, given by

\[
- \frac{\partial \omega}{\partial \rho} = \nabla \cdot \mathbf{v} = \nabla \cdot \mathbf{v} + \frac{\partial \phi}{\partial x} \frac{\partial (f^{-1})}{\partial y},
\]

where \( \mathbf{v} = \mathbf{v}_g + \mathbf{v}_a \), and the first equality derives from mass continuity. The last term in Eq. (4) can be written

\[
\frac{\partial \phi}{\partial x} \frac{\partial (f^{-1})}{\partial y} = -\frac{\cos \theta}{2 \Omega a \sin^2 \theta} \frac{\partial \phi}{\partial x} = -\frac{\beta}{f} \mathbf{v}_a \cdot \mathbf{n},
\]

where \( \mathbf{n} \) is the unit normal vector to the earth.
This term goes to infinity on the equator, expressing the invalidity of the definition of geostrophy there. In the Tropics and subtropics, this term is a potentially important source of vertical velocity in the presence of pronounced longitudinal structure in the geopotential field. The longitudinal geopotential gradients (meridional geostrophic wind) are small over West Africa (Fig. 4), and \((-\beta/f)u_x \approx 10^{-6} \text{ s}^{-1}\). This is the same order of magnitude as \(\partial u_y/\partial x\), but \(\partial u_y/\partial y\) is an order of magnitude larger (see Fig. 11). Thus, most (but not all) of the vertical velocity is associated with the meridional divergence of the ageostrophic wind.

5. Influence of SST distributions

Significant empirical relationships have been noted between SST anomalies and African precipitation variability (e.g., Folland et al. 1986; Rowell et al. 1995). In addition, there are suggestions in the literature that there is a relationship between the strength and/or position of the African easterly jet and precipitation (Fontaine and Janicot 1992; Rowell et al. 1992). Therefore, it is of interest to consider the dependence of the jet on structure in the SST field. Understanding this connection may help us to better understand the mechanisms that connect SST distributions and African rainfall.

A GCM simulation was carried out to isolate the influence of longitudinal structure in the SSTs near the African continent. In SIM3 (see Table 1), cold SST anomalies are superimposed on the zonally uniform SST distribution in the Gulf of Guinea and the eastern Atlantic to make the SSTs off the west coast of Africa and in the Gulf of Guinea more realistic (Fig. 12). The SST anomalies are Gaussian shaped, with half-widths, locations, and magnitudes chosen to mimic the effects of the eastern boundary currents. Both anomalies have amplitudes of 4 K. One is centered at 21°S and 15°E, with a longitudinal half-width of 15° and a latitudinal half-width of 20°. This approximates the influence of the Benguela Current and upwelling in the Gulf of Guinea on SSTs. The other anomaly is centered at 21°N and 15°W, with a longitudinal half-width of 20° and a latitudinal half-width of 15°. This is meant to represent the effects of the Canary Current.

The presence of these cool SSTs has some effect on the jet, making it about 1 m s\(^{-1}\) (20%) stronger and

---

**Figure 12.** Prescribed SSTs and calculated land surface temperatures from GCM SST anomaly experiment. Contour interval is 2 K.
located a couple of degrees (one GCM grid point) farther south (not shown). These changes are consistent with differences in the surface temperature gradient in the lower troposphere over West Africa (cf. Figs. 8b and 12), so the differences in the jet are captured by differences in the geostrophic zonal wind.

The location and intensity of the jet is tied to the region of strong temperature and geopotential gradients between $10^\circ$ and $20^\circ$N. Cold SSTs in the Gulf of Guinea cause this gradient to increase because lower atmosphere temperatures over the land between $8^\circ$ and $12^\circ$N are up to 1.5 K cooler due to cold air advection across the Guinean coast. The cool SSTs imposed in the Northern Hemisphere eastern Atlantic are not influential in modifying the jet since they do not impact the meridional geopotential gradient in the model.

For completeness, a GCM experiment was run with uniform soil moisture (at 5 cm) and the SST anomalies described above. In this case the AEJ does not appear, confirming the fundamental role of the soil moisture distribution.

6. Influence of atmospheric boundary layer transports

The communication of the surface temperature gradient into the lower troposphere is a significant element determining the intensity of the African easterly jet. Since the magnitude of the meridional temperature gradient decreases with height from the surface, to be replaced by a negative temperature gradient in the free atmosphere (Fig. 5), one might expect that enhancing fluxes from the surface to the atmosphere would cause the jet to be stronger. In fact, the opposite occurs in the GCM.

Momentum, sensible heat, and moisture transports from the surface to the lowest level of the model atmosphere are parameterized using the standard bulk aerodynamic formulation (e.g., Washington and Parkinson 1983). The vigor of these transports is determined, in part, by the setting of a “drag coefficient,” $C_D$. A higher drag coefficient enhances the transports, and is designed to represent the effects of a “rougher” surface on turbulent transports within the mixed layer. (The depth of the mixed layer in the model is prescribed, and given as the depth over which heat, momentum, and moisture are vertically diffused. The top of the mixed layer is $\sigma = 0.777$ in all of the experiments discussed here, and this setting may be related to the fact that the jet appears too near the ground in the simulations.)

In SIM1, the surface drag coefficient over land is the same as over the ocean surface ($C_D = 0.001$), and it is the same for momentum, heat, and water. Figure 13a shows the jet in SIM4, a GCM experiment that differs from SIM1 in that the surface drag coefficient on the land surface is three times that over the ocean surface (Table 1). Including a mechanism for enhanced momentum, sensible heat, and moisture fluxes due to a more well-developed mixed layer over land causes the strength of the West African jet to diminish by about 20% ($1 \text{ m s}^{-1}$ in the GCM). As in SIM3, the SST anomaly simulation, the difference in the magnitude of the jet maximum is accounted for by differences in the magnitude of the geostrophic zonal wind. Consistent with that decrease, the meridional temperature gradient between $10^\circ$ and $20^\circ$N is smaller when land surface drag is enhanced.

The enhanced heat fluxes to the atmosphere over land cause land surface temperatures to be generally smaller (cf. Figs. 8b and 13b). Over tropical and Sahelian West Africa, evaporation is important for cooling the surface (Table 2 and Fig. 10) and evaporation rates are larger when the surface drag is larger. An increase in condensational heating and precipitation over West Africa results (Cook 1995), improving the model’s simulation of West African precipitation. Because the added energy is going primarily into evaporating water, the lower atmosphere is cooler with enhanced fluxes from the land surface. Temperatures in the lower atmosphere are controlled, in large part, by sensible heat fluxes from the surface and diffusion and dry convection in the lower
atmosphere. The climatological sensible heat flux from the surface to the atmosphere is smaller when $C_d$ is larger because the difference in the surface and lowest-level air temperatures is reduced by the strong evaporative cooling of the surface.

Over the Sahara, where evaporative cooling is disabled, sensible heating of the lower atmosphere is enhanced by the larger $C_d$ and the lower atmosphere warms. The difference in the lower atmospheric temperature between the Tropics and the subtropics is, therefore, greater with enhanced drag, suggesting an increase in the meridional temperature gradient and a strengthening of the African easterly jet. However, as for the SST anomaly experiment, the critical factor determining the magnitude of the jet is the gradient at latitudes close to the jet. The evaporation rate increases are greatest just north of the jet, and this decreases the temperature gradient between 10° and 20°N. A decrease in the magnitude of the jet results.

This mechanism might be too subtle for these idealized GCM experiments to capture definitively. The horizontal resolution is coarse, and the treatment of surface processes is very simple. For example, the influence on the jet might be quite different if the soil moisture were able to respond instead of being fixed. These results suggest that the sensitivity to land surface drag should be explored within a mesoscale model with a more sophisticated land surface treatment.

7. Role of the African easterly jet in determining the precipitation climatology

The presence of the African easterly jet has been associated with the occurrence of African wave disturbances and, arguably, with the modulation or even the instigation of intense, small-scale precipitation events (e.g., Payne and McGarry 1977; Chen and Ogura 1982; Rowell and Milford 1993). The jet is thought to be hydrodynamically unstable and African wave disturbances may be an expression of this instability (e.g., Thorncroft and Hoskins 1994a,b). However, there are opposing views in the literature. Schubert et al. (1991), for example, suggest that the reversed potential vorticity gradients that mark the region of instability between the ITCZ and the African easterly jet are due only to the presence of a well-defined ITCZ over West Africa. Thorncroft (1998, manuscript submitted to Quart. J. Roy. Meteor. Soc.) finds that both the jet and the ITCZ contribute to the reversal of the potential vorticity gradient and, therefore, the unstable environment.

While African wave disturbances are resolvable in GCMs with horizontal resolution equivalent to approximately 5° long or finer (e.g., Reed et al. 1988), the squalls that may form preferentially near their troughs are not. To address connections between the African easterly jet and precipitation in the context of the large-scale circulation, the column moisture budget climatology in the model is examined.

Figure 14a shows the precipitation climatology from SIM1. This is a relatively realistic simulation of the African precipitation climatology, especially considering the idealized boundary conditions. [For comparison, Figs. 1 and 2 of Cook (1997) show observed and analyzed precipitation climatologies, and a more realistic simulation with the GCM.] Precipitation magnitudes in East Africa fall within the range of the observed and reanalyzed climatologies. The main fault of the SIM1 precipitation is weak West African precipitation.

Figure 14b shows the precipitation climatology from SIM2. This simulation differs from SIM1 only in having uniform prescribed soil moisture (Table 1), and because of this, there is no African easterly jet. The precipitation maximum over East Africa is shifted to the west relative to that in SIM1, and the West African precipitation is stronger. This suggests that the presence of the jet, and the associated barotropic and baroclinic instabilities, is not necessary for precipitation to occur in West Africa, at least in the GCM.

Despite the presence of higher soil moisture values between 5°S and about 12°N in SIM1 (see Fig. 8a), precipitation across tropical and Sahelian west and central Africa is significantly smaller (and less realistic) in SIM1 than in SIM2. As seen in Fig. 15a, precipitation rates are up to 4 mm day$^{-1}$ smaller in SIM1. Figure 15b shows the difference in surface evaporation rates between the two model simulations. Evaporation rates in SIM1 are slightly larger than in SIM2 equatorward of 15°N, where the SIM1 soil moisture is greater than in SIM2. North of about 15°N, where SIM1 is drier than SIM2, evaporation rates are smaller in SIM1. Clearly, differences in evaporation do not explain the differences in the precipitation distribution.

Figure 15c shows the difference in the vertically integrated column moisture convergence between SIM1 and SIM2. This difference field is very similar in structure and magnitude to the precipitation difference field (Fig. 15a). Thus, the influence of the soil wetness distribution on the precipitation field is exerted indirectly, through the atmosphere’s hydrodynamical response to surface heating, and not directly through the evaporation field.

The different soil moisture prescriptions in SIM1 and SIM2 result in very different climatologies. For example, the thermal low over northern Africa is deeper and situated farther north in the simulation with realistic soil moisture (SIM1). Also, there are pronounced differences in the low-level flow, and these are primarily responsible for the differences in precipitation shown in Fig. 15a. However, the role that the African easterly jet plays in determining the precipitation field can be isolated by comparing the moisture convergence fields at the height of the jet.

Figures 16a and 16b show the water vapor flux convergence $-\nabla \cdot (p_q \mathbf{v})$ within the GCM layer from about 623 to 728 mb for SIM1 and SIM2, respectively, where $\Delta \sigma$ is the layer thickness and $p_q$ is surface pres-
Fig. 14. Precipitation climatology for (a) SIM1 and (b) SIM2, described in Table 1. Contour interval is 2 mm day$^{-1}$.

sure. Values are calculated on $\sigma$ levels and converted to pressure levels; the quantity $(p,q_v)$ is sampled daily in the course of the GCM integration, so the effects of transients as well as the time-mean fluxes are included. Arrows in Figs. 16a and 16b indicate the wind in the center of the layer, at 676 mb. The AEJ appears as the easterly flow over West Africa between about 10° and 15°N in Fig. 16a. In SIM2 (Fig. 16b), there is weak westerly flow on the coast, which strengthens inland. In SIM1, there is weak moisture divergence immediately north and south of the jet and in a broad band across the Sahel to the east of the jet (Fig. 16a). In SIM2, a region of moisture divergence is centered over West Africa, as the westerly flow carries moisture into East Africa. Thus, the presence of the jet inhibits moisture divergence in West Africa south of about 15°N and enhances moisture divergence at Sahelian latitudes. According to these simulations, then, a strong jet would be associated with weaker precipitation in the Sahel and stronger precipitation north of the Guinean coast. However, it must be kept in mind that a change in the jet’s intensity or location is likely associated with other changes in the system, and the low-level response to those changes may mask the role of the African easterly jet.

The influence of the jet on the column moisture convergence is largest below the level of the jet maximum. Figure 17a shows the layer water vapor flux convergence for the layer centered at 777 mb (825.5–728.4 mb) from SIM4, which has the strongest West African precipitation of the four simulations and the most realistic representation of the deep boundary layer and
Fig. 15. Differences in GCM climatologies, for SIM1–SIM2, for (a) precipitation, (b) evaporation, and (c) vertically integrated column moisture convergence. All contour intervals are 2 mm day$^{-1}$. 
low-level moisture fields. As in all of the simulations with the jet, within the 825- to 728-mb GCM layer, and also in the 728- to 623-mb layer above, moisture divergence occurs in a broad band over the Sahel, dipping equatorward over tropical West Africa (Fig. 17a), and this diminishes the total column moisture convergence in these regions. This is in contrast to the vertical structure to the east, where moisture converges throughout the column. The vertical structure in the water vapor flux convergence seen in the model simulations is similar in the NCEP analysis. However, the NCEP vertical resolution is not well suited for capturing this structure, with reporting levels only at 850 and 700 mb.

Longitudinal structure in the moisture convergence field is due to structure in both the moisture and wind fields, but the latter dominates. Figure 17b shows the water vapor mixing ratio at 777 mb from the GCM. There is a positive moisture gradient, with East Africa more moist than the west, especially near the latitude of the jet. Thus, the African easterly jet is flowing down the moisture gradient, advecting moisture off the African continent over West Africa.

Figure 17c shows wind convergence and wind vectors at 777 mb. The moisture convergence pattern is reproduced in the wind convergence field, with divergence over much of West Africa and the entire Sahel region. The superposition of the easterly flow of the jet near the coast and westerly flow into the deep convergence system over East Africa creates the regions of wind and moisture divergence.
In the SST anomaly GCM simulation (SIM3), in which the African easterly jet is stronger, the moisture divergence at 777 mb is larger than in SIM4 by a factor of 2. The correlation between the strength of the jet and the moisture divergence suggests a mechanism by which the soil moisture field, as an important governor of the jet through the surface temperature distribution, may influence precipitation rates over West Africa. A positive feedback system may be present, in which dryness over Sahelian West Africa is associated with a stronger African easterly jet, more pronounced moisture divergence below the level of condensation, reduced precipitation, and additional surface drying.

Rowell et al. (1992) also suggest that such a feedback among midlevel moisture divergence, the jet, and rainfall exists, and they emphasize the position of the jet. In an analysis of output from a series of GCM simulations with SSTs for individual years, they find that the location of the moisture divergence is highly correlated with both the position of the African easterly jet and the magnitude of Sahelian rainfall.

8. Conclusions

Results from GCM simulations and comparisons with the NCEP reanalysis show that there are two necessary and sufficient conditions for the formation of the African easterly jet over West Africa in the model, namely, summer insolation and a realistic soil moisture distribution. The jet is essentially geostrophic and owes its existence to the presence of a positive surface temperature gradient, which, according to the thermal wind relation, induces easterly shear over the surface monsoon westerlies. The height of the jet is determined by the level at which the surface-induced positive temperature gradient is replaced by the free atmosphere's negative temperature gradient. Other possible sources of the positive temperature gradient, such as insolation alone or in combination with cloud or SST distributions, are eliminated as sources of the jet.

Variations in surface conditions (SST and land surface roughness) are shown to influence the position and intensity of the jet somewhat. There is the potential for fairly subtle differences in the jet in association with surface condition differences. At least in the idealized GCM experiments, the jet features are not determined by the continental-scale meridional temperature gradient, but by the intensity and position of the strong gradient near the latitude of the jet.

While the zonal flow of the jet is very nearly geostrophic, the meridional flow is largely ageostrophic. Between 10° and 25°N, a thermally direct circulation forced by low-level heating is generated in both the GCM simulations and the NCEP reanalysis. Strong upward vertical velocities, with magnitudes greater than those within the ITCZ, are centered near 17°N and 800 mb. The ageostrophic flow converges below this level, and diverges above it, with the entire circulation confined below 500 mb. Such a circulation converts potential to kinetic energy and can maintain the African easterly jet against dissipation.

There are differences in the vertical structure of the circulation and precipitation fields over West and East Africa, and the African easterly jet plays a central role in distinguishing these two regions. Over East Africa,
where virtually all of the low-level moisture convergence is driven by mid- and upper-tropospheric condensational heating (Cook 1994, 1997), moisture convergence occurs at every level below the condensational heating maximum. In contrast, the total column moisture convergence over West Africa is determined not only by the response to condensational heating aloft, but also to low-level heating. The African easterly jet is part of the response to low-level heating, being the equatorward portion of the Saharan high that is forced by dry convection and the diffusion of heat from the surface in the lowest few hundred millibars of the atmosphere. This dynamics gives West African precipitation enhanced sensitivity to surface conditions. The African easterly jet is associated with the divergence of moisture below the level of condensation, so a strong jet may be connected with low precipitation over the western Sahel.

This study provides the fundamental reason for the occurrence of the African easterly jet. We have identified the cause of the jet as the soil moisture distribution, but using this knowledge to further our understanding of the region’s observed variability and predictability requires a more complex setting, including two-way interactions between the surface and the atmosphere, topography, and realistic SST distributions.

Acknowledgments. The author thanks Brian Belcher for his assistance in running and analyzing GCM experiments, and Steve Colucci and John Chiang for helpful discussions. Figures were generated using the GrADS system. Research support is from the National Science Foundation through Grant ATM-9300311.

REFERENCES


Trenberth, K. E., 1992: Global analyses from ECMWF and atlas of 1000 to 10 mb circulation statistics. NCAR Tech. Note STR. [Available from NCAR, P.O. Box 3000, Boulder, CO 80307-3000.]

