Simulation and Diagnosis of the Regional Summertime Precipitation Climatology of South America

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ABSTRACT

The regional precipitation climatology of South America during austral summer is studied by means of an R30 general circulation model (GCM). Results from three perpetual January experiments, which differ only in their distributions of topography and sea surface temperature (SST), are presented. The precipitation field of the most realistic experiment compares well with the observed January precipitation climatology of South America, reproducing, in particular, five regions of maximum precipitation. To understand how structure in the surface conditions is mapped onto the precipitation field, the results of the three GCM experiments are compared. Continentality, through the generation of a thermal low, is responsible for much of the structure in the modeled precipitation field of South America. Topography introduces orographic precipitation maxima in the Central and Southern Andes and modifies precipitation rates elsewhere. Longitudinal structure in SSTs, which is also largely an expression of continentality, is not a dominant source of structure for the South American precipitation field. However, the positions and magnitudes of some of the precipitation maxima (especially those in the east) are moderately affected by SSTs.

Analysis of the atmospheric water vapor budget associates structure in the precipitation field with structure in the moisture flux convergence field. Connections with the large-scale circulation are explored to explain the convergence of moisture flux in each region of enhanced precipitation. Comparisons with observed low-level wind and moisture fields suggest that the mechanisms responsible for the modeled precipitation maxima are, for the most part, reflective of those in the real world.

1. Introduction

The annual amount of rainfall (and snowfall) received by a region is an important characteristic of the regional climate. Deviations from the climatological precipitation rate, especially for extended periods of time, can have significant agricultural, economic, and social consequences. The importance of understanding a region's precipitation climatology and its variability, therefore, cannot be overstated.

In northeastern Brazil, for example, the temporal and spatial distribution of precipitation is highly variable, and droughts are a recurring problem. Accordingly, this region has received much attention from the research community (e.g., Hastenrath and Heller 1977; Moura and Shukla 1981; Kousky 1985; Mechoso et al. 1990). Precipitation variability has been studied in other regions of South America as well, such as southern Brazil (Kousky and Casarin 1986), the Amazon Basin (Marengo et al. 1993), and Uruguay (Pisciottano et al. 1994). In addition, the influence of the El Niño–Southern Oscillation phenomenon on South American precipitation has been extensively studied (e.g., Quinn and Neal 1982; Kousky et al. 1984; Aceituno 1988), as have the potential effects of Amazonian deforestation (e.g., Dickinson and Henderson-Sellers 1988; Nobre et al. 1991; Lean and Rowntree 1993).

Although these studies all contribute to our understanding of regional South American precipitation, they primarily focus on its natural variability and its susceptibility to human-induced change. Relatively few studies (e.g., Nishizawa and Tanaka 1983; Horel et al. 1989; Figueroa and Nobre 1990) have tried to understand the structure in the climatological precipitation field of South America and how it is related to the large-scale circulation. For this reason, the mechanisms responsible for the regional structure in the precipitation field are not fully understood.

In order to increase our understanding of this issue, the regional precipitation climatology of South America during austral summer is analyzed using a relatively high-resolution general circulation model (GCM). Structure in the precipitation field is related to features of the large-scale circulation through analysis of the atmospheric water vapor budget. In addition, the results of three mechanistic GCM experiments reveal the individual roles of continentality, topography, and sea surface temperatures (SSTs) in generating this structure. (We define "continentality" as the contrast be-
between land and sea in the absence of topography and longitudinal SST variations.) To cleanly diagnose the roles of these three surface features, other GCM boundary conditions are kept simple. It is not necessarily the intention of this study to generate the most realistic simulation of precipitation possible (again, mechanisms are the focus). On the other hand, the modeled precipitation field in our most realistic experiment compares very favorably with observations. This is encouraging, especially since many GCMs have difficulty accurately simulating regional South American precipitation (Legates and Willmott 1992).

A description of the GCM and the three experiments is given in section 2. In section 3 the modeled South American precipitation climatology from the most realistic experiment is compared with observations, and a diagnosis of how structure at the surface leads to structure in the precipitation field is presented. The atmospheric water vapor budget is analyzed in section 4, and connections with the large-scale circulation are discussed in section 5. Comparisons with observations are presented in section 6 to assess the relevance of the analysis to the real climate. Section 7 addresses concerns regarding model simplifications, while section 8 presents a summary of the main conclusions.

2. Model and experiments

a. Model

The model used in this study is a version of the GCM developed and maintained by the Climate Dynamics Group at NOAA’s Geophysical Fluid Dynamics Laboratory [GFDL; see Manabe (1969) for a general description]. It is a spectral model with R30 resolution, which corresponds to 3.75° longitude by about 2.25° latitude on the transform grid. The model solves the primitive equations plus a prognostic equation for water vapor mixing ratio on 22 vertical levels in sigma coordinates. The sigma levels (at $\sigma = 0.015, 0.045, 0.073, 0.092, 0.107, 0.125, 0.146, 0.170, 0.198, 0.231, 0.269, 0.313, 0.365, 0.426, 0.496, 0.578, 0.676, 0.777, 0.866, 0.935, 0.979$, and 0.997) have been chosen to give higher vertical resolution near the surface and the tropopause. The model time step is 20 minutes and variables are sampled daily. Cumulus convection is parameterized by means of a moist convective adjustment.

In order to isolate the effects of continentality, topography, and SSTs on the South American climate, some of the boundary conditions have been kept simple. The surface drag coefficient is fixed and globally uniform so that the effects of surface roughness variations are not considered. Surface albedo is fixed at a value of 0.1 everywhere except where snow accumulates on land surfaces and in regions of glaciated land (e.g., Antarctica and Greenland). In such areas the albedo is calculated within the model and depends on the surface temperature and snow depth. The surface hydrology employs the “bucket method,” in which soil moisture is expressed as an amount of water, in centimeters, with 15 cm considered saturation (the “field capacity”). Typically, soil moisture is calculated from the precipitation and evaporation, with amounts in excess of the field capacity being designated as runoff. Here, however, the soil moisture is not allowed to respond to the precipitation field but is instead fixed at 5 cm for all land points. This eliminates structure in the precipitation field that would result from a geographic distribution of soil moisture, allowing one to focus on the effects of other boundary conditions. Clouds are not predicted but are fixed at their observed annual mean values. The distribution is zonally uniform and uses hemispherically symmetric Southern Hemisphere values from London (1957). Ozone concentrations are also zonally uniform and the same in both hemispheres. The model is run in a “perpetual January” mode so that the sun’s zenith angle represents austral summer solstice conditions. Diurnal solar forcing is not included, and sea ice is not allowed to form.

b. Experiments

Results from three experiments are discussed. We refer to them as the SST, mountain, and no-mountain experiments, and they differ only in their distributions of topography and SSTs (Table 1). Each experiment is at least 2100 days long, with the first 300–400 days devoted to spinup from a dry, isothermal atmosphere at rest. The remaining days are averaged to form the climatology.

The boundary conditions for the three experiments are depicted in Fig. 1 for the vicinity of South America. The topography used in the mountain and SST experiments (Fig. 1b) has been filtered (Lindberg and Broccoli, personal communication 1994) to reduce Gibbs oscillations present near narrow topographic features (such as the Andes). This results in a smoothed representation of the Andes, with the maximum height reduced to roughly 2 km (as opposed to the observed elevation of about 3.8 km). Nonetheless, the precipitation field is dramatically improved as a result of the reduction in Gibbs ripples. In addition to being lower in elevation, the filtered topography is also broader. To avoid an elevated sea surface, the model’s continental

<table>
<thead>
<tr>
<th>Experiment name</th>
<th>Boundary conditions</th>
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<tr>
<td></td>
<td>continents</td>
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<tr>
<td>SST</td>
<td>global</td>
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<tr>
<td>Mountain</td>
<td>global</td>
</tr>
<tr>
<td>No-mountain</td>
<td>global</td>
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boundary is extended in certain areas, especially along the west coast of South America (Fig. 1a).

A zonally uniform SST prescription, $T_s$, is used in the no-mountain and mountain experiments (Fig. 1a) and is given (in Kelvin) by

$$T_s = 287 - 10 \sin \phi - (0.333)(40)(3 \sin^2 \phi - 1),$$

(1)

where $\phi$ is latitude. This SST distribution is similar to that of an average solstice, with a temperature maximum of roughly 301 K at about $7^\circ$S and temperatures of 250 and 270 K at $90^\circ$N and $90^\circ$S, respectively. The realistic SSTs used in the SST experiment (Fig. 1b) are those of Shea et al. (1990) and have a zonal-mean distribution that differs from $T_s$ (e.g., note the positions of the equatorial SST maxima in Fig. 1).

The mass of the atmosphere is the same in all three experiments. As a result, the globally averaged sea level pressure in the no-mountain experiment is lower than in the other two experiments, because the atmosphere fills the void left by the topography. To remove this "hydrostatic effect" of topography from the analysis, the results of the no-mountain experiment are interpolated to slightly lower pressure levels than those of the other two experiments. The ratio of the pressure levels is equal to that of the globally averaged sea level pressures.

3. Regional diagnosis of South American precipitation

Figure 2a presents observed January precipitation rates for South America from Legates and Willmott (1990), while modeled precipitation rates from the SST experiment are shown in Fig. 2b. The observed data have been interpolated from a half-degree grid to the model's R30 resolution. The position of the oceanic intertropical convergence zone (ITCZ) is well represented in the model, and four distinct continental precipitation maxima are evident in both the model and the observations. These maxima are located in the central, northwest, west central, and southern regions of South America—regions referred to here as the Amazon, Northern Andes, Central Andes, and Southern Andes regions, respectively (Fig. 3). A fifth rainfall maximum occurs in both the model and observations in the form of a northwest–southeast-oriented band extending off the southeast coast into the South Atlantic. This band is known as the South Atlantic convergence zone (SACZ) and, while represented in Fig. 2a, is even more prominent in cloud observations (see, e.g., Kodama 1992). The dry regions of South America are also well represented in the model, namely, along the central west coast, the very southeast portion of the continent, and the eastern tip of Brazil.

Some discrepancies between the modeled and observed precipitation distributions are evident. The precipitation maximum of the Northern Andes does not extend as far out over the ocean as in the observations. Also, the modeled precipitation maximum in the Southern Andes is weaker than the observed, while in the Central Andes the maximum is stronger (by more than 6 mm day$^{-1}$ in some places) and more extensive than in the observations.

Apart from the above exceptions, the positions and magnitudes of the regional precipitation maxima compare favorably with the observations, especially considering that the discrepancies may well be within the
uncertainty in the observations themselves. (The distribution of observing stations over some areas of South America is very sparse.) We feel that this agreement is strong enough to suggest that an analysis of the mechanisms that lead to the modeled precipitation distribution will be informative about processes in the real world. Favorable comparisons between other modeled and observed fields (to be shown later) support this suggestion.

Even with the model’s simplified boundary conditions, the GCM reproduces the observed structure of the South American precipitation field with reasonable accuracy. This suggests that continentality, topography, and/or SSTs are more important sources of structure than are other boundary conditions such as surface albedo, wetness, and roughness. Section 7 provides further evidence for this suggestion.

Ultimately, any structure in the precipitation field of the SST experiment is due to the presence of continents, topography, and structure in the SST distribution. To distinguish between these three influences we compare, by region, the precipitation field of the SST experiment (Fig. 2b) with that of the mountain (Fig. 2c) and no-mountain (Fig. 2d) experiments.

a. Amazon region

A precipitation maximum is present in the Amazon region in each of the three experiments, indicating that continentality is the source of this particular maximum. The addition of topography acts to augment the precipitation rate in the Amazon region (Figs. 2c and 2d), while SSTs affect the longitudinal position of the maximum, in addition to its magnitude (Figs. 2b and 2c).
b. Northern Andes

As was the case for the Amazon region, continentality is the primary cause of the Northern Andes precipitation maximum (Fig. 2d). Topography displaces the maximum about 10° to the northwest, enhancing its distinction from the Amazonian maximum (Fig. 2c). Realistic SSTs further distinguish the Northern Andes maximum from the remaining precipitation field by shifting it about 5° to the north (Fig. 2b). A similar shift of the ITCZ also takes place.

c. Central Andes

The precipitation maximum in the Central Andes does not exist in the absence of mountains (Fig. 2d). Only when topography is added does this maximum appear (Fig. 2c). The effect of SSTs is limited (Fig. 2b), though precipitation in the northwest portion of the Central Andes is weaker in the presence of realistic SSTs.

d. Southern Andes

The Southern Andean precipitation maximum is also the direct result of topography (Figs. 2c and 2d), while the effect of SSTs is minimal.

e. SACZ

As is the case for the Amazon region, precipitation in the SACZ occurs in the absence of topography and longitudinal structure in SSTs (Fig. 2d). Mountains slightly enhance the precipitation in this region (Fig. 2c), while SSTs affect both its position and intensity (Fig. 2b).

While the effects of the earth's topography on precipitation in the Andean regions predominantly reflect the influence of the Andes, this may not be the case in the Amazon or the SACZ. For example, in the GCM the precipitation in Southern Africa is considerably altered by the presence of African topography. This could be affecting South American precipitation through a Walker circulation.1

4. Analysis of the atmospheric water vapor budget

A more complete understanding of the regional South American precipitation field can be obtained by relating the structure in the precipitation field to features of the large-scale circulation through the atmospheric water vapor budget (see, e.g., Peixoto and Oort 1992). The climatological precipitation rate, $\bar{P}$ (mm day$^{-1}$), is equal to the evaporation rate, $E$, minus the vertically integrated sum of the water vapor flux divergence and the local rate of change of water vapor mixing ratio according to

$$\bar{P} = E - \int_0^{\infty} \left[ \nabla_3 \cdot q u_3 + \left( \frac{\partial q}{\partial t} \right) \right] \frac{dp}{g \rho_w}$$

$$= - \int_0^{\infty} \rho_s \left[ \nabla_3 \cdot q u_3 + \left( \frac{\partial q}{\partial t} \right) \right] \frac{d\sigma}{g \rho_w} + E, \quad (2)$$

where $p_s$ is the surface pressure, $\nabla_3$ is the three-dimensional divergence operator in pressure coordinates, $q$ is the water vapor mixing ratio, $u_3 = (u, v, w)$ is the three-dimensional wind velocity (in pressure coordinates), $g$ is the acceleration due to gravity, $\rho_w$ is the density of water, $d\sigma$ ($dp$) is the infinitesimal sigma-level (pressure level) thickness, and overbars indicate the climatological mean. Splitting the integral in (2) into time-mean and transient components gives

$$\bar{P} = - \frac{\bar{E}}{g \rho_w} \int_0^{\infty} \left( \nabla_3 \cdot \bar{q} \bar{u}_3 \right) d\sigma + \bar{T} + E, \quad (3)$$

where $\bar{T}$ is the integrated water vapor flux convergence by the transient eddies and can be approximated as

$$\bar{T} \approx - \frac{1}{g \rho_w} \nabla_3 \cdot \left( \int_0^{\infty} \left( \bar{p}, \bar{q} \bar{u}_2 - \bar{p}, \bar{q} \bar{u}_2 \right) d\sigma \right), \quad (4)$$

where $\nabla_3$ is the horizontal divergence operator and $u_2$ is the horizontal wind velocity. [See the appendix for a derivation of (4).] Expanding the divergence operator in (3) into its horizontal and vertical components gives

1 Teleconnections between Africa and South America have been noted in the literature (Vines 1982; Moura and Kagano 1983), and we have observed in some of our GCM experiments that Africa can have a significant influence on the precipitation in eastern South America.
\[ P = -\frac{P_s}{g \rho_w} \int_0^1 \left[ \nabla_{2p} \cdot q u_2 + \frac{\partial}{\partial p} (q \omega) \right] d\sigma + T + E, \]

where the overbar notation has been dropped and the horizontal divergence is calculated on constant pressure surfaces. Integrating the vertical derivative and writing the remaining integral as a finite sum, (5) becomes

\[ P = -\frac{P_s}{g \rho_w} \sum_{\sigma=0}^1 (\nabla_{2p} \cdot q u_2) \Delta \sigma - \frac{1}{g \rho_w} (q \omega), + T + E, \]

where the subscript \( s \) indicates evaluation at the surface (\( \sigma = 1 \)) and \( q \omega \) is assumed to vanish at the top of the atmosphere. Equation (6) can be decomposed as

\[ P = C + A + O + T + E, \]  

where

\[ C = -\frac{P_s}{g \rho_w} \sum_{\sigma=0}^1 (q \nabla_{2p} \cdot u_2) \Delta \sigma, \]

\[ A = -\frac{P_s}{g \rho_w} \sum_{\sigma=0}^1 (u_2 \cdot \nabla_{2p} q) \Delta \sigma, \]

and

\[ O = -\frac{1}{g \rho_w} (q \omega), \]

The terms \( C, A, O, T, \) and \( E \) will hereafter be referred to as the convergence,\(^2\) advection, orographic, transients, and evaporation terms, respectively. Term \( C \) represents precipitation associated with horizontal wind convergence, while \( A \) is that of horizontal moisture advection. Here \( O \) is the vertical moisture flux convergence, but since \( \omega \approx 0 \) everywhere at the surface except where winds flow across topography, \( O \) represents orographic precipitation.\(^3\)

The five terms on the right-hand side of (7a), together with their sum, are shown in Fig. 4 for the SST experiment. There is a significant amount of structure in each term, except for the evaporation field (Fig. 4e). Even in regions where evaporation rates are large (relative to the precipitation rate), this term does not exhibit structure similar to that of the precipitation field. The slight discrepancy between \( P \) (Fig. 2b) and the sum of the five terms (Fig. 4f) is the result of approximations involved in calculating the horizontal derivatives and the transients term and in extrapolating \( q \omega \) to the surface.

Table 2 summarizes Fig. 4 for each South American region. The five water vapor budget components have each been averaged over the areas shown in Fig. 3 and written as a percentage of the regionally averaged precipitation rate. In each region of enhanced precipitation the diagnosis of the atmospheric water vapor budget indicates strong dynamical support for the rainfall amplifications. The above decomposition shows that this support can occur in different ways, as summarized below for each region.

a. Amazon region

The precipitation maximum in the Amazon region is primarily associated with large-scale wind convergence (Fig. 4a), and some water vapor is removed by transient eddies (Fig. 4d). Though evaporation does not cause the enhanced precipitation rates in this region, it does supply roughly 30% of the available moisture (Table 2).

b. Northern Andes

As in the Amazon region, the precipitation maximum of the Northern Andes is associated with strong low-level wind convergence (Fig. 4a). Transient eddies remove some water vapor from the region (Fig. 4d). Orographic precipitation (Fig. 4c) is weak in the model despite the mountainous terrain of the Northern Andes. While this may also be the case in the real world, it is possible that the topography of this region (which is much higher and narrower than portrayed in the model) plays a more significant role than is represented here. For example, Figueroa and Nobre (1990) suggest that precipitation along the eastern and western flanks of the Northern Andes is primarily orographic in nature.

c. Central Andes

Rainfall in the Central Andes region is mainly orographic, with the wind convergence also playing a significant role (Table 2). The relative importance of these two components varies greatly within the region. Strong orographic precipitation occurs along the eastern flank of the Central Andes (Fig. 4c), while strong wind convergence occurs at higher elevations (Fig. 4a) along a ridge running from northwest to southeast. There is also a significant orographic component along the western flank of the Central Andes, but this is offset by divergent winds (Fig. 4a) and the advection of drier air (Fig. 4b). The importance of orographic uplift for precipitation in the eastern Central Andes has been mentioned in the literature (e.g., Figueroa and Nobre 1990).

\(^2\) The term "convergence" is used rather loosely here and refers to the vertically integrated product of wind convergence and mixing ratio. Most of the structure in this term is determined by structure in the wind convergence field, not the moisture field.

\(^3\) Positive values of \( C, A, O, T, \) or \( E \) do not necessarily represent realized precipitation amounts, only the contribution toward the total precipitation. Since a term can be offset to a large degree by another term of opposite sign (as the orographic term often is by the advection term), it is possible to have a term whose value is large and positive in a region where there is, in fact, little or no precipitation.
Fig. 4. The (a) convergence, (b) advection, (c) orographic, (d) transients, and (e) evaporation terms in the water vapor budget equation [(7a), SST experiment only], along with (f) their sum. Contour interval is 3 mm day$^{-1}$ (solid lines). The ±1 and ±2 mm day$^{-1}$ contours are also shown (dashed lines), and negative values are shaded.
Table 2. Components of the water vapor budget \((7a)\) for the SST experiment, written as a percentage of the regional precipitation rate.

<table>
<thead>
<tr>
<th>Region</th>
<th>C</th>
<th>A</th>
<th>O</th>
<th>T</th>
<th>E</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amazon</td>
<td>87</td>
<td>-8</td>
<td>11</td>
<td>-18</td>
<td>29</td>
</tr>
<tr>
<td>Northern Andes</td>
<td>96</td>
<td>-8</td>
<td>8</td>
<td>-25</td>
<td>28</td>
</tr>
<tr>
<td>Central Andes</td>
<td>43</td>
<td>-20</td>
<td>70</td>
<td>-9</td>
<td>26</td>
</tr>
<tr>
<td>Southern Andes</td>
<td>-21</td>
<td>-61</td>
<td>129</td>
<td>1</td>
<td>51</td>
</tr>
<tr>
<td>SACZ</td>
<td>32</td>
<td>24</td>
<td>11</td>
<td>-39</td>
<td>66</td>
</tr>
</tbody>
</table>

\(C = -\frac{P_s}{g \rho_w} \sum_{\sigma = 0}^{1} (q \nabla u \cdot u_2) \Delta \sigma = C_z + C_m, \) \(8a\)

where

\[ C_z = -\frac{P_s}{g \rho_w} \sum_{\sigma = 0}^{1} \left( \frac{q}{R \cos \phi} \frac{\partial u}{\partial \lambda} \right) \Delta \sigma \] \(8b\)

and

\[ C_m = -\frac{P_s}{g \rho_w} \sum_{\sigma = 0}^{1} \left( \frac{q}{R \cos \phi} \frac{\partial}{\partial \phi} (v \cos \phi) \right) \Delta \sigma \] \(8c\)

\((\lambda, \phi, \text{and} \ R \text{ are longitude, latitude, and radius of the earth, respectively}). \ Figure \ 6 \ shows \ C_z \text{ and } C_m \text{ for each experiment.} \)

The Amazonian precipitation maximum in all three experiments is associated with a zonal convergence maximum (Figs. 6a, 6c, and 6e) that varies in intensity and position in accordance with the precipitation (Figs. 2b–2d). This convergence maximum is related to the counterclockwise turning of the northeasterly trade winds crossing South America (Fig. 5). These winds attain a northwesterly direction in the Amazon region, where they converge with easterlies emanating from the South Atlantic high. This convergence is primarily associated with the continental low (Figs. 5e and 6e) and not with mechanical funneling of the northeasterly trades along the concave eastern flank of the Andes, as has been suggested by Nobre et al. (1991).

5. Connections with the large-scale circulation

The decomposition of the atmospheric water vapor budget provides a means of relating structure in the South American precipitation field to features of the large-scale circulation. This in turn helps to explain the modeled effects of continentality, topography, and SSTs on the precipitation distribution.

The low-level wind and geopotential height fields are shown in Fig. 5 for the three GCM experiments defined in Table 1. In each experiment, the strong South Pacific and South Atlantic highs flank South America to the west and the east, respectively. In addition, the northeasterly trade winds turn into the cyclonic flow of a continental low (or trough) positioned over tropical and subtropical portions of South America. Most of the structure in the precipitation field can be related to these circulation features.

a. Amazon region

As discussed above, the modeled Amazonian precipitation is primarily associated with low-level wind convergence. To better relate this to the large-scale circulation, the convergence term (Eq. 7b) is split into its zonal and meridional components:

\[ C = -\frac{P_s}{g \rho_w} \sum_{\sigma = 0}^{1} (q \nabla u \cdot u_2) \Delta \sigma = C_z + C_m, \) \(8a\)

where

\[ C_z = -\frac{P_s}{g \rho_w} \sum_{\sigma = 0}^{1} \left( \frac{q}{R \cos \phi} \frac{\partial u}{\partial \lambda} \right) \Delta \sigma \] \(8b\)

and

\[ C_m = -\frac{P_s}{g \rho_w} \sum_{\sigma = 0}^{1} \left( \frac{q}{R \cos \phi} \frac{\partial}{\partial \phi} (v \cos \phi) \right) \Delta \sigma \] \(8c\)

\((\lambda, \phi, \text{and} \ R \text{ are longitude, latitude, and radius of the earth, respectively}). \ Figure \ 6 \ shows \ C_z \text{ and } C_m \text{ for each experiment.} \)

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b. Northern Andes

Precipitation in the Northern Andes is associated with zonal wind convergence to the east (Fig. 6a) and meridional wind convergence to the west (Fig. 6b). As is the case for the ITCZ, the meridional component is associated with the convergence of the trade winds (Fig. 5a). Cyclonic flow about the continental low enhances this convergence in the Northern Andes and also forms monsoonal westerlies that flow onto the continent at about 2ºS. These westerlies, which converge with the northeasterly trade winds, are the source of the zonal wind convergence in the eastern Northern Andes (Fig. 5a).

As indicated by the no-mountain experiment, the enhanced meridional convergence (Fig. 6f) and the monsoonal westerlies (Fig. 5e) are the direct result of continentality and the thermal low. The effects of longitudinal structure in the SSTs are minimal, but the role of the Andes is significant. The addition of topography does not simply shift the broad meridional convergence maximum (Fig. 6f) to the northwest, as might be expected from comparing the precipitation fields (Figs. 2c and 2d). Instead, topography causes the continental portion of the convergence maximum to be concentrated over the Central Andes, while the oceanic portion is displaced to the north (Figs. 6d and 6f). In addition, the Andes cause the monsoonal westerlies to
Fig. 5. Geopotential height (contour interval 10 m) and wind vectors at (a) 950 mb and (b) 785 mb for the SST experiment. (c), (d) As in (a), (b), respectively, but for the mountain experiment. (e), (f) As in (a), (b), respectively, but for the no-mountain experiment at 924 and 764 mb. The vector in the lower-right corner denotes a wind speed of 20 m s⁻¹, and contours less than 550 (left panel) and 2150 m (right panel) are not plotted. Hatching indicates regions below ground.
FIG. 6. The (a) zonal and (b) meridional convergence terms [(8)] for the SST experiment. (c), (d) As in (a), (b), respectively, but for the mountain experiment. (e), (f) As in (a), (b), respectively, but for the no-mountain experiment. Contour interval is 3 mm day$^{-1}$, and negative values are shaded.
shift from about 8°S to 3°S (Figs. 5c and 5e), where there is a minimum in the topography (Fig. 1b), indicating that the Andes may be mechanically funneling this flow. This northward shift of the westerlies introduces the zonal convergence maximum in the eastern Northern Andes (Figs. 6c and 6e).

c. Central Andes

Precipitation in the Central Andes is associated with orographic uplift along the eastern slopes and meridional wind convergence at higher elevations. As shown in Figs. 5a and 5c, the orographic uplift is associated with north to northwesterly winds that impinge obliquely on the topography. These winds are part of the same northwesterly wind regime that converges with the tropical easterlies in the Amazon region.

Strong meridional convergence in the Central Andes (Figs. 6b and 6d) is associated with small-scale, cyclonic circulation along the western portion of the continental low (Figs. 5b and 5d). This small-scale convergence may be due to strong sensible and latent heating over the Andes. This would be encouraging since the description of the observed convection in this region is very similar (Gutman and Schwerdtfeger 1965; Rao and Erdogan 1989).

d. Southern Andes

The simulated precipitation in the Southern Andes is purely orographic (Table 2). This orographic precipitation is associated with the midlatitude westerlies (Fig. 5), which experience mechanical uplift by the meridionally oriented topography of this region (Fig. 1b).

e. SACZ

The SACZ is characterized by low-level wind convergence and moisture advection (Table 2), which both contribute positively to the regional precipitation. In addition, moisture flux divergence by the transient eddies is significant in this region. The wind convergence in the SACZ is meridional, with some offset by zonal divergence (Fig. 6). This meridional convergence is associated with northerly winds along the western flank of the South Atlantic high (eastern flank of the continental low) that converge in the southwest corner of the high, where the flow is westerly (Fig. 5).

The strength of the advection and transients terms in the SACZ suggests that the structure in the atmospheric moisture field plays an important role in determining the precipitation rate in this region. Figure 7 shows the mixing ratio and horizontal wind velocity at 850 mb from the SST experiment. The positive advection of water vapor into the SACZ is evident since the northerly and northwesterly winds along the southwest portion of the South Atlantic high flow down a strong moisture gradient. The mixing ratio forms a tongue-like maximum that extends off the coast of South America toward the southeast (Fig. 7).

The advection and transients terms roughly cancel each other in the SACZ region (Figs. 4b and 4d) because both are related to the moisture gradient. To illustrate this, the vertically integrated transient water vapor flux \( \Sigma \left[ (\rho_q \bar{u}_2 - \bar{p} \bar{q} \bar{u}_2 - \bar{p} \bar{q} \bar{u}_2) \Delta T / g \rho_w \right] \) for the SST experiment is shown in Fig. 8. Note the relatively strong southeastward flux of moisture along the axis of the SACZ, approximately directed across the gradient in mixing ratio (Fig. 7). The direction of the transient flux agrees with intuition, since one would expect wind speed to be positively (negatively) correlated with the mixing ratio when the flow is coming from a region of higher (lower) mixing ratio. Thus, the transient moisture flux vectors in Fig. 8 are directed away from each other along the axis of the tongue-like mixing ratio maximum, yielding the divergence shown in Fig. 4d. Also apparent in Figs. 4d and 8 is some weak transient moisture flux convergence just to the south of the SACZ. Thus, instead of concluding that transient eddies reduce the SACZ's precipitation, it may be more appropriate to say that they displace the SACZ to the south.

The modeled transient moisture flux in the SACZ represents roughly 25% of the total flux (not shown) and so is an important component of the moisture transport in this region. This is in agreement with observational studies such as Chen (1985). The fact that the transient moisture flux is particularly strong in the vicinity of the SACZ (and other regions not shown, such as the storm tracks of the North Atlantic and the North Pacific) indicates the significance of transient weather systems (e.g., extratropical cyclones and frontal activity) for this region. Similar descriptions of the SACZ
have been noted in the literature (e.g., Kousky 1979; Oliveira and Nobre 1986; Kodama 1993). Continentiality has been shown to be the cause of precipitation in the SACZ, and the preceding analysis indicates that the explanation of this is related to the presence of the continental low. Specifically, the northerly winds along the western flank of the South Atlantic high, which are associated with the meridional wind convergence and southward advection of moisture in the SACZ, can be attributed to the geopotential gradient between the continental low and the South Atlantic high. In addition, the majority of the structure in the SACZ's mixing ratio and transient moisture flux fields is present in the no-mountain experiment (not shown). The significance of the continental low for the maintenance of the SACZ has been suggested before by Kodama (1993), who also noted that the SACZ (like the continental low) is weak during austral winter.

6. Comparison with observations

To increase our confidence that the analysis of the GCM experiments is providing insight into how the South American precipitation distribution is established in nature, we compare our results with observations. In this section, we try to go beyond a simple comparison of the observed and modeled precipitation fields (Figs. 2a and 2b) to support our conclusions about the mechanisms associated with the regional South American precipitation maxima.

The dataset used is a climatology based on initialized analyses from the European Centre for Medium-Range Weather Forecasts (ECMWF) and represents an average of ten months of January data from 1980 to 1989 (Trenberth 1992). Archived on a 2.5° grid at seven pressure levels, the data has been horizontally interpolated to the model's R30 resolution.

This comparison examines fields that are notoriously difficult to observe (e.g., mixing ratio and wind convergence), especially over South America, where the observing network is very sparse in many regions. To some degree, we are comparing two models (the GCM and the model used by ECMWF). Nevertheless, for the purposes of this comparison, the ECMWF analyses are probably the best currently available.

a. Amazon region

Figures 9a and 9b show the observed convergence and advection terms, respectively. As in the model
(Fig. 4a), the observations show a convergence maximum in the Amazon region (Fig. 9a). However, the modeled maximum reaches a magnitude of more than 15 mm day\(^{-1}\), while the observed is just over 6 mm day\(^{-1}\).

A breakdown of the observed convergence term into its zonal (Fig. 10a) and meridional (Fig. 10b) components supports the model’s result (Figs. 6a and 6b) that the Amazonian precipitation maximum is associated with zonal wind convergence. Also, the magnitudes of the modeled (Fig. 6a) and observed (Fig. 10a) zonal convergence maxima are in better agreement than the full convergence term, so the model’s overestimation of the convergence term in this region is related to a lack of meridional divergence (Figs. 6b and 10b).

Figure 11 shows low-level wind and geopotential height fields from observations and the SST experiment. The modeled northeastercly trade winds, continental low, and South Pacific and South Atlantic highs (Figs. 11b and 11d) all compare favorably with the observations (Figs. 11a and 11c). In particular, the proposed association between these features and the Amazonian convergence maximum is borne out in the observations, which show northwesterly winds converging in the Amazon region with the easterlies of the South Atlantic high (Fig. 11c). These easterlies are stronger in the model than in the observations (by more than 5 m s\(^{-1}\) in some places), which partially accounts for the stronger zonal wind convergence in the model.

b. Northern Andes

The model’s simulation of zonal convergence in the eastern Northern Andes (Fig. 6a) and meridional convergence in the western Northern Andes (Fig. 6b) is corroborated by the observations (Figs. 10a and 10b). As in the Amazon region, however, the lack of significantly offsetting divergence in the model results in an overestimation of the net convergence (Figs. 4a and 9a).

The meridionally converging trade winds of the ITCZ and the western Northern Andes are well simulated by the model (Figs. 11a and 11b), as are the weak monsoonal westerlies, at least at 1000 mb. The significance of the monsoonal westerlies for the precipitation in this region has also been noted by Figueroa and Nobre (1990). At 850 mb both the GCM and the observations show easterlies converging in the eastern portion of the Northern Andes (Figs. 11c and 11d), but the modeled monsoonal westerlies and generally weak flow to the west do not match the ECMWF observations, which show easterlies of almost 10 m s\(^{-1}\).

c. Central Andes

The modeled meridional convergence maximum of up to 15 mm day\(^{-1}\) in the Central Andes (Fig. 6b) compares reasonably well with an observed maximum of over 9 mm day\(^{-1}\) (Fig. 10b). The net convergence is excessive (Figs. 4a and 9a), however, as was also the case for the simulated precipitation rate in this region (Figs. 2a and 2b). Nevertheless, the observations do show a band of positive convergence in the Central Andes (Fig. 9a), supporting the model’s indication that orographic precipitation does not act alone in this region.

The north to northwesterly winds associated with the modeled orographic precipitation along the eastern flank of the Central Andes are in good agreement with the observations at 850 mb (Figs. 11c and 11d).

d. Southern Andes

In the Southern Andes, the modeled midlatitude westerlies (Fig. 11d) compare favorably with the ob-
served (Fig. 11c). In addition, the simulation of dry-air advection (Fig. 4b) and meridional divergence (Fig. 6b) is supported by the observations (Figs. 9b and 10b).

e. SACZ

The observations corroborate the model's simulation of a meridional convergence maximum in the SACZ (Figs. 6b and 10b). In this case the modeled net convergence is weaker than the observations by about 1–2 mm day⁻¹ (Figs. 4a and 9a). There is good agreement between the modeled and observed advection terms in the SACZ (Figs. 4b and 9b), where a positive band of more than 2 mm day⁻¹ is present.

Figure 11 reveals a good correspondence between the modeled and observed winds in the SACZ. In particular, the northerly winds between the continental low and the South Atlantic high, which are important for the convergence and advection in this region, are well simulated by the model.

Since the structure in the moisture field is especially important for the SACZ, the observed mixing ratio and winds at 850 mb are shown in Fig. 12. Comparison with Fig. 7 reveals that although the magnitude of the modeled mixing ratio is low (by 1–4 g kg⁻¹), there is good correspondence between the overall structure in the two fields. The simulation of a strong moisture gradient in the SACZ and a tongue-like maximum in mixing ratio is supported by the observations, though the modeled maximum is lo-
cated farther southwest and is more distinct than the observed.

7. Sensitivity to simplified boundary conditions

a. Land surface conditions

The preceding section reveals a good correspondence between the results of the SST experiment and observations. Nevertheless, considering the simple boundary conditions used in the three GCM experiments, one might be concerned that the simulated effects of continentality, topography, and SSTs do not accurately reflect the behavior of the real atmosphere. For example, it is possible that the assumptions of uniform surface roughness, albedo, and soil moisture may be eliminating important structure in the South American precipitation or evaporation fields. We offer evidence from a GCM experiment that suggests this is not the case. This experiment was run with realistic January values of albedo and soil moisture (from Matthews 1985; and Mintz and Walker 1993, respectively) and with the surface drag coefficient a factor of 3 larger over land than over ocean. In most other respects the experiment is similar to the SST experiment.4

The basic structure in the South American precipitation field for the “land surface experiment” (not shown) is very similar to that of the SST experiment, with the five regions of precipitation maxima being clearly present. Except for the Amazonian precipitation maximum, which is positioned farther northwest, the magnitudes and positions of the precipitation maxima remain relatively unchanged. And while the evaporation field contains more structure than that of the SST experiment, the difference amounts to less than 1 mm day⁻¹ in the hydrologically active regions of South America. For these reasons, the basic conclusions of the present study regarding the roles of continentality, topography, and SSTs remain unchanged. In fact, these results reinforce the idea that continentality, topography, and the large-scale circulation play a more fundamental role in establishing regional structure in the South American precipitation field than do geographic variations in land surface conditions and the evaporation field. That is not to say that such geographic variations are not important. For example, while the Amazonian precipitation maximum is the result of continentality, its location is sensitive to land surface conditions.

b. Zonally uniform SSTs

It may also be of some concern that the differences between the mountain and no-mountain experiments reflect the effects of topography only in the presence of zonally uniform SSTs. Given that the precipitation rate in eastern South America is particularly sensitive to the SST distribution, one might suspect that the response to topography would be different in the presence of realistic SSTs. Two additional GCM experiments shed some light on this issue. The first is similar to the SST experiment but has different land surface conditions and 30 vertical levels. The second experiment is identical to the first, except with no topography. Thus, a comparison of the two offers a clean diagnosis of the effects of topography in the presence of realistic SSTs.

These two experiments confirm that continentality (topography) is primarily responsible for enhanced precipitation rates in the SACZ and Amazon regions (Central and Southern Andes). The fundamental role of continentality for enhancing precipitation rates in the Northern Andes is also confirmed. However, Northern Andean topography more strongly concentrates the precipitation into a well-defined maximum than in the case with zonally uniform SSTs. In addition, the effects of topography on precipitation in the Amazon and SACZ are different from what is indicated by a comparison of Figs. 2c and 2d. While we do not focus on this issue in the present study, the lack of agreement indicates the need for further research on the effects of topography on precipitation in eastern South America.

8. Summary and conclusions

The observed January precipitation climatology of South America reveals five reasonably well-defined precipitation maxima, which occur in the Amazon region, the Northern, Central, and Southern Andes, and
the SACZ. We have modeled this regional precipitation event in a perpetual-January, R30, GCM experiment, where the only structure at the surface is the presence of continents, topography, and realistic SSTs. Despite the lack of structure in other land surface conditions, the model is able to simulate reasonably well the five regional precipitation maxima over South America. Two other GCM experiments are the relative roles of continentality, topography, and SSTs in generating this structure in the precipitation field. Analysis of the atmospheric water vapor budget provides a fuller understanding of these mechanisms by relating the precipitation enhancements to features of the large-scale circulation. Comparisons with observed height, wind, and moisture fields are used to validate the mechanisms identified in the model analysis.

In the GCM, the presence of the South American continent (without topography and with zonally uniform SSTs) is responsible for establishing precipitation maxima in the Amazon, SACZ, and Northern Andes regions in association with the summertime thermal low. Enhanced precipitation in these regions is associated with low-level wind convergence maxima that are directly related to the interaction of the continental low with the South Atlantic high and the northeasterly trade winds. Precipitation in the SACZ is also enhanced by strong southward advection of moisture between the continental low and South Atlantic high. Additionally, moisture flux convergence by transient eddies is significant in the SACZ.

Topography introduces orographic precipitation maxima in the GCM along the eastern and western flanks of the Central and Southern Andes, respectively, where low-level winds impinge on the mountain barrier. Strong precipitation at higher elevations of the Central Andes is associated with locally driven small-scale convergence, not with orographically induced mechanical uplift. In the Northern Andes orographic precipitation is weak (perhaps too much so), but orography helps to define the precipitation maximum. The effects of orography on precipitation in the Amazon and SACZ are regarded as inconclusive and worthy of further investigation. Considering that the GCM's precipitation field has shown a sensitivity to African topography, it would be especially interesting to distinguish the effect of global topography on the precipitation in the Amazon and SACZ from the effect of just the Andes.

The five regional South American precipitation maxima are simulated in the absence of zonal structure in SSTs and geographic structure in South American land surface conditions. This suggests that continentality and topography are more fundamental for establishing these precipitation maxima. However, the positions and magnitudes of some of these maxima are quite sensitive to SSTs and land surface conditions, particularly in the Amazon and SACZ. Other studies have also shown that precipitation in eastern South America is sensitive to SSTs, especially in connection with droughts in northeastern Brazil (e.g., Mechoso et al. 1990). Similarly, numerous studies of Amazonian deforestation have demonstrated connections between land surface conditions and precipitation in the Amazon region (e.g., Lean and Rowntree 1993).

Comparisons between modeled and observed precipitation, moisture flux convergence, and low-level winds show good agreement in the Amazon region and SACZ, suggesting that the modeled effects of continentality and the thermal low on precipitation in these two regions are robust. However, some significant discrepancies between the model and observations are present in the three Andean regions. In the Central Andes, for example, the modeled precipitation maximum is much stronger and larger in extent than the observed, while in the Southern Andes, the modeled precipitation is too weak. However, the low-level wind and moisture flux convergence fields in these two regions are modeled reasonably well, suggesting that the proposed mechanisms are appropriate. In the Northern Andes the shallow, onshore westerlies are stronger and more extensive in the model than in the observations, and orographic precipitation is only weakly simulated.

Given that the model's representation of the Andes is (by a factor of 2 or more) too low in elevation and too broad in longitudinal extent, it is certainly possible that this is responsible for some of the above discrepancies. For example, raising the elevation of the Andes in the model increases the orographic precipitation in the Northern and Southern Andes, thereby improving the simulation. However, the simulation is worsened in the Central Andes, where it seems that the modeled excessive precipitation may be more related to the broadness of the mountains (primarily a limitation of the model resolution). Since filtering the topography acts to lower and broaden the Andes, it might seem that such a technique is not advantageous. However, it has been our experience and that of others (e.g., Navarra et al. 1994) that the overall simulation of the precipitation field with filtered topography is still a considerable improvement over that with unfiltered topography.

The present study was performed with very simplified boundary conditions. Even so, the favorable comparison with observations and results from more realistic GCM experiments suggest that the conclusions presented here are representative of the real climate. Nevertheless, it would be interesting to perform studies similar to the present one but with different models to test the sensitivity to various model characteristics. For example, one might wonder how different parameterizations of cumulus convection or different treatments of the boundary and surface layers influence the results. The effects of diurnal solar forcing and different cloud treatments are also not investigated here. Finally, a complete study of the regional South American precipitation climatology would have to consider the influence of seasonality. We have only dealt here with the
month of January since this is the height of the rainy season in many (but certainly not all) regions of South America. Considering the marked seasonal changes that take place in SST, the continental low, and the associated large-scale circulation, it is reasonable to expect that the relative effects of continentality, topography, and SSTs would vary seasonally.

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APPENDIX

Derivation of the Transients Term

The exact form of the transients term given in (3) can be found by subtracting (2) from (3):

\[ \bar{T} = - \frac{1}{g \rho_w} \int_0^1 \left\{ \frac{\partial}{\partial \sigma} \left[ p_s \left( \frac{\partial q}{\partial t} + \frac{\partial q}{\partial \sigma} \right) \right]_{\rho_s} \right\} d\sigma. \quad (A1) \]

The purpose of this appendix is to show the approximations involved in going from (A1) to its approximate form:

\[ \bar{T} \approx - \frac{1}{g \rho_w} \nabla \cdot \left[ \int_0^1 \left( p_s q - \frac{\partial q}{\partial \sigma} \right) d\sigma \right]. \quad (4) \]

Expanding \( \nabla_{\rho_s} \cdot q u_{\rho_s} \) into its horizontal and vertical components and transforming to sigma coordinates gives

\[ \nabla_{\rho_s} \cdot q u_{\rho_s} = \frac{\partial}{\partial x} (qu)_s + \frac{\partial}{\partial y} (qv)_s \]

where \( \nabla_{\rho_s} = (\partial/\partial x + \partial/\partial y, \partial/\partial \sigma) \), and \( u_{\rho_s} = (u, v, \partial q/\partial t) \). From (A2),

\[ p_s \left[ \nabla_{\rho_s} \cdot q u_{\rho_s} + \frac{\partial q}{\partial t} \right] = \nabla \cdot \left( p_s q u_{\rho_s} \right), \quad (A3) \]

where it has been assumed that \( \partial q/\partial \sigma(p, q) = 0 \) (by definition of climatological mean).

In a fashion similar to the derivation of (A3), it can be shown that

\[ \bar{p} \nabla_{\rho_s} \cdot q \bar{u}_{\rho_s} = \nabla \cdot \left( \bar{p} \bar{q} \bar{u}_{\rho_s} \right) \]

\[ + \frac{\partial}{\partial \sigma} \left\{ \bar{q} \left[ u' \frac{\partial p'}{\partial x} + v' \frac{\partial p'}{\partial y} + \frac{1}{\sigma} \left( \sigma' p' \right) \right] \right\}, \quad (A4) \]

where \( \sigma = \partial \sigma/\partial t \) and primed quantities indicate deviations from the climatological mean (e.g., \( u' = u - \bar{u} \)). Incorporating (A1) and (A3) into (A4) allows the transients term to be written as

\[ \bar{T} = - \frac{1}{g \rho_w} \int_0^1 \left\{ \frac{\partial}{\partial \sigma} \left[ \frac{\partial q}{\partial t} + \frac{\partial q}{\partial \sigma} \right]_{\rho_s} \right\} d\sigma. \quad (A5) \]

Performing the integral over sigma for terms that involve \( \partial q/\partial \sigma \) and making use of the boundary condition, \( \sigma = 0 \) at \( \sigma = 1 \), (A5) becomes

\[ \bar{T} = - \frac{1}{g \rho_w} \nabla \cdot \left[ \int_0^1 \left( p_s q - \frac{\partial q}{\partial \sigma} \right) d\sigma \right] \]

\[ + \frac{\bar{q}}{g \rho_w} \left[ u' \frac{\partial p'}{\partial x} + v' \frac{\partial p'}{\partial y} \right], \quad (A6) \]

where the subscript, \( s \), indicates evaluation at the surface (\( \sigma = 1 \)). Equation (4) is, therefore, obtained from (A6) if

\[ u' \frac{\partial p'}{\partial x} + v' \frac{\partial p'}{\partial y} \approx 0. \quad (A7) \]

It is not the intention here to "prove" (A7) but to simply note that the GCM results indicate such an approximation is, in fact, valid.

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