Mechanisms by Which Surface Drying Perturbs Tropical Precipitation Fields

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ABSTRACT

The observed precipitation climatology in austral summer has a pronounced longitudinal gradient across Africa and South America. A low-resolution general circulation model (GCM) with a simple continent centered on the equator is used to understand how the presence of the land surface generates this gradient, and the role of surface wetness in determining its magnitude. In the model, precipitation is enhanced on the east coast of the continent in the summer hemisphere tropics with magnitude and location independent of surface wetness. Precipitation rates are lower in the continental interior and in the west as the surface becomes drier, resulting in a longitudinal precipitation gradient that is similar to observations.

Modeled low-level moisture convergence and wind convergence anomalies mimic the precipitation anomalies over all but the driest land surfaces. A linearized primitive equation model is used to identify the physical mechanisms responsible for the GCM's dynamical response. Dry convection and condensational heating force most of the anomalous convergence over the land surface in the GCM, with sensible heating and transient eddies playing more minor roles. At the latitude of the intertropical convergence zone (ITCZ), dry convection drives anomalous convergence at low levels, and this convergence is larger over drier (warmer) surfaces. Anomalous divergence develops in response to decreased condensational heating below 680 mb. The dependence on surface wetness arises because the relative strength of these opposing responses depends on the degree of warming over the land surface.

Low-level convergence over the eastern portion of the land surface in the model is forced by condensational heating in the middle and upper troposphere. Here, diabatic heating is balanced by adiabatic cooling, and the positive vertical velocities induce convergence below 830 mb by continuity. The magnitude of the response is largely independent of land surface drying and warming. The longitudinal precipitation gradient develops when even moderate surface drying affects precipitation in the continental interior and west, but not in the east.

1. Introduction

In austral summer, the large-scale precipitation field in the tropics is perturbed in the same way over Africa and South America. As seen in Fig. 1a, rainfall maxima occur over the eastern portions of these continents, and minima occur in the west. This results in precipitation rates that are up to an order of magnitude larger in the east. The signal during boreal summer is not as consistent (Fig. 1b); the shape of the African continent in the Northern Hemisphere admits an important source of moisture from the south, and precipitation is enhanced in the west near the Gulf of Guinea (Cadet and Nnoli 1987). Also, South America does not extend very far into the Northern Hemisphere, so there is a strong moisture source to the north.

Several features of land surfaces have been associated with perturbing precipitation distributions in climate models and, presumably, in the real world. These features include orography, surface albedo, surface roughness, and surface wetness and water retention properties [see the review by Nicholson (1988)]. Perhaps the first step in building an understanding of how this influence arises is to study the atmosphere's response to anomalous heating at the surface without the complications of the hydrological cycle. A range of atmospheric models have been successful in providing insights into this problem (e.g., Gill 1980; Simmons 1982; Ting and Held 1990).

The process that distinguishes SST anomalies from ground temperature anomalies is the relationship between ground temperature and potential evaporation. The disabling of evaporation as a surface cooling mechanism that can accompany drying affects the vertical distribution of diabatic heating throughout the troposphere as well as the heating at the lower boundary. The response of the atmosphere, in turn, affects the surface heat budget and the surface wetness through modifications of the precipitation, low-level wind, and temperature fields. Modeling studies by Walker and Rowntree (1977), Sud and Fennessy (1984), and Rind (1982) highlight the important influence of land surface wetness on precipitation. When we try to understand the source of this influence in depth, however, a classic problem arises: the complexity required to generate reasonably realistic simulations in models often pre-
Fig. 1. Observed precipitation from Legates and Willmott (1990) on the R15 GCM's transform grid for (a) DJF and (b) JJA. Contour intervals are 0.5, 1, 2, 4, 8, and 12 mm day\(^{-1}\). Values greater than 4 mm day\(^{-1}\) are shaded.

includes a clean analysis of cause and effect. With a GCM resolving even 2 or 3 degrees of latitude, one cannot prescribe a realistic distribution of land surface characteristics without introducing variations at the smallest resolved scale. Trying to diagnose processes in GCMs at these scales introduces uncertainties that are compounded when competing processes are operating and integration times are short.

Another approach to the problem is to use simple boundary conditions in a GCM, choosing the boundary conditions and designing integrations to highlight certain processes and relationships, and avoiding problems associated with the practical limitations of integration times and model resolution. By imposing simple boundary conditions in a GCM, the full complexity of the atmospheric dynamics and physics is retained but the sources of structure in the predicted fields are limited. In this way relationships between a given surface characteristic and the atmospheric response can be clearly established. Experiments can be designed to yield statistically significant results without extremely long integration times by focusing on a subset of climate processes. Then, having established an understanding of the basic processes operating in the simple situation, the same model can be used with increasingly complex boundary conditions to build up a physical understanding of the more realistic system.

To complement more realistic modeling studies, this work sacrifices some realism to study the effects of surface wetness in isolation by using multiple integrations of a low-resolution general circulation model (GCM) with simple boundary conditions. This approach to studying land surface/atmosphere interactions in the tropics and subtropics is used by Cook and Gnanadesikan (1991), hereafter referred to as CG. The focus is on understanding how the atmospheric dynamics responds to modifications of the diabatic heating fields due to the presence of land surfaces with various surface wetness values, and the implications of that response for the continental precipitation distribution. Geographical distributions of orography, continentality, surface albedo, and surface roughness are eliminated from the model boundary conditions and replaced with a simplified surface to isolate the GCM's response and sensitivity to various surface wetness prescriptions. A steady-state, linear, primitive equation model is used to further explore the physical mechanisms responsible for the GCM's response.

The GCM and the linear model are described in section 2; GCM experiments and their results are described and compared with observations in section 3. Analysis of an experiment with intermediate surface wetness is presented in section 4, followed by results for wetter and drier surfaces in sections 5 and 6, respectively. Section 7 contains conclusions.

2. Model descriptions

Two models are used to investigate the dependence of the tropical precipitation field on surface wetness. One model is a GCM, with a simple continental surface and zonally and hemispherically symmetric atmospheric forcing similar to CG. The other model is a three-dimensional stationary wave model with atmospheric dynamics that is essentially a linear version of the GCM's governing equations. The GCM is used in a series of integrations to establish a clean relationship between surface wetness and precipitation. The linear model is used to further understand how the zonal asymmetries in the GCM climatologies are generated. Each model is described below.

a. General circulation model

The GCM is a version of the atmospheric model developed and maintained by the Climate Dynamics group at NOAA's Geophysical Fluid Dynamics Laboratory. The GCM has nine vertical levels on sigma surfaces and a rhomboidal-15 truncation that is equivalent to a horizontal grid resolution of approximately 7.5\(^{\circ}\) longitude and 4.5\(^{\circ}\) latitude. The reader is referred to Manabe (1969) for a general description of this model. Here the description will focus on the treatment of processes that are most relevant to this particular study since the purpose is not to produce the most realistic global climate simulation, but to focus on a subset of processes in a particular region.

Although the fully nonlinear, nonsteady treatment of the atmospheric dynamics and physics of the GCM
is retained, a number of simplifications of the boundary conditions are specified. Clouds are fixed, and the same in both hemispheres; sea ice is not allowed to form; sea surface temperatures are fixed and zonally uniform; prescribed ozone concentrations are zonally uniform and hemispherically symmetric; land/sea distributions are simplified (see next section).

The explicit treatment of the response of the atmospheric dynamics to a given three-dimensional diabatic heating field through the primitive equations is one of the strengths of this class of models. However, calculating the form of the diabatic heating—the forcing for the dynamical response—implies the parameterization of processes that are not explicitly treated. The treatment of the following is of particular concern for this study: 1) relationship between surface wetness and ground temperature; 2) connection between surface features and atmospheric diabatic heating fields. These are discussed below.

Surface wetness affects ground temperature by modifying the surface heat budget. The degree of surface wetness is expressed in terms of the amount of water, in centimeters, held in soil that has a 15-cm field capacity according to a simple "bucket" formulation. The surface wetness is prescribed in all of the GCM experiments, so there is no adjustment of this field when ground temperature or precipitation rates change. The connection between the surface wetness and ground temperature is accomplished by a surface heat balance calculation in which the land surface is assumed to have zero heat capacity (see CG for a summary). In this calculation, the influence of surface wetness on the heat budget is modified by a restriction of evaporation for surfaces with wetness values below 11.25 cm. The purpose of this parameterization is to represent the tendency of vegetation to decrease transpiration in the face of dry conditions (stomatal resistance). Milly (1992) presents the history and evaluates the accuracy of the bucket parameterization.

Establishing the relationship between surface conditions and the atmospheric diabatic heating fields involves a number of processes that are not explicitly treated. Of particular relevance for this study are the dry-adiabatic adjustment (dry convection) and the calculation of condensational heating rates.

Dry convection is the model's adjustment of the temperature profile to the adiabatic lapse rate if the solution of the governing equations results in a dry-adiabatic unstable column. Pairs of adjacent layers are tested, and if the dry-adiabatic lapse rate is exceeded, heat is removed from the lower layer and deposited in the upper layer to bring the lapse rate to the dry adiabat. Iteration over the atmosphere proceeds until the entire column is stabilized.

Condensational heating occurs in the model atmosphere due to two processes—moist-adiabatic instability and large-scale condensation. The steps in these calculations are detailed in CG. In essence, the moist convection adjustment corrects temperatures to the moist-adiabatic lapse rate by condensing water. To improve model efficiency, this process is done noniteratively so it is possible for a region to be "destabilized" by the adjustment of adjacent layers. When this adjustment is complete, any grid points that did not participate in the moist adjustment but are still saturated are heated by an amount representing the condensation of enough water vapor to bring the point to saturation. Water that condenses by either process is assumed to contribute directly to precipitation, with no opportunity for reevaporation as it falls through the atmosphere. (Tests performed in conjunction with these experiments show that the order of operation is immaterial, i.e., the climatology does not change significantly if large-scale condensation precedes the moist-adiabatic adjustment. Also, the computer time-saving device of not iterating to stabilize the entire atmospheric column each time step does not change the solution, at least for these simple experiments.)

Weaknesses in the moist-convective adjustment parameterization are well documented (e.g., Frank 1983). A misrepresentation of scales is inherent in the parameterization, since moist convection in nature occurs on scales smaller than those resolved by the model. Meehl and Albrecht (1991), along with others, point out that a model with this moist-convective parameterization will have a tropical troposphere that is too cold and dry, and they suggest additional ventilation of heat and moisture from the boundary layer to compensate. Despite these limitations, however, it is not clear that more complex parameterizations result in a better simulation of climatological precipitation, temperature, and water vapor distributions on large space scales. Also, as discussed later, the observed feature of the tropical precipitation distribution that we are studying is captured here and in other models using this parameterization. Using the moist-convective adjustment parameterization for this particular application is appealing since it is fundamentally driven by the vertical temperature structure of the atmosphere and, as such, is directly and relatively simply related to the physical mechanisms highlighted in this study.

b. Steady-state linear model

The linear model, described in Ting and Held (1990), takes as its governing equations a linearization of the GCM's governing equations. It is a purely dynamical model, with no prognostic equation for water vapor and no surface heat budget calculation. It is a steady-state model, requiring no integration in time, with the governing equations solved by matrix inversion. The model solves for the deviations from zonal means for the horizontal motion and temperature fields on each of nine sigma levels that are the same as the GCM's sigma levels. The spectral treatment of horizontal dependence is the same as in the GCM, with the same R15 truncation.
Input to the linear model includes zonal mean fields to linearize about and three-dimensional forcing functions to generate zonal asymmetries. All of these fields are from the GCM climatologies described above. The forcing functions can represent the effects of topography, latent heating, radiative heating, sensible heating, and the cumulative effects of transient eddies on the time-mean climate. The total forcing may be one of these, or any combination. For example, Cook and Held (1992) use the linear model with the GCM to isolate and analyze the response to orography in middle latitudes.

Two criteria must be met before we can declare that the linear model is a good tool for analyzing the GCM response. First, since the linear model does not have a prognostic equation for water vapor, the distribution of water vapor mixing ratio should not be a major source of structure for the total column anomalous moisture convergence. This is equivalent to requiring that the anomalous moisture convergence distribution be similar to the anomalous wind convergence distribution in the GCM. The second criterion is that the linear model solution, generated by linearizing about the GCM’s perturbed mean fields and forced with all possible sources of zonal asymmetry, reproduce the GCM’s perturbations with sufficient accuracy to suggest that the dynamical response in the GCM is essentially linear. The degree to which these criteria are met must be examined before the linear model can be applied.

When applicable, the linear model is useful for identifying the processes responsible for the atmosphere’s response to the surface wetness field. Even with simple GCM experiments, the effects of each of the modified diabatic heating fields on the dynamics cannot be cleanly separated. The linear model cannot make the connection between the dynamics and the precipitation field, but it can demonstrate exactly how convergence is forced in the GCM given the diabatic heating perturbation if the above criteria are satisfied.

3. GCM experiments

The GCM is used in a perpetual season mode, with a winter hemisphere and a summer hemisphere. Day length, solar zenith angle, and SSTs are used to impose solstice conditions. Diurnal variation is not included, so the incoming solar radiation is the same at a given grid point each time step. Zonally uniform SSTs are prescribed to give values similar to the observed average ocean temperatures, with a temperature maximum of 301 K located in the summer hemisphere tropics at about 7°C. Experience with this configuration in CG shows that these simple boundary conditions provide a stable and reasonable zonally averaged control climate.

The only differences between the hemispheres in the GCM experiments are in the solar insolation and SST prescriptions, and the only source of zonal asymmetry is the presence of a continental surface. The continent is 60° wide in longitude, and extends to about 33° in each hemisphere. The land surface is flat and has the same surface roughness and albedo as the ocean surface.

In addition to an all-ocean control integration, five GCM experiments are discussed, with land surface wetness \( w = 1, 2, 5, 10, \) and 15 cm. A wetness value of 15 cm represents a saturated surface—the “swamp” boundary condition—and evaporation, \( E_r \), is at the potential rate, \( E_r \). With surface wetness at 10 cm, there is some restriction on the evaporation, \( E = 0.89 E_r \). The value 5 cm corresponds roughly to “intermediate” surface wetness conditions, and evaporation is restricted to less than half the potential rate \( (E = 0.44 E_r) \). Dry surface conditions are represented by wetness values of 1 and 2 cm, where \( E = 0.09 E_r \) and 0.18 \( E_r \), respectively.

To generate each climatology, the GCM was integrated 300 days from an initially isothermal atmosphere at rest. The quasi-equilibrium climate states represent averages over the subsequent 800 days of integration. To test the stability of these averages, several experiments were integrated for longer periods (up to 1800 days). No changes in the climate state resulted from additional integration length.

Figure 2a shows the tropical precipitation field from the all-ocean control integration. The ITCZ is centered over the imposed SST maximum near 7° in the summer hemisphere. Precipitation in the model is somewhat weaker than the observed; the modeled maximum of 6.0 mm day\(^{-1}\) compares well with the 6.3 mm day\(^{-1}\) from the observed DJF mean, but the observed tropical maximum is broader in latitude because the ITCZ stays in the Northern Hemisphere over much of the Pacific during these months. During JJA, the ITCZ forms a more continuous band, so the observed tropical maximum is narrow as in the model. However, the observed maximum then reaches about 9.5 mm day\(^{-1}\).

Figures 2b–f shows the precipitation difference in the vicinity of the continent for each of the five experiments. The field contoured is the deviation from the control zonal mean. Negative (stippled) regions indicate that the effect of the land surface is to decrease precipitation. The precipitation perturbations for the two wettest cases are very similar to each other (Figs. 2b and 2c). Precipitation within the ITCZ is enhanced, especially in the east, and the precipitation maximum is located deeper into the summer hemisphere over land. Precipitation decreases on the equator in the west.

Over progressively drier surfaces (Fig. 2d, 2e, and 2f), the precipitation decrease spreads eastward and extends deep into the summer hemisphere. With intermediate surface wetness, there is no significant precipitation increase at the latitude of the ITCZ in the continental interior and in the west. Decreases up to 3 mm day\(^{-1}\), a halving of the precipitation rate, occur in the continental interior in the two dry cases. One
feature of the precipitation perturbation emerges independent of surface wetness. In all cases, a precipitation enhancement of about the same magnitude occurs at the same latitude on the east coast.

The purpose of this paper is to understand how these precipitation anomalies are generated, and how they are related to the surface wetness prescription. We start with the analysis of the intermediate wetness case, and then compare with wetter and drier surfaces.

4. Intermediate surface wetness

For the climatology, the precipitation perturbation, $P^*$, is equal to the sum of the evaporation perturbation, $E^*$, and the vertically integrated water vapor flux convergence perturbation according to the budget equation

$$P^* = E^* - \Sigma (\Delta \sigma / g) \text{div}(p_r q V)^*,$$

where $P^*$ and $E^*$ are climatological precipitation and evaporation anomalies (asterisks indicate the deviation from the control zonal mean), $\Delta \sigma$ is the sigma-layer thickness, $g$ is the acceleration due to gravity, $p_r$ is surface pressure, $q$ is water vapor mixing ratio, and $V$ is the horizontal wind speed. The summation is over all vertical levels.

Virtually all of the moisture convergence occurs in the lowest four layers of the model ($\sigma = 0.68, 0.83, 0.94, \text{and} 0.99$). (In the following analysis quantities are calculated on pressure surfaces $p = 680, 830, 940$, and $990 \text{ mb}$, which are nearly coincident with the sigma surfaces since surface pressure variations are small.) At and above $680 \text{ mb}$, the model atmosphere is too dry...
to offer a large contribution to the horizontal moisture convergence. Because the lowest model layer at 990 mb is very thin, the lion’s share of the moisture convergence occurs at the 940- and 830-mb levels. Figure 3 shows the anomalous water vapor flux convergence, mass weighted and integrated over the 830-mb and 940-mb levels. The pattern is very similar to that of the precipitation anomaly (compare Figs. 3 and 2d; 1 mm day$^{-1}$ = 1.16 × 10$^{-5}$ kg H$_2$O m$^{-2}$ s$^{-1}$). The precipitation decrease over the western half of the winter hemisphere tropics coincides with a region of anomalous low-level moisture divergence, while the precipitation increase off the summer hemisphere east coast is associated with moisture convergence in these layers.

The localized divergence maximum in the west over the ocean surface is not reflected in the precipitation field because it is associated with a region of evaporation increase, where the tropical wintertime easterlies blow continental air with low mixing ratio out over the ocean surface. This moisture diverges in the lowest layers of the model below the level of condensation. Elsewhere, the evaporation anomaly is small compared with the precipitation anomaly; $E^*$ is negative in the winter hemisphere, with values up to $-1$ mm day$^{-1}$, and positive in the summer hemisphere, with values up to 0.5 mm day$^{-1}$. Over the oceans, evaporation proceeds at the potential rate. Over the land surface, evaporation is at 44% of the potential rate, but $E_P$ is larger due to the strong dependence of the saturation vapor pressure on temperature.

Figure 4 shows moisture convergence in the individual layers summed in Fig. 3. The dipole pattern in the 830-mb difference field (Fig. 4a) indicates that the moisture divergence maximum that occurs in the tropical winter hemisphere shifts into the summer hemisphere and strengthens by about 50% across the western half of the land surface. A region of anomalous convergence coincides with the location of the east coast precipitation enhancement. At 940 mb, strong moisture convergence occurs in the zonal mean at the latitude of the ITCZ, with strong moisture divergence in the winter hemisphere tropics. Thus, the difference field (Fig. 4b) indicates that the zonal mean pattern is enhanced over the land surface, especially near both coasts. There is no obvious latitudinal shift of the moisture convergence belt as it crosses the land surface at 940 mb.

Figures 3 and 4 imply that one can gain an understanding of how precipitation is modified by the presence of the land surface through understanding differences in the 830- and 940-mb moisture convergence fields. The next step is to consider whether these differences are connected with differences in the wind convergence, the mixing ratio fields, or both. Figures 5a and 5b show that the 830- and 940-mb wind convergence anomalies from the intermediate surface wetness GCM experiment are very similar to the moisture
the first criterion for applying the linear model discussed in section 2. The degree of linearity of the GCM’s dynamical response to the continent must also be established.

Figures 6a and 6b are the linear model wind convergence anomalies at 830 and 940 mb, respectively, from the linear model. To generate these solutions, the linear model is linearized about the zonal mean fields from the intermediate surface wetness experiment. Zonal asymmetries are forced by radiative, latent, and sensible heating plus dry convection and transient eddy heat transports. The agreement with the GCM’s wind convergence anomalies (Figs. 5a and 5b) is good. The dipole pattern emerges in the west at 830 mb, somewhat stronger than in the GCM. The 940-mb perturbation is also reproduced reasonably well in the linear model solution, with the main difference being an underestimate of the magnitude of the convergence enhancement in the summer hemisphere tropics in the east.

This level of agreement between the GCM and linear models is sufficient to suggest that linear dynamics plays

convergence anomalies. At 830 mb, the divergence is enhanced and shifted in the west, and anomalous convergence (a weakening of divergence) appears off the east coast. At 940 mb, the coastal moisture convergence enhancement is captured in the wind convergence field, as is enhanced divergence in the winter hemisphere tropics. The only feature missing is the divergence maximum off the west coast, as this feature is related to structure in the moisture field.

Thus, the precipitation anomaly is replicated in not only the low-level moisture convergence abnormality, but also in the wind convergence field. This suggests that to understand why the precipitation is perturbed by the presence of the continent, we need to understand the response of the low-level dynamics, and this motivates the application of the linear primitive equation model.

a. Linear model applicability

The similarity between the moisture convergence and wind convergence anomalies in the GCM satisfies

Fig. 5. Anomalous wind convergence from the intermediate surface wetness GCM experiment at (a) 830 mb and (b) 940 mb. Contour intervals are $10^{-4}$ s$^{-1}$, and negative values are shaded.

Fig. 6. Anomalous wind convergence from the linear model with full forcing and basic state from the GCM climatology with intermediate surface wetness at (a) 830 mb and (b) 940 mb. Contour intervals are $10^{-4}$ s$^{-1}$, and negative values (divergence) are shaded.
a significant, if not dominant, role in establishing the response, and that further application of the linear model will be useful.

b. Linear model diagnosis

The full linear model solution shown in Figs. 6a and 6b represents the superposition of the effects of differences in the zonal mean fields, transient eddy transports of heat and momentum, and radiative, sensible, dry-convective, and condensational heating. The linear model diagnosis consists of decomposing the full response to understand what factors are important for establishing the convergence field.

Differences in the zonal mean wind and temperature fields between the control and intermediate surface wetness GCM climatologies are small, and the linear solutions exhibit virtually no sensitivity to them for this case. The linear model solution, when the model is linearized about unperturbed mean fields and forced by three-dimensional heating distributions that represent the effects of dry convection and condensational heating, is also very similar to the full linear solution. The effects of sensible, radiative heating and transient heat transports on the perturbations are less important.

At 830 mb, forcing by condensational heating is responsible for most of the response; the wind convergence perturbation forced by the three-dimensional condensational heating field in the linear model is nearly identical to Fig. 6a. The only other nonnegligible contribution at this level comes from sensible heating, which strengthens the anomalous moisture convergence (diminished divergence) over the winter tropics. But it is clear from the linear model that the shift of the winter hemisphere divergence maximum into the summer hemisphere and enhanced convection off the east coast is forced by the modification of the latent heating field over the land surface.

At 940 mb, heating associated with dry and moist convection together force the wind convergence perturbation. The responses to each forcing are shown in Fig. 7; a superposition of the anomalies shown in Figs. 7a and 7b gives Fig. 6b. Dry convection (Fig. 7a) amplifies the zonal mean pattern (convergence in the summer hemisphere tropics, divergence in the winter hemisphere tropics). Anomalous convergence is strongest near the coasts, while the divergence maximum is in the interior. Convergence forced by condensational heating (Fig. 7b) opposes the response to dry convection in the tropics of both hemispheres. The response to moist convection is especially strong in the east.

The three-dimensional forcing field that produces the response in Fig. 7a is represented by the 940-mb heating field shown in Fig. 8a. The dry-convective heating anomaly is defined as the anomalous temperature tendency due to dry-adiabatic adjustment in the GCM, multiplied by \( cp \Delta p / g \), where \( \Delta p \) is the pressure thickness of the model layer centered at \( \sigma = 0.94 \). The structure of the forcing is similar, but weaker, at adjacent levels. This forcing is closely related to the ground temperature difference, which is shown in Fig. 8b. Land surface temperatures are generally lower than the prescribed SSTs throughout the winter hemisphere, and higher in the summer hemisphere. In the summer hemisphere higher ground temperatures are associated with a higher dry-convective heat flux from the surface. As seen in Fig. 8a, two zonally oriented regions of maximum heating occur, one near 25° and a weaker maximum near 5°. As Fig. 7a shows, only weak horizontal convergence is associated with the subtropical heating maximum; much of the atmosphere's response to this heating is rotational, largely nondivergent flow around a 3-mb-deep thermal low. In the tropics, however, the dynamical response is more highly convergent and leads to the large response in the moisture convergence field. The dry-convective heating maximum off the west coast in the winter hemisphere occurs when
middle troposphere in the east, with a deep layer of enhanced condensation from 830 mb to nearly 205 mb. At these levels, diabatic and adiabatic heating are closely related according to

\[ \frac{Q_v}{c_p} = -\omega S, \]

where \( Q_v \) is the heating due to condensation, \( c_p \) is the specific heat at constant pressure, \( \omega \) is vertical \( p \) velocity, and \( S \) is the static stability. This balance has been shown to be relevant for determining surface wind and pressure perturbations in response to thermal forcing in simpler models (Gill 1980). A close relationship between diabatic and adiabatic heating is well established in the GCM throughout the tropics over both land and ocean surfaces (cf. Figs. 10 and 12 in CG).

The east coast condensation, convergence, and precipitation maxima are the response to this enhanced upper-level heating. This result is verified by forcing the linear model with upper- and midlevel (680 mb and above) and lower-level (940 and 990 mb) condensational heating individually. Since the 830-mb level is a transition region, forcing at this level is excluded from both cases. Figure 10a shows the linear response to upper and midlevel condensational heating, and Fig. 10b is the response to lower-level heating. The quantities plotted are anomalous wind convergence, vertically integrated over the 940- and 830-mb levels, since the two levels have similar perturbations in both cases. Both fields are a bit weak due to the loss of the 830-mb forcing, but two distinct distributions have emerged. The low-level response to the upper-level heating (Fig. 10a) has a distribution very similar to the condensational heating and precipitation anomalies, with decreased convergence in the winter hemisphere tropics and increased convergence on the east coast in the summer hemisphere tropics. Thus, the horizontal wind convergence under the condensational heating maximum on the east coast is the response, by continuity, to midtropospheric vertical motion.

The low-level response to the low-level forcing (Fig. 10b) consists of a simple north–south dipole with anomalous divergence in the summer hemisphere tropics and anomalous convergence in the winter hemisphere. This represents a weakening of the zonal mean pattern at 940 mb that is responsible for maintaining the ITCZ precipitation in the GCM.

Other forcing terms play more minor roles in establishing the total response. Radiative effects are minimal for the intermediate surface moisture case. Forcing by thermal transient eddies decreases the amplitudes of both the divergence and convergence maxima by about 30% at 830 mb, but is not effective at 940 mb. The net result is a decrease in the vertically integrated anomalous convergence at the ITCZ latitude. This suggests that the differences between the GCM and linear solutions at 830 mb (Figs. 5a and 6a) could be due to a weak response of the linear model to transient forcing, as would be the case if nonlinear dynamics were more
important in the response to thermal transients than in the response to diabatic heating in this region.

5. Wetter surfaces

Over the wetter land, with surfaces at saturation and two-thirds of saturation (\(w = 15\) and 10 cm, respectively), the enhancement of the ITCZ that occurs for intermediate surface wetness extends farther west into the continental interior (see Figs. 2b and 2c). For example, the central portion of the saturated continent at the latitude of the ITCZ receives 30% more than the zonal mean precipitation, while there is no increase in this region over the continent with intermediate surface wetness.

As in the intermediate surface wetness case, the column moisture convergence is concentrated in the lowest layers of the model, and the anomalous moisture convergence mass integrated over the 830- and 940-mb levels is similar to the precipitation perturbation. The enhanced precipitation in the continental interior over the wetter surface is associated with an increase in the 940-mb moisture convergence; the 830-mb field for the saturated case is quite similar to its counterpart for the intermediate moisture case. This suggests that the difference in the precipitation perturbations can be understood by determining the cause of the differences in the 940-mb convergence fields.

The wind convergence anomaly at 940 mb is also nearly identical to the moisture convergence anomaly over the wetter surfaces. In particular, the enhancement in the continental interior is reflected in this field. The fact that the wind convergence carries the same signal as the moisture convergence encourages use of the linear model to understand the difference in the precipitation fields. The linear model also satisfies the second criterion; the anomalous wind convergence fields are similar to the GCMs when the linear model is forced with full forcing (diabatic heating plus thermal transients, linearized about the perturbed zonal mean fields) from the GCM. (The linear model response at 830 mb is stronger than the 830-mb GCM convergence in the west, as it was for the intermediate wetness case, but the positioning of the maximum and minimum is good.) The response that we want to focus on, namely, enhanced convergence in the summer hemisphere continental interior at 940 mb without enhanced divergence aloft, is represented in the linear solution. As in the intermediate wetness case, the perturbation of the zonal mean fields, sensible heating, and transient eddies play relatively minor roles in the linear model solutions over the wet surface. Convergence induced
by dry convection and condensational heating dominates the response.

Forcing by dry-convective heating, and the response to that forcing, is smaller when the ground temperature perturbation is smaller. Over the saturated surface, warming in the summer hemisphere tropics is much less than over the intermediate wetness surface, especially at the latitude of the ITCZ. Thus, the mechanism that was responsible for a tendency for precipitation to increase over the intermediate wetness surface is not responsible for the increase over the wetter surfaces.

At 830 mb (not shown), the response to condensational forcing resembles that for the intermediate case (Fig. 6a), except that the divergence maximum over the saturated surface in the west is about half as strong. This reduced divergence contributes a little to the precipitation increase, but the largest effect is at 940 mb.

Figure 11 shows the 940-mb wind convergence anomaly due to the condensational heating from the saturated surface GCM climatology. Between the equator and 15° in the summer hemisphere, the divergence that occurred in the continental interior of the intermediate wetness case (Fig. 7b) has been replaced by convergence. Two distinct maxima occur; one on the east coast as for the drier case, and a second of comparable magnitude in the interior. This feature of the convergence field is responsible for the maintenance of the ITCZ in the continental interior.

Figure 12 shows the 940-mb condensational heating anomaly over the saturated continent at 11° latitude. The decrease in condensational heating at low levels from the drier case (Fig. 9) does not occur over the wetter surface. The temperature increase is not large enough to significantly decrease the relative humidity and, thereby, condensation. The enhanced condensation aloft is a little weaker in the east, but extends across the continent. The result is enhanced low-level convergence and precipitation in the continental interior.

6. Drier surfaces

When the land surface is drier than "intermediate wetness," precipitation rates are less than the zonal mean in the west and the continental interior. A close relationship between low-level moisture convergence and precipitation anomalies does not exist in the experiments with \( w = 1 \) and 2; evaporation from the land surface becomes a smaller fraction of the potential evaporation, the evaporation anomaly becomes large, and all three terms in Eq. (1) are important. Large evaporation decreases at low latitudes, up to \( 3 \times 10^{-3} \) kg H₂O m⁻² s⁻¹, rival precipitation decreases, leading to small \( P - E \) and moisture convergence perturbations close to the equator. Evaporation decreases are larger than precipitation decreases over most of the winter.
hemisphere; precipitation and evaporation decreases are comparable over most of the summer hemisphere land surface.

Figure 13 shows moisture convergence anomalies from the GCM experiment with $w = 1$. Figure 13a is the mass-weighted vertical integral over the 830- and 680-mb levels, and Fig. 13b is for 940 mb. Despite the extreme dryness of the surface and the low mixing ratio of the air, strong moisture convergence persists at 940 mb between 5° and 15° in the summer hemisphere, and is even stronger than over wetter surfaces. The strong perturbation winds can apparently compensate for the reduced moisture content of the air to maintain moisture convergence at low levels. In this case, however, moisture convergence does not lead to precipitation. Divergence aloft is also strong (Fig. 13a), and much of the moisture that converges at the 940-mb level diverges at 830 and 680 mb. Note in particular the fate of the moisture that is brought onto the west coast by the monsoon wind; divergence aloft is even stronger than the 940-mb moisture convergence, and this region experiences decreased precipitation. The moisture convergence maximum associated with the precipitation enhancement on the east coast is not present at 940 mb, but appears in the 830–680-mb layer.

If applicable, the linear model wind convergence would be diagnostic of the $P - E$ field, not the precipitation field as in the other cases. A comparison of the GCM's moisture convergence and wind convergence anomalies, however, indicates that understanding the dynamical response alone will not be sufficient to explain the $P-E$ distribution over dry surfaces. In the continental interior and over the western portion of the continent, strong mixing ratio gradients introduce structure in the moisture convergence that is not carried in the wind convergence field. Thus, the linear model is not a useful diagnostic tool for this case.

Without this tool, the physical processes that force moisture convergence and precipitation anomalies in the GCM cannot be identified. However, the analysis for the cases with wetter surfaces suggests analogies. For example, it is reasonable to assume that the east coast precipitation maximum is forced by middle- and upper-level condensational heating, but Fig. 13 indicates that the moisture converges higher in the atmosphere (at 830 and 680 mb) over drier surfaces. Figure 14 shows how the transport of moisture at 830 and 680 mb is related to the precipitation anomaly. The vectors are the anomalous 830-mb water vapor flux, $f$, where

$$f = \left( \frac{\Delta \sigma}{g} \right) \left[ (p_0 q_u)^*i + (p_0 q_v)^*j \right]$$

(3)
in the tropics. In austral summer, precipitation rates are larger than the zonal mean over the eastern portions of Africa and South America, and smaller in the west. This east-west pattern does not emerge clearly for boreal summer when this signal may be obscured over Africa by the role of the Gulf of Guinea as a moisture source and strong circulations to the east associated with the Indian monsoon and Indonesian heating. South America extends to only about 10°N, and the Caribbean Ocean to the north complicates the response in that region in summer.

A low-resolution GCM with simple boundary conditions captures the longitudinal precipitation gradient seen in the austral summer observations. It is always possible that the model is not behaving in the same way as the real world; even when a model simulation is similar to observations, the model can be right for the wrong reasons. The results presented here could also be dependent on the model physics, especially the moist convection scheme. Barring these possibilities, the model results suggest that the observed precipitation gradient occurs because of the presence of land surfaces extending into the subtropics. Analysis of the GCM results with a linear model allows us to identify the physical processes that control the precipitation gradient. Different processes dominate for different values of land surface wetness, leading to the simulated dependence of the precipitation perturbation on surface wetness shown in Fig. 2.

Over very wet land surfaces, the moderate warming of the surface and lower atmosphere leads to an enhancement of the ITCZ across the continent. Static stability decreases, resulting in larger vertical velocities in response to condensational heating above 830 mb [Eq. (2)]. By continuity, convergence below 830 mb is enhanced, feeding more moisture into the ITCZ.

As the land surface becomes drier, two other processes become important. Surface warming leads to enhanced dry convection, which tends to increase precipitation rates in the continental interior by forcing low-level convergence. In addition, the temperature change of the lower atmosphere can become large enough so that the relative humidity is significantly reduced. The resulting decrease in low-level condensation is associated with decreased low-level convergence and precipitation. In the case with intermediate surface wetness, these effects approximately balance over much of the continental interior and the west, and the precipitation perturbation is small. For very dry surfaces, the low-level absolute humidity decreases, as does the relative humidity. Low-level condensational heating is further reduced, as is the low-level moisture convergence and precipitation. A monsoon circulation develops in the west, but it is too shallow to bring rainfall onto the land surface. Enhancement of low-level convergence by midtropospheric condensation persists on the east coast in all cases, and the longitudinal precipitation gradient results.

[see Eq. (1)]. Most of the moisture is fed into the region of enhanced precipitation from the west and the continental interior, although water vapor transport from the east over the oceans is also involved.

As for the case with intermediate surface wetness, diminished condensational heating over land must also be instrumental in decreasing precipitation rates over the continental interior. In contrast to the intermediate wetness case, the absolute humidity drops as well as the relative humidity, so evaporation restrictions (and not just temperature increases) contribute to the decreased relative humidity. Again, by analogy with the other cases, decreased moisture convergence due to the loss of low-level condensation is probably opposed by increased low-level convergence associated with increased dry convection.

7. Conclusions

Observations show that the presence of continents perturbs the large-scale, time-mean precipitation field.
Undoubtedly, continental shape, topography, albedo, and a host of other surface features contribute to determining the observed climatology. The results presented here, however, show that the basic pattern (wet in the east, dry in the west) could occur simply because of the presence of the African and South American continents.

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