Influence of Surface Processes over Africa on the Atlantic Marine ITCZ and South American Precipitation

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Abstract

Previous studies show that the climatological precipitation over South America, particularly the Nordeste region, is influenced by the presence of the African continent. Here the influence of African topography and surface wetness on the Atlantic marine ITCZ (AMI) and South American precipitation are investigated.

Cross-equatorial flow over the Atlantic Ocean introduced by north–south asymmetry in surface conditions over Africa shifts the AMI in the direction of the flow. African topography, for example, introduces an anomalous high over the southern Atlantic Ocean and a low to the north. This results in a northward migration of the AMI and dry conditions over the Nordeste region.

The implications of this process on variability are then studied by analyzing the response of the AMI to soil moisture anomalies over tropical Africa. Northerly flow induced by equatorially asymmetric perturbations in soil moisture over northern tropical Africa shifts the AMI southward, increasing the climatological precipitation over northeastern South America. Flow associated with an equatorially symmetric perturbation in soil moisture, however, has a very weak cross-equatorial component and very weak influence on the AMI and South American precipitation. The sensitivity of the AMI to soil moisture perturbations over certain regions of Africa can possibly improve the skill of prediction.

1. Introduction

Because of its economic and social impact on the population of the neighboring land regions, the dynamics of Atlantic marine ITCZ (AMI) merits the recent efforts to understand its various aspects. In addition to its direct dependence on the sea surface temperatures (SSTs), the AMI is also sensitive to remote influences like ENSO. The role played by the surface features of the adjacent land, however, has not received as much attention. Cook et al. (2004) examined the nature of intercontinental teleconnections between Africa and South America. In their model simulations, Africa influences South American precipitation more than South America does Africa. The strongest intercontinental forcing occurs during the austral summer (January) when the presence of Africa introduces a large (about 40%) suppression of rainfall over the Nordeste region of Brazil, and rainfall enhancements on the northern coast and over the South Atlantic convergence zone (SACZ). Heating within the West African monsoon system causes drying of about 1–2 mm day\(^{-1}\) along the northeastern coast of South America. As a natural extension of the above work, this study is aimed at addressing the following questions:

- What is the influence of African topography on the climatology of the Atlantic marine ITCZ and South American precipitation?
- How do variabilities in surface wetness over Africa influence the AMI?

Since the pioneering work of Charney and Eliassen (1949), many studies have investigated the influence of large-scale orography on atmospheric circulation. The role of African orography on the local climate was studied numerically by Semazzi (1980a,b). He conducted a series of simulations based on a primitive equation barotropic model to investigate the influence of large-scale orographic forcing on the climate of Africa. In these experiments, observed mean zonal flow was prescribed in the presence of African orography. His results show that African orography introduces an equatorial pressure trough and a pair of cyclones on either side of the equator over the Atlantic Ocean. Later, Se-
mazzi and Sun (1997) used the National Aeronautics and Space Administration (NASA) Goddard Earth Observing System (GEOS-1) GCM to show that the large-scale orography of Africa plays a major role in determining the climate of the Sahel and the Guinean regions during Northern Hemisphere summer. Downstream of the orography, their model produces a leeward low centered to the north of the Sahelian region. They also found that cyclonic circulation associated with this trough plays an important role in transporting moisture from the midsection of the Atlantic Ocean into the western parts of the Sahel region. The influence of orography on the position of the local ITCZ has also been studied. Sultan and Janicot (2003) suggested that orographically forced circulations could induce a northward relocation of the ITCZ over West Africa during summer due to a monsoon enhancement, and that the observed abrupt shift of the ITCZ over land could be associated with the presence of the Atlas–Ahggar Mountains to the north, and the development of the leeward trough that reinforces an active ITCZ through enhanced moist air advection into the continent. Ringler and Cook (1999) showed that interactions between mechanical and thermal forcing can be a two-way process. For example, condensational heating resulting from flow over topography is likely to cause significant changes to the circulation and this, in turn, can modify the mechanical forcing associated with topography. So a full description of the influence of orography on the position of the local ITCZ has also been studied. Sultan and Janicot (2003) suggested that orographically forced circulations could induce a northward relocation of the ITCZ over West Africa during summer due to a monsoon enhancement, and that the observed abrupt shift of the ITCZ over land could be associated with the presence of the Atlas–Ahggar Mountains to the north, and the development of the leeward trough that reinforces an active ITCZ through enhanced moist air advection into the continent. Ringler and Cook (1999) showed that interactions between mechanical and thermal forcing can be a two-way process. For example, condensational heating resulting from flow over topography is likely to cause significant changes to the circulation and this, in turn, can modify the mechanical forcing associated with topography. So a full description of the influence of orography on local climate should account for its combined mechanical and thermal effects.

The impact of local soil moisture on the climate of the Sahel region of Africa has also been an area of interest. For example, Douville (2002) used a GCM to show that the relationship between precipitation and soil moisture over tropical Africa is not only due to a simple recycling mechanism between evaporation and moisture convergence. Increases in evaporation over the Sahel region lead to precipitation anomalies to the southwest in association with the westward advection of water vapor by easterly waves. Soil moisture is known to be an important determinant of surface temperature through regulation of the relative magnitudes of the sensible and latent heat fluxes from the surface. Shinoda and Yamaguchi (2003) used observational data to compare temperature and rainfall during decades of wet (1950s) and dry (1980s) surface conditions over the African Sahel. Their results show that increases (decreases) in the accumulated precipitation leads to increases (decreases) in soil moisture with a time lag. This results in decreased (increased) sensible heat flux and daily maximum surface temperature and a simultaneous increase (decrease) in the latent heat flux. This kind of relationship between soil moisture and surface temperature is clearly observed only during the beginning 1–2 months of the dry season before the surface becomes desiccated. Since the soil moisture reservoir has memory longer than that of most other atmospheric processes, anomalous perturbations associated with soil moisture are likely to persist for a longer time.

A more recent effort involving 12 AGCMs [the Geophysical Fluid Dynamics Laboratory (GFDL) model among them] to identify regions of strongest influence of soil moisture on precipitation shows that northern tropical Africa is one of the “hot spots” where the impact of soil moisture on precipitation is very strong (Koster et al. 2004).

The emphasis of this study is to evaluate the role of African topography on the climatology of as well as to investigate the possible impact of variability on surface wetness over Africa on the AMI and South American precipitation. The approach involved is climate modeling using idealized boundary conditions. A description of the GCM simulations implemented is given in section 2. Analysis of the influence of African topography on the AMI and South American climate is presented in section 3 and their responses to circulations induced by equatorially symmetric and asymmetric prescribed weak anomalous soil moisture over tropical Africa as well as the implied link between the Nordeste region in January and that over Africa in an earlier month is assessed using 25 yr of Global Precipitation Climatology Project (GPCP; Huffman et al. 1996) precipitation data.

To gain further understanding of the underlying dynamics, the linear response of the tropical atmosphere to forcings of similar structure as those prescribed in the GCM experiments is considered. A simple analytical model under steady, prescribed heating and friction is presented. The results are compared with those of the GCM simulations. The overall conclusions drawn from this study are summarized in the last section.

2. Models and simulations

a. Model description and validation

A series of simulations with a version of the GFDL R30 14-level (Gordon and Stern 1982) atmospheric GCM is discussed. Simple physical parameterizations, with the bucket hydrology model at the surface and the moist convective adjustment scheme, are used. The model solves the primitive equations and an equation to predict water vapor mixing ratio distributions. Each simulation begins with an isothermal dry atmosphere at rest. The integration length is 2000 days with the first
200 days removed as a spinup period. Statistical significance tests for differences between results of a pair of experiments for any field are made using a Student’s t test for each grid point. A difference is then assumed to be statistically significant if its probability is 0.1 and below (i.e., 90% confidence level and above), else the difference is filtered out.

The experiments in this study are different from those described in Cook et al. (2004) in that a realistic distribution global land surface with topography as well as SST are prescribed here, while those in Cook et al. (2004) involve one or two continents and a zonally uniform SST. The control simulation has a realistic representation of land surface features on all continents, including climatological soil moisture from Mintz and Walker (1993) shown in Fig. 1a, surface albedo from Matthews (1984), and global topography. Realistic SSTs based on Shea et al. (1990) are also prescribed. A perpetual season format is used, with solar insolation and SSTs held fixed at January values. This approach allows one to efficiently produce a quasi-stationary climatology for 1 month with relatively short integration times. The 2000-day runs make the system “forget” its initial conditions. The results should be interpreted as solutions to a boundary value problem involving time-independent climatologies. The surface drag coefficient used in the sensible and latent heat flux parameterizations over land is greater than that over the ocean by a factor of 3. This simulation is used to validate the model’s ability to capture the observed climatology, and to compare with the Flat Africa experiment described below.

In the Flat Africa experiment, African topography (Fig. 1b) is removed while that of every other continent is retained. All other surface features, including surface albedo and soil moisture, are the same as in the control simulation. Comparison with the control simulation provides a basic understanding of the influence of African topography on the structure of the AMI and precipitation over tropical South America.

Two additional simulations include only the African continent, with the remainder of the globe covered with zonal uniform SSTs derived from Shea et al. (1990). In one of these, Flat Africa Alone, African topography is removed. A difference between these two simulations is used to isolate the wind and pressure fields associated with African topography.

The goal of another set of experiments is to evaluate the response of the AMI to typical anomalous circulations induced by variations in surface conditions over tropical Africa, and to evaluate the degree to which Africa can force precipitation variability over South America. To do this, soil moisture is selected as an independent variable and slightly perturbed about the climatology to change the local surface heat flux and ultimately the circulation.

In the asymmetric simulation, a localized soil moisture anomaly of Gaussian structure with an amplitude of +1 cm and latitudinal and longitudinal half-widths of 5° and 30°, respectively, is centered at 7°N and 25°E (Fig. 2a). In the symmetric simulation, a localized Gaussian soil moisture anomaly of amplitude of −1 cm with latitudinal and longitudinal half-widths of 5° and 30°, respectively, is located at 25°E over 0° latitude.
A brief summary of the experiments is given in Table 1. The geographical locations of the soil moisture anomalies are estimated based on the deviations of the monthly climatological soil moisture distribution over tropical Africa during the 1950s (wet) and 1980s (dry) from a 50-yr average obtained from National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis. Since the magnitudes of the perturbations are not very accurate (e.g., see Philippon and Fontaine (2002)), the prescribed perturbations are chosen to be smaller to make sure the subsequent linear analysis is justified.

An evaluation of the control experiment builds confidence in the model’s ability to represent the region’s large-scale climatology. January winds and geopotential heights at 866 hPa from the control simulation with both continents are displayed in Fig. 3a. The strong height gradient in the subtropical Southern Hemisphere is similar to that displayed in the NCEP–NCAR reanalysis (Fig. 3b). Both the African monsoon and the flow into the northeastern coast of South America are realistic. The South Atlantic high is well simulated and the flow across the southern Atlantic Ocean is captured. The comparison between the NCEP–NCAR reanalysis data and the GCM output is also good at higher levels. For example, Fig. 4 shows the control simulation result and reanalysis data at 250 hPa. The strong subtropical westerly flows in both hemispheres are well simulated. The Bolivian and southern African highs are reproduced in the simulation.

Figures 5a and 5b display January rainfall climatologies over Africa, South America, and the Atlantic Ocean obtained from two sets of observations. The South American monsoon is active, and a precipitation maxima is centered near 10°S in the Amazon basin with the South Atlantic convergence zone (SACZ) extending to its southeast. The SACZ is not very well defined in the Legates and Willmott (1990) gauge-based dataset (Fig. 5a) and the two datasets disagree on whether another precipitation maximum occurs just north of the equator (Fig. 5b). Note the strong zonal rainfall gradient between the Amazon region and the Nordeste to its east. Over southern Africa, the precipitation maximum is over the eastern half of the continent. The AMI is centered near 3°N in both observations, and is stronger over the western tropical Atlantic than in the east.

Figure 5c shows the January precipitation climatology from the control GCM simulation. Primary features of the observed South American precipitation climatology are represented, including the Amazon maximum and the SACZ. Rainfall over the high Andes is too strong in the model because smoothing of the topography, which is necessary to avoid spectral ringing, lowers the elevation. As a result, specific humidity levels near the surface are too high (Lenters and Cook 1995). The western portion of the AMI is well simulated, but the rainfall does not extend completely across the continent.

![Table 1. GCM experiments](image)

<table>
<thead>
<tr>
<th>Name</th>
<th>Soil moisture</th>
<th>Continents</th>
<th>Topography</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>Climatological</td>
<td>All</td>
<td>Full topography</td>
</tr>
<tr>
<td>Flat Africa</td>
<td>Climatological</td>
<td>All</td>
<td>No African topography</td>
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<tr>
<td>Africa Alone</td>
<td>Climatological</td>
<td>Africa only</td>
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<td>Flat Africa Alone</td>
<td>Climatological</td>
<td>Africa only</td>
<td>No African topography</td>
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<tr>
<td>Asymmetric</td>
<td>Wet northern equatorial Africa</td>
<td>All</td>
<td>Full topography</td>
</tr>
<tr>
<td>Symmetric</td>
<td>Dry equatorial Africa</td>
<td>All</td>
<td>Full topography</td>
</tr>
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(Fig. 2b).
the equatorial Atlantic to West Africa. The latitude of maximum precipitation is very slightly farther south than in the observations, at 2°N. Maximum precipitation rates over Africa are a little farther south than the Amazon maximum in both the observations and the model, but the model places the rainfall maximum too far west (probably due to poor resolution of the southern African plateau topography).

b. Linear tropical model

Simplified linear models are useful in diagnosing the first-order structure of circulations obtained from GCM simulations. A two-dimensional dynamical model of a hydrostatic, steady-state perturbation of a stratified atmosphere on an equatorial β plane driven by differential heating and controlled by friction can be written as (Moura and Shukla 1981)

\[ \epsilon u - \gamma v = -\frac{\partial \phi}{\partial x}, \]  
\[ \epsilon v + \gamma u = -\frac{\partial \phi}{\partial y}, \]  
\[ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p} = 0, \text{ and} \]  
\[ \omega = -\frac{\rho \hat{q}}{S(p)}. \]  

Fig. 3. (a) Jan wind and geopotential height at 866 hPa from control and (b) Jan wind and geopotential height at 850 hPa from the NCEP–NCAR reanalysis. The contour interval is 30 gpm and the vector scale is in m s\(^{-1}\). Dotted contours of 1585 gpm for (a) and 1505 for (b) are also included.
where $u$, $v$, and $\omega$ are the $x$, $y$, and $p$ components of velocity; $\phi$ is the geopotential height, and $\epsilon$, $S(p)$, and $\dot{q}$ are the friction coefficient, stability parameter, and the heating rate, respectively. All the variables are appropriately made nondimensional.

In the Tropics, horizontal density and temperature gradients are very small. Consequently, on large space scales, the dominant balance in the thermodynamic equation is between diabatic heating and adiabatic cooling. In other words, tropical heating is primarily balanced by vertical motion, and horizontal temperature advection is small. A detailed evaluation of a similar approximation is given by Sobel et al. (2001). Having this in mind, the horizontal structure of the diabatic heating and that of the ascent–descent of air over the region are assumed to be equivalent. As will be shown later, this approximation is quite good in regions close to the equator but could result in significant error away from the equator. The details of the solution to (1)–(4) are given in the appendix.

The underlying assumption in this linear analysis is that perturbations in soil moisture content, surface fluxes, surface temperature, and diabatic heating are small. Also, what is usually observed in the atmosphere and nonlinear models is that forced perturbations of the low-level wind field influence the moisture distribution and surface heat exchange, which ultimately modifies the forcing field itself. These feedback processes are not included in these simple linear calculations.
3. Results

a. Influence of circulations associated with topography

To understand the structure of the orographic forcing field, and the first-order dynamical response, differences between the Africa Alone and Flat Africa Alone simulations are considered. Orographic forcing can be mechanical, in association with obstruction of the flow, or thermal. Thermal forcing consists of largely low-level sensible heating perturbations and midtropospheric perturbations of the diabatic heating field. Figure 6 shows the difference in the column-average condensational heating associated with African topography. This orographic thermal forcing is comparable to the climatological heating from the control simulation in this region, which is 2–6 K day\(^{-1}\).

Differences in pressure and wind fields between Africa Alone and Flat Africa Alone are shown in Fig. 7a. An anomalous low is centered over the topography of southern Africa (and the region of anomalous diabatic heating). The response to this heating over southern Africa introduces a region of high geopotential heights over the southern Atlantic Ocean. This is similar to the observed anomalous high over the eastern South Pacific in association with the South American monsoon heating, and over the eastern North Atlantic due to the Asian monsoon (Rodwell and Hoskins 2001). The simulated weak low in the northern subtropics is to the west of the topography, suggesting that it is primarily a response to mechanical forcing. A typical orography located at Northern Hemisphere subtropics introduces a low to its immediate west in the absence of diabatic heating (Hoskins and Karoly 1981).

Figure 7b shows differences in pressure and wind fields between control and Flat Africa. Structures simi-
lar to those in Fig. 7a are observed, but the Southern Hemisphere geopotential height anomaly is weaker in the presence of the realistic distribution of continents and SST. The low pressure anomaly off the north coast of West Africa is stronger. Despite these differences, the anomalous winds are similar throughout the region. This similarity justifies the use of the more realistic simulation to examine the influence of African topography on the AMI and South America.

In both cases shown in Fig. 7, the presence of African topography is seen to introduce an anomalous meridional pressure gradient across the equator centered near 20°W. As a result, the irrotational part of the anomalous wind, displayed in Fig. 8a, has a significant southerly component crossing the equator. The center
of convergence (ascent) is located on the western coast of central Africa, and the divergence center (descent) is over the northeastern tip of South America (the Nordeste region of Brazil). So the overall effect is the introduction of an anomalous circulation over the Atlantic Ocean that is essentially a Walker circulation, but with the down branch shifted about 8° south of the equator.

Figure 8b shows the nondivergent wind difference associated with African topography. A pair of cyclonic anomalies are generated over the eastern Atlantic in both hemispheres, similar to the classical Gill–Matsuno response to tropical heating (e.g., Gill 1980), but here the response is to orography. These cyclones introduce a strong westerly wind anomaly over the equatorial Atlantic.

The precipitation anomaly associated with these circulation features and, therefore, the presence of African topography is displayed in Fig. 9. (Note that in the GCM simulations, precipitation and circulation feedback are included.) The influence of African topography is particularly large over the northeastern parts of

![Fig. 8. (a) Irrotational wind (m s⁻¹) and velocity potential for control minus Flat Africa. (b) Nondivergent wind and for control minus Flat Africa at 866 hPa. The vector scale is in m s⁻¹.](image-url)
South America and the western equatorial Atlantic where the rainfall is up to 6 mm day\(^{-1}\) (about 80\%) lower. This rainfall deficit is consistent with the anomalous circulation shown Fig. 8a. The AMI is shifted to the north by about 5° of latitude, resulting in a precipitation increase over the northern edge of the continent. Comparison of Fig. 8a of Cook et al. (2004) with Fig. 9 shows that the precipitation over northeastern South America attributed to African topography is comparable to that associated with the presence of the continent as a whole. There is little influence from Africa over the southern and southwestern parts of the continent. As shown by Cook et al. (2004), the response is particularly strong over land because of feedbacks from the land surface.

Precipitation deviations can be related to contributions from column moisture convergence, moisture advection, orographic uplift, transient eddy activity, and evaporation through conservation of water in the atmospheric column. Following Lenters and Cook (1995), each component of the column moisture budget was calculated. Comparison of the anomalous rainfall (Fig. 9a) and column moisture convergence fields (Fig. 9b) shows that differences in the latter account for most of the rainfall deviation over the western Atlantic and South America. Since most of the atmospheric mois-
ture is at low levels, this implies that anomalous structure in the low-level wind fields is closely associated with the precipitation anomaly shown in Fig. 9a. Over the Gulf of Guinea, in association with the strong moisture gradients of the coastal region, moisture advection is important for the column moisture budget. As seen in Fig. 9c, positive moisture advection balances the column moisture divergence so that the overall precipitation response is small.

In addition to modifying the meridional position of the AMI, the African topography in these simulations impacts the surface heat balance in the eastern equatorial Atlantic through its influence on the low-level equatorial easterly flow. The equatorial westerly components of the pair of the cyclones significantly weaken the easterly trade winds across the breadth of the Atlantic Ocean. This can have important implications for the strength of the AMI because the convection is ultimately driven by sensible and latent heat transfers from the surface, which in turn are sensitive to the magnitude of the surface winds. On large time scales the ocean temperature is likely to respond to the changes in surface heat fluxes as well as to the decrease in upwelling due to weaker easterly flow. These processes are beyond the scope of this study.

b. Influence of circulations associated with perturbations in soil moisture

Prescribing wet surface conditions (shown in Fig. 2a) over northern tropical Africa introduces the precipitation anomaly shown in Fig. 10a. A southward shift of precipitation is apparent from the dipolar structure over equatorial Africa and South America. The Nordeste region gains an average of about 3 mm day$^{-1}$ of precipitation (about 40% of the climatological value), while northern parts of the continent get about 2 mm day$^{-1}$ drier. Moisture budget analysis shows that, as was the case for the precipitation perturbations associ-
ated with topography discussed above, column moisture convergence (Fig. 10b) accounts for most of the precipitation anomaly and is closely related to the low-level wind convergence. The AMI in the central and western Atlantic is shifted toward the summer hemisphere and slightly strengthened, also in association with perturbation of the low-level wind convergence (not shown).

To understand the physical processes involved in this response, the forcing function over Africa is identified by examining the local effects of changes in surface wetness on the diabatic heating. The soil moisture perturbation changes the surface heat budget. Figure 11 shows differences in each component of the surface heat budget. Wet surface conditions reduce the sensible heat and net upward longwave radiation fluxes (Figs. 11a and 11b, respectively), and increase the latent heat flux (Fig. 11c). These differences are accompanied by a cooling of the surface (shown in Fig. 11d) because the enhanced availability of moisture favors evaporation.

Fig. 11. (a) Sensible heat flux (W m\(^{-2}\)), (b) net upward longwave flux (W m\(^{-2}\)), (c) latent heat flux (W m\(^{-2}\)), and (d) surface temperature (K) differences (asymmetric minus control). Contour intervals are 15 W m\(^{-2}\) for (a)–(c), and 2 K for (d).
and heat transfer through the latent heat flux. Figure 12a shows the full wind and streamfunction differences at 866 hPa associated with the wet surface conditions imposed over northern equatorial Africa (asymmetric minus control). In the Northern (winter) Hemisphere, there is a substantial acceleration of the tropical easterly flow over West Africa and the eastern Atlantic. Close to the equator, the flow anomaly has a strong northeasterly component over the Gulf of Guinea. The strongest differences in the wind occur over South America, where a pronounced cyclonic flow is associated with enhanced precipitation. This precipitation difference is associated with a southward shift of the ITCZ (Fig. 10), and amplification by land surface feedbacks (Cook et al. 2004).

The irrotational part of the flow is shown in Fig. 12b, along with velocity potential contours. Cooling over Africa introduced local anomalous divergence and descent (not shown). Of particular relevance for the positioning of the AMI is the resulting northerly component to the flow that extends westward over the tropical Atlantic. As in the case for orographic forcing discussed above, the large-scale Walker circulation is perturbed with anomalous convergence in the Southern Hemi-
sphere, but with divergence on the equator. Further insight is provided by an analysis using the linear model (described in section 2 and the appendix). When forcing with similar structure to that of the asymmetric simulation is imposed in the linear framework, the anomalous full and irrotational wind fields shown in Figs. 13a and 13b, respectively, result. Similarities between the linear and GCM solutions break down north of about 15°N where temperature advection becomes important and the weak temperature gradient approximation (4) is not valid. Elsewhere, however, the general features of the linear model solution agree with the low-level wind field obtained from the GCM simulations (Fig. 12). In particular, the linear model captures the northerly flow over the equatorial region that is responsible for the meridional migration of the ITCZ in the GCM. Since divergence is prescribed indirectly through the specification of the forcing in the linear model, land surface feedback is absent and, therefore, the localized convergence anomaly seen over South America in the GCM (Fig. 12b) does not occur.

The above results are consistent with the westward control hypothesis (Xie 1996), according to which, because long antisymmetric Rossby modes can propagate...
only westward, the position of a marine ITCZ is controlled by the eastern continent. Antisymmetric modes generated by a dipolar distribution of anomalous SST over the Atlantic Ocean have also been known to have a similar effect on the AMI and precipitation over the Nordeste region (see Moura and Shukla 1981; Hastenrath and Giescher 1993 and references therein). In their observational studies, Gu and Zhang (2001) used outgoing longwave radiation (OLR) data to investigate the relative roles of waves in forming the structure of the ITCZ. They found that the latter is composed mainly of nonpropagating random shallow and deep convective clouds as well as zonally propagating synoptic-scale organized deep convective systems. They estimated that westward-propagating waves account for 25%–40% of the total deep convective clouds and 10%–15% of the ITCZ cloudiness. Contributions from eastward-propagating waves were found to be much smaller. The above results suggest that the primary mode of intercontinental interactions involves antisymmetric Rossby waves propagating westward across the Atlantic. This could explain the relatively weak influence of South America on precipitation over tropical Africa discussed in Cook et al. (2004).

Prescribing the equatorially symmetric soil moisture anomaly shown in Fig. 2b introduces a temperature increase of about 2 K locally because the reduced surface moisture favors sensible heat transfer over the latent heat flux. Because the circulation anomaly in this case reflects the equatorial symmetry of the forcing, statistically significant cross-equatorial flow and precipitation perturbation are virtually absent.

The experiments evolved from an initial pair of experiments designed to investigate the response of South America to wet and dry conditions over Africa centered at the equator, which showed that the response arises from the strong sensitivity to soil moisture perturbation over the regions between the Sahara and the equator. So the experiment involving wet surface conditions being symmetric about the equator also results in a response similar to that of the asymmetric case discussed herein because the system is most sensitive to the northern half of the soil moisture anomaly. The reason for this is described in some detail by Koster et al. (2004), which also shows that precipitation is most sensitive to soil moisture in the transition regions between wet and dry areas. Evaporation in wet regions is less sensitive to soil moisture perturbations. In dry areas, while it is sensitive to soil moisture, the response is generally weak because of the lack of moisture. The soil moisture anomaly prescribed in the asymmetric simulation is more or less in a transition region. The symmetric experiment is on a wet region and its influence on precipitation over both local and remote regions is smaller.

The correlation between areal average precipitation over northeastern South America (0°–10°S, 45°–30°E) in January and precipitation over Africa during December is shown in Fig. 14. The analysis is done on 25 yr of GPCP precipitation data. Precipitation over northeastern South America has some positive correlation at the 90% significance level with that over northern tropical Africa and is negatively correlated with that of equatorial Africa; this observation is consistent with the model results, which suggest that precipitation over the Nordeste region of South America could be modulated by wetness over northern tropical Africa. Further studies are necessary to evaluate the potential of soil moisture to serve as a link between precipitation over Africa and that over South America at some later time.

4. Discussion

In addition to its strong dependence on local SSTs, the AMI is sensitive to remote influences like the eastern equatorial Pacific Ocean. The role played by the physical features of, and variabilities on the surfaces of the surrounding continents, have not received as much attention. Cook et al. (2004) showed that the hydrodynamical response to forcing associated with the existence of Africa significantly modifies the precipitation climatology in parts of South America. In the first part of this study the role of African topography on the climatology of the AMI and South American precipitation is investigated. Then the impacts of variations in surface wetness over Africa on AMI and South
America are evaluated. A series of atmospheric GCM experiments with idealized boundary conditions are presented. To efficiently produce climatological values, the simulations are performed in the perpetual January mode with solar insolation and SSTs held at fixed January values. The influence of the presence of Africa on South America has been shown to be particularly strong during January (Cook et al. 2004).

African topography introduces strong divergence over northeastern South America by shifting the AMI northward. This leads to an 8 mm day$^{-1}$ (50%) drop in precipitation in the Nordeste region and up to a 6 mm day$^{-1}$ rise to the north. This response is due to the fact that the response to topography is different in the winter and summer hemispheres. In the winter hemisphere, the topographic forcing is primarily mechanical (obstruction of the flow) and introduces a weak low to the west. In the summer hemisphere, there is strong coupling with the diabatic heating field and the low is located on top of the orography and the South Atlantic high is enhanced. This asymmetry leads to north–south geopotential height gradients and a southerly flow across the equator that shifts the AMI northward, leaving the equatorial western Atlantic dry.

Wet surface conditions over northern tropical Africa reduce the sensible heat transfer in favor of evaporation. This reduces the local surface temperature and induces low-level divergence and descent. This results in a cross-equatorial geopotential height gradient to the immediate west, northerly flow across the equator, and a southward shift of the AMI. A dipolar precipitation perturbation that extends into the equatorial western Atlantic Ocean and northeastern South America accompanies these circulation anomalies, increasing rainfall over the Nordeste region by 3 mm day$^{-1}$ (about 40% of the climatological rainfall).

Dry surface conditions centered over equatorial Africa increase the local surface temperature and introduce an anomalous forcing that is symmetric about the equator. The change in cross equatorial flow and precipitation response over the AMI and South America and the Atlantic Ocean is found to be small.

The possibility of using December precipitation over Africa, which contributes to soil moisture, with January precipitation over northeastern South America is assessed using observational data. The results suggest future research should further investigate possible causal relationships between precipitation over northern tropical Africa and the Nordeste region of South America. In summary, the primary influence of African topography on South America comes through the equatorial asymmetry of the associated circulations and its influence on the meridional position of the Atlantic ITCZ. Conversely, soil moisture anomalies that induce cross-equatorial flow over the Atlantic shift the AMI in the direction of the flow. This study concentrates on how large-scale variations over Africa influence South America and the AMI.

The experiments discussed here are simplified and can only provide a basic idea of the processes involved with this relationship. The impact of the degradation of vegetation across Africa on other regions of the globe is also an important practical problem not addressed in this study. While this study provides some basic mechanisms of the influence of circulations associated with the African land surface on the AMI and South America precipitation, the simulations have some notable shortcomings that limit their accuracy: their perpetual-season format, fixed SSTs, soil moisture, and albedo, as well as the relatively low resolution of the GCM. Intercontinental interactions also include smaller-scale disturbances (e.g., African easterly waves and hurricanes over the eastern United States), so further analysis should involve investigation of the multiscale and two-way nature of the interactions of circulations with land and ocean surfaces. Analysis of time-dependent simulations with a higher-resolution model, realistic parameterizations of air–sea–land interactions, and boundary conditions is crucial to this end.

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APPENDIX

Linear Model Solution

To solve the system of Eqs. (1)–(4), the prescribed diabatic heating is assumed to be of the form

$$q = X(x)Y(y)P(p), \quad (A1)$$

where $X(x)$, $Y(y)$, and $P(p)$ in (A1) are the zonal, meridional, and vertical structures, respectively, of the prescribed heating field, which is assumed to be independent of the pressure and wind field. Here, $P(p)$ is prescribed to be zero at the surface and is an increasing function of altitude in the lower part of the troposphere, so that the boundary condition of impermeability of the surface can be satisfied. The constant Rayleigh friction $\varepsilon$ is made nondimensional by $\beta L$; east-
ward and northward distances, $x$ and $y$, by $L$; and the eastward and northward velocities, $u$ and $v$, by $V$. The vertical velocity, $\omega$, and the geopotential, $\phi$, are nondimensionalized with $(PV/L)$ and $L^2V$, respectively. The diabatic heating is made nondimensional by $(S, l, k)$ and the static stability by a typical value of $H_s, N_s$, where $H_s$ is the scale height and $N_s$ is the Brunt–Veisilä frequency; $\beta = 2(\Omega/a)(=2.28 \times 10^{-11} \text{ms}^{-1})$, $V = 10 \text{m s}^{-1}$. The length scale $L$ is of the order of the equatorial Rossby radius and is taken to be equal to 1000 km.

Combining (1)–(4) gives

$$
\epsilon \frac{\partial^2 u}{\partial y^2} + \epsilon \frac{\partial^2 v}{\partial x^2} + \frac{\partial u}{\partial x} = \left( \epsilon \frac{\partial}{\partial y} - y \frac{\partial}{\partial x} \right) \frac{\partial}{\partial p} \left( \frac{p}{S(p)} \right).
$$

(A2)

Equation (A2) admits a solution of the form

$$
\left[ \frac{d}{dp} \frac{p}{S(p)} \right] \sum_k v_k(y)e^{ikx} \text{ and (A3),}
$$

substituting (A3) into (A2),

$$
\epsilon \frac{d^2}{dy^2} v_k(y) - (k^2e - ike)v_k(y) = \epsilon q_k \frac{d}{dy} Y(y) - ikq_k \bar{y} Y(y).
$$

(A4)

Here $q_k$ is the $k$th Fourier coefficient of the $x$ structure of the heating given by

$$
q_k = \frac{\int_{-S}^{S} X(x)e^{-ikx} \, dx}{2S},
$$

(A5)

where $S \sim 40L$ is the nondimensionalized circumference of the earth. To obtain a solution that satisfies the boundary conditions that $v_k(y)$ should vanish at $y = \pm \infty$, $v_k(y)$ can be expanded as

$$
v_k(y) = \sum_{n=0}^{n=\infty} a_n \psi_n(y),
$$

(A6)

In (A6) $\psi_n(y) = e^{-1/2y^2}H_n(y)$ is the $n$th-order Hermite function whose properties are

$$
\int_{-\infty}^{\infty} \psi_n \psi_m \, dy = \pi^{1/2} 2^n m! \delta_{nm},
$$

(A7)

where $\delta_{nm}$ is the Kronecker delta, and

$$
\frac{d^2}{dy^2} \psi_n(y) = n(n-1)\psi_{n-2} - (n + 1/2)\psi_n + 1/4\psi_{n+2}.
$$

(A8)

Substituting (A6) into (A4) and using (A8) gives

$$
\left( -k^2 + \frac{ik}{\epsilon} - n + 1/2 \right) a_n + 1/4a_{n-2} + (n + 1)(n + 2)a_{n+2} = c_n - \frac{ikd_n}{\epsilon},
$$

(A9)

where

$$
c_n = \frac{\int_{-\infty}^{\infty} \psi_n \frac{d}{dy} Y(y) \, dy}{2^n n! \pi^{1/2}} \text{ and (A10),}
$$

$$
d_n = \frac{\int_{-\infty}^{\infty} \psi_n Y(y) \, dy}{2^n n! \pi^{1/2}}.
$$

(A11)

After substituting the forcing structure $Y(y)$ into (A10) and (A11), (A9) results in a pair of tridiagonal matrix systems, and $v_k(y)$ is obtained by solving the pair of matrix equations by inversion. Combining (3) and (4) and using (A3) gives

$$
iku_k + \frac{\partial}{\partial y} v_k(y) + Y(y)q_k = 0.
$$

(A12)

After solving for $u_k$, the Fourier components of (1) can be used to obtain $\psi_k$:

$$
\epsilon u_k - v_k = -ik\psi_k.
$$

(A13)

Summing over all $k$, the final solution is obtained.

REFERENCES


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