The Effect of Tropical Atlantic Heating Anomalies upon GCM Rain Forecasts over the Americas

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ABSTRACT

Severe droughts occurred over eastern sections of North America and central sections of South America in 1986 and 1988. We summarize data suggesting that both periods were characterized by above-normal tropical Atlantic sea surface temperatures and convection, and investigate the response of a general circulation model to positive heating anomalies in the tropical Atlantic sector. An eight-case control ensemble of 30 day global predictions is made starting from the atmospheric state observed on 1 January of each year from 1977 through 1984. The same eight cases are integrated in a second experimental ensemble that is identical to the first control ensemble, except that a heating term is added to the thermodynamic equation in a region centered at 30°W, 6.6°N. This is intended to simulate the latent heating of enhanced tropical Atlantic convection. The third ensemble is identical to the second, except the heating is centered at 6.6°S.

Both heated ensembles produce reductions of forecast precipitation over most of North and South America, but these appear to have greater statistical significance over North America. Here the greatest precipitation reductions are forecast over the southern and eastern United States, and this response does not change substantially between the two experiments. The South American response is more sensitive to the placement of the heating anomaly. When the anomaly is located north of the equator, drying occurs over northeast Brazil; meanwhile this region receives increased rainfall when the anomaly is located south of the equator. Both experiment ensembles display a region of reduced rainfall over the Andes Mountains, and over southern portions of Brazil. However, only the former region is statistically significant above the 95% confidence level. The present usage of real initial data and an ensemble of cases permits us to draw quantitatively meaningful estimates of the time scale of response and case-to-case variability. For presently tested cases, the South American response is evident by day 5, but exhibits substantial intersample variability, and the North American response is fully established by day 10, and exhibits less intersample variability. The model drying effects can be explained only partly by enhanced local subsidence; much of the rainfall reduction appears to be related to a reorientation of the synoptic scale wave pattern in which the lower tropospheric circulation is unfavorable for water vapor inflow from source regions over the tropical Atlantic and Amazon Basin.

1. Introduction

Prolonged drought conditions are of particular practical concern in those regions of the world in which agriculture is well developed and dependent upon recurring rainfall before and during the growing season. The two major regions of the western hemisphere possessing these characteristics are the agricultural belts of the central and eastern United States, and the region of South America comprising southern Brazil, northern and central Argentina, Uruguay and Paraguay. These areas ordinarily have sufficient local precipitation from winter through summer to make it impractical to construct irrigation systems based on distant water sources, and to consequently leave them vulnerable even to the occasional droughts that may occur in only rather restricted regions.

The problem has received increased attention in recent years because of extensive droughts and crop failures during 1986 over most of the southeastern United States (Halpert and Ropelewski 1987), and during 1988 over much of the United States (Trenberth et al. 1988). Although they have been less well publicized in North America, severe drought also occurred over

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Figure 1 reproduces the precipitation anomaly charts for the period of December 1985 through February 1986 (i.e., winter 1986) from Arkin and Janowiak (1987). These display extensive tongues of abnormally dry conditions over the southeastern United States and Central regions of South America. Figure 2 reproduces the 200 mb wind anomaly for January 1988 (taken from Kousky 1988). This pattern is similar to the anomaly for the 1986 winter (Arkin and Janowiak 1987) in the Northern Hemisphere, but the midlatitude westerly anomaly over South America and the South Atlantic is somewhat weaker in 1986. The wind anomalies at 850 mb are similar in the two periods as well (Fig. 6b of Arkin and Janowiak 1987; and Fig. 25 of Kousky 1988). Both display accentuated subtropical anticyclones above the North and South Atlantic and imply enhanced low-level convergence over the tropical Atlantic Ocean.

Outgoing longwave radiation (OLR) maps display anomalously low values over the tropical Atlantic in both years (Fig. 4b of Arkin and Janowiak 1987; and Fig. 29 of Kousky 1988). These imply increased convection over the tropical Atlantic, in consistency with the increased low-level convergence found there. The sea surface temperature anomalies (SSTA) are also positive over the tropical and South Atlantic Ocean in both years (Fig. 23 of Arkin and Janowiak; and Fig. 22 of Kousky 1988).

The association of rainfall in the tropical Atlantic sector with anomalous circulation (and by inference, rainfall) in extratropical latitudes is not new. Namias (1972) found correlations of rainfall in the northeast of Brazil with the subtropical anticyclone and higher latitude cyclones in data studies on time scales as short as five days. Moura and Shukla (1981) have studied these correlations in extended integrations of a general circulation model initialized with idealized data.

However, in more recent years, there has been an even greater research emphasis upon correlations of tropical and extratropical weather with SSTA of the tropical Pacific Ocean associated with the El Niño, Southern Oscillation (ENSO). Trenberth et al. (1988) and Kousky et al. (1984) present recent observational studies of ENSO related signals.

Negative SSTA anomalies were observed over much of the equatorial east Pacific during the droughts of 1986 and 1988, with the latter becoming notable during the northern spring and summer. Buchmann et al. (1986, 1988) have performed numerical forecast experiments designed to study the impact of Pacific tropical heating anomalies upon South American rainfall and Paegle et al. (1987) have used the same approach to search for the influence of such anomalies upon North American weather. Those calculations imply that a relatively cool eastern tropical Pacific Ocean reduces rainfall over subtropical portions of eastern South America, and diminishes cyclonic activity over North America. Consequently, two of the elements contained in the observational data during recent drought years may be partly explained by cooler than normal conditions of the tropical east Pacific.

Those studies leave two major questions with regard to tropical–extratropical interactions. First, they did not address the precipitation response in the extratropics of North America. However, we do not expect this to be statistically significant because the model circulation response had marginal statistical significance over the eastern half of the United States (Paegle et al. 1987) where the droughts have been most severe. Second, they did not study the relative role of tropical Atlantic heating modifications, but did conclude (Buchmann et al. 1988) that the tropical east Pacific SST modification experiments show little agreement.
with the North Atlantic teleconnection pattern observed by Namias (1972), particularly on the shorter (five day) time scales covered in the latter study. Consequently, our prior results suggest that the Atlantic and east-American anomalies observed in 1986 and 1988 may have sources which are independent of Pacific Ocean influences.

It is the purpose of the present experiment to further probe this question. We presently use a newer version of the general circulation model previously employed by Buchmann et al. (1986, 1988) and by Paegle et al. (1987) in an eight case ensemble of 30 day, real data forecasts wherein the tropical Atlantic heating is modified to simulate enhanced convective latent heating observed there during 1986 and 1988. The goal is to inquire whether the meridional structure of the response anomaly is more in agreement with the observed anomaly pattern of the drought years than was the case with our prior Pacific heating experiments. Partial reasons to expect a positive conclusion are provided in a study by Moura and Shukla (1981) using a different model, and a more limited sample. Their forcing was provided by heating mainly north of the equator, rather than south of the equator, which is more common in the northern winter.

In the present study, an eight case ensemble initialized with data for the first day of each year from 1977 to 1984 is used, in order to obtain statistically reliable results. If the model were perfect this would be tantamount to an “observed” sample of eight events rather than just the two observed cases of 1986 and 1988. This sample is further amplified by repeating the eight case ensemble separately for cases heated just north of the equator and for cases heated just south of the equator.

Section 2 describes the model and outlines the experiment design. Sections 3 and 4 present the response of the precipitation and circulation, respectively, to heating of the tropical Atlantic. Section 5 describes statistical significance of the results, and section 6 summarizes conclusions.

2. Model and experiment design

The model used for this study is the National Center for Atmospheric Research (NCAR) general circulation model (GCM). This is a spectral global model that employs sigma coordinates, is initialized with real data, and run at a horizontal rhomboidal spectral truncation retaining 15 waves. The model (CCM) is an improved version of the community climate model (CCMO) employed in our earlier studies of South American rainfall (Buchmann et al. 1986, 1988; and Paegle et al. 1987).

The present model uses 12 layers in the vertical and has several substantial improvements in the radiation code. The new radiation calculations retain a solar zenith angle dependence in albedo specification. The algorithm for absorption of solar radiation by water vapor and oxygen has been improved, and the water vapor absorption of the longwave radiation has been replaced by a nonisothermal emissivity scheme. Calculations of carbon dioxide and ozone absorption have also been improved, and the emissivity of stratiform clouds is now determined by their liquid water content.

Other improvements include vertical finite difference approximations that have better energy conserving properties, and retention of frictional dissipation of kinetic energy to heat in the thermodynamic equation. The second-order lateral diffusion previously used in the troposphere has been replaced by a more scale-selective fourth-order diffusion. However, second-order diffusion is retained in the stratosphere. The diffusion includes corrections to constrain its principal effect in the pressure surface rather than the sigma surface, and to thereby minimize spurious vertical diffusion associated with steep mountains.
The vertical eddy mixing coefficient now depends upon local shear and stability throughout the atmosphere, and imposes an effective turbulent adjustment that obviates the need for a separate dry convective adjustment in the troposphere. As a column becomes unstable, the vertical diffusion coefficient becomes sufficiently large that the diffusion relaxes the column back to adiabatic stratification. Furthermore, condensation now occurs at 100% relative humidity instead of 80% as in the previous model, and evaporation is decreased over deserts and grasslands. The surface drag coefficient now depends upon stratification, the equation of state uses the virtual temperature rather than the actual temperature, and the specific heat at constant pressure varies with moisture. Further details on these and other aspects of the model are provided by Williamson et al. (1987).

The experiments of the present study are composed of three ensembles, each consisting of eight cases initialized from eight different dates starting on 1 January of each year from 1977 to 1984. The first ensemble is the control, in which the model is run in an unmodified form for 30 days for each of the eight separate cases. In the second and third ensembles, a heating term is added to the thermodynamic equation. In the case of the second ensemble, this heating term maximizes at 6.6°N, 30°W, and in the case of the third ensemble the heating is strongest at 6.6°S, 30°W. The term decreases radially outward with a Gaussian profile that has a half width of approximately 8° in latitude and 16° in longitude. It is strongest at 400 mb in both cases, with a local maximum of 8°C/day, and a column average of 5°C/day. This corresponds to the latent heating associated with a precipitation rate of approximately 1.7 cm/day.

The imposed heating induces rising motion that locally enhances the rainfall and adds an additional latent heating that approximately equals the imposed source. The net result is a local heating that is more characteristic of rather active individual events than of seasonal averages. Consequently, the model is modified more strongly than the actual atmosphere. One reason for this is to maintain consistency with the degree of heat forcing applied in our earlier studies that focused upon the effect of relatively strong, short-term tropical heat enhancements over the Pacific Ocean. Another reason is that general circulation models commonly respond rather little to weaker, possibly more realistic long term heating, and the forcing is consequently increased with respect to observed values, [e.g., compare Moura and Shukla’s (1981) imposed Atlantic modifications to observed values, and Blackmon et al.’s (1983) Pacific modifications to observations]. However, our best justification is that the seasonal latent heating anomalies are composed of individual events of larger magnitude, and these larger magnitudes may be appropriate for present subseasonal integration periods.

We have examined the precipitation rates and latent heating rates produced by this version of the NCAR model in perpetual season integrations. For one month averages, the maximum local precipitation rates are approximately 2.5 cm/day, and the maximum mid-tropospheric heating rates are about 11°C/day, (not shown). Consequently, although the presently implemented forcing is rather strong when compared to typical tropical activity over the Atlantic Ocean, it is not unreasonably high compared to other tropical events simulated by the model.

3. Precipitation modifications

The present section presents the response of the precipitation forecasts. Time-averaged precipitation and its case-to-case variability are discussed in more detail by Buchmann et al. (1986) for both control and (east Pacific) heated cases. The principal conclusions of that study conform with present results; i.e., control integrations for the northern winter produce averaged rainfalls that maximize in the interior of the Amazon Basin over South America. Over North America, the heaviest rainfall is predicted to occur near the coasts.

Figure 3 presents the 30-day averaged response of the rainfall over South America for each of the experimental cases in ensemble 2 (heating maximum at 6.6°N). Each of these cases displays a substantial decrease of precipitation around northeast Brazil, in conformity with Moura and Shukla’s (1981) study. However, there is no systematic signal in the central portion of South America. Here, the influence of tropical North Atlantic heating anomalies varies substantially from case to case.

The situation is rather different for the case of ensemble 3 (heating maximum at 6.6°S). In this case (Fig. 4) the precipitation over the extreme eastern tip of Brazil is generally enhanced by the tropical latent heating maximum, and it is diminished south of this region in most cases. This becomes more evident in statistical analysis (section 5).

We conclude that the details of the precipitation response to tropical Atlantic heating anomalies depend upon the latitudinal position of the heating anomaly, and only tropical heating in the South Atlantic provides consistent rainfall suppression in this region. This agrees with the observation that the largest Atlantic sector SSTA during 1986 and 1988 occurred south of the equator.

Figure 5 displays the precipitation response over North America in ensemble 2 (heating maximum at 6.6°N). Six of the eight cases show rainfall decreases over the southeastern United States in the tropically heated experiments. A similar conclusion characterizes the results of ensemble 3, (heating maximum at 6.6°S). These are shown in Fig. 6, and six of the eight cases display rainfall reduction in the southeastern sections of the United States. The composite of ensembles 2
Fig. 3. Time averaged precipitation response in ensemble 2 (experiment-control) for (a) 1977, (b) 1978, (c) 1979, (d) 1980, (e) 1981, (f) 1982, (g) 1983, (h) 1984. Contour interval equals $1.5 \times 10^{-4}$ m s$^{-1}$. Dashed contours are negative.
Fig. 4. As in Fig. 3, for ensemble 3.
Fig. 5. As in Fig. 3, for Northern Hemisphere. Contour interval equals $6 \times 10^{-9}$ m s$^{-1}$. 
Fig. 6. As in Fig. 4, for Northern Hemisphere. Contour interval equals $6 \times 10^{-9}$ m s$^{-1}$. 
and 3 indicates that 12 of 16 cases of tropical Atlantic heating modifications produce distinct drying in the southeastern portion of the United States, and in these cases the pattern is rather similar to the observations of Fig. 1a. The details of this picture are apparently not very sensitive to the position of the heating anomaly with respect to the equator. Although the maximum precipitation decrease is in the southeast section of the United States, there are general decreases over most of the country. The decreases located further north may have even greater relative impact than those in the southern areas, because the control precipitation of ensemble 1 (not shown) is smaller toward the north. This point is emphasized by statistical analysis (section 5).

Figure 7 presents the ensemble averaged time evolution of the precipitation response over South America for the second ensemble (heated at 6.6°N). The response at day 5 is almost as strong as it is at later days, implying that an accurate short to medium range forecast of precipitation over much of South America may require an accurate forecast of conditions over the tropical Atlantic Ocean. This conclusion is not particularly surprising. Figure 8 displays the ensemble averaged time evolution of the precipitation response over North America for the same ensemble. A weak negative signal is perceived at day 5, and becomes fully established by day 10, implying the importance of conditions over the tropical Atlantic for accurate medium range rain prediction over North America.

4. Circulation responses

It is of interest to compare the degree to which the precipitation modifications described in the previous section are directly explained by modifications of the vertical motion, as opposed to other possible influences such as moisture transport. Figure 9 displays the ensemble averaged time evolution of the vertical motion response over South America for the second ensemble. It indicates areas of alternating sign, but generally the regions of increased subsidence (positive values in Fig. 9) correlate well with regions of decreased precipitation (negative values in Fig. 7) over the eastern tip of Brazil. Thus, it appears that, here, the modifications of the vertical motion field correlate rather directly with the precipitation.

In other regions, this correlation is not equally evident, and it is particularly weak in more remote regions. Figure 10 displays the ensemble-averaged time evolution of the vertical motion response over North America for the same ensemble. Comparing this with the precipitation modification of Fig. 8 shows little detailed correlation. The vertical motion response displays enhanced subsidence in a northeast to southwest line along the East Coast, while the precipitation changes feature a more nearly east–west line of reduction.

It is likely that other effects such as modification of horizontal moisture transport also affect rainfall response. This possibility is supported in the response of the horizontal circulation to the heating changes. Figures 11 and 12 display the time evolution of the 700 mb geopotential height response for ensembles 2 and 3 respectively. In both cases, at almost all displayed times, there is a general enhancement of the relative northerly flow over eastern North America that may be inferred from the increased heights centered approximately over Hudson Bay, and the lowered heights centered approximately over the central Atlantic Ocean. The implied northerly flow modifications over the eastern United States would probably allow less moist air to penetrate this region from the Gulf of Mexico and the Atlantic.

An analogous situation occurs over South America for ensemble 3 (Fig. 12), in that relatively colder, and probably relatively drier southerly winds are implied by the gradient of the height field change at most times. However, a similar change is not equally obvious for ensemble 2, (Fig. 11).

Figures 11 and 12 help to explain the relatively drier conditions over North America in terms of relatively less effective moisture transport of the lower tropospheric air stream over the most strongly affected region. The North Atlantic anomaly patterns of these figures are consistent with the anomaly correlations found by Namias (1972) between the rainfall over northeast Brazil and the 700 mb height pattern over the North Atlantic. Figure 4 of Namias' (1972) study indicates lower heights over the central North Atlantic, and higher heights over northeast Canada and northern portions of the North Atlantic during relatively dry periods over northeast Brazil.

However, the response patterns of Figs. 11 and 12 are not particularly consistent with the lower tropospheric flow deviations that occurred over the Atlantic Ocean in 1986 and 1988. As mentioned in the Introduction, those displayed anomalously strong anticyclonic circulations over the Central Atlantic Ocean in both hemispheres.

Figure 13 shows the ensemble and time-averaged 200 mb wind responses. Each case displays enhanced anticyclonic flow in the subtropics of the Atlantic region, straddling the heated tropical sector. These are associated with enhanced westerly flow in midlatitudes above both the North and South Atlantic. This response is similar to the observed anomaly of the drought year displayed in Fig. 2, although the westerly response maximum is stronger and situated somewhat equatorward and eastward of the observed circulation anomaly.

5. Statistical significance of rainfall response

The rainfall changes (Figs. 5 and 6) show that the precipitation decreased over the southeastern United States in 12 cases out of the total of 16 experiments in
Fig. 7. Time evolution of precipitation response averaged over ensemble 2 for (a) day 5, (b) day 10, (c) day 15, (d) day 20, (e) day 25, (f) day 30. Contour interval is $4 \times 10^{-8}$ m s$^{-1}$. 
ensembles 2 and 3. However, the response signal was not equally clear over South America, (Figs. 3 and 4), and the time evolutions of the precipitation responses over North America display several instances with reversals from the typical pattern. Although the ensemble-averaged time evolution and the case-by-case differences within each ensemble are useful to describe the response time scale and internal variability, respectively, the inferences regarding the averaged response are rather subjective.

The large temporal variability of the response in Figs. 7 and 8 suggests that the results at any particular time may have relatively small statistical reliability. This is confirmed in calculations of the $T$ statistic, which
FIG. 9. Time evolution of vertical motion response averaged over ensemble 2 for (a) day 5, (b) day 10, (c) day 15, (d) day 20, (e) day 25, (f) day 30. Contour interval equals $3 \times 10^{-4}$ mb s$^{-1}$. 
quantifies the probability that the difference between the forced and control ensembles did not occur merely as a sampling fluctuation of a randomly varying process. This probability varies with the number of degrees of freedom, and an overall measure is obtained by a "significance" estimate. However, time-averaged responses are more significant.

Figure 10. As in Fig. 9, for Northern Hemisphere.

Figure 14a presents the time-averaged response of the second ensemble over South America. This field is negative almost everywhere over South America, and the response is especially pronounced over eastern and western sections of Brazil. The $T$-statistic for this case is displayed in Fig. 14b; this shows values approaching 10 in the center of the heated area, and 4 over South
Fig. 11. Time evolution of geopotential height response averaged over ensemble 2 for (a) day 5, (b) day 10, (c) day 15, (d) day 20, (e) day 25, (f) day 30. Contour interval equals 20 m.
Fig. 12. As in Fig. 11, for ensemble 3.
America. The lowest contour value of the significance analysis (Fig. 14c), equals 95%, and this is exceeded over most of the heated region, and in an area over western Brazil.

Figure 15a presents the time-averaged response of ensemble two over North America. This is negative over most of the southern and eastern United States; the corresponding T-test values approach 4 over the eastern United States (Fig. 15b); and the significance level exceeds 99% over a fairly large section of the eastern United States (Fig. 15c). The North American precipitation decrease is statistically more significant than the central South American decrease in the present 8-case ensemble of tropical Atlantic heating centered north of the equator.

The averaged response of ensemble 3 (South Atlantic heating) does not present major differences in the statistics. Figure 16a displays the time-averaged response over South America. This indicates a region of decrease over southeastern Brazil that was also surmised from the individual charts of section 3. However, the T-statistic is not particularly large here (Fig. 16b). Conse-
Fig. 14. (a) Ensemble and time-averaged precipitation response for ensemble 2. Contour interval is $5 \times 10^{-4}$ m s$^{-1}$. (b) T-statistic for response depicted in Fig. 14a. Contour interval is 2. (c) Probability that the response depicted in Fig. 14a is statistically significant. Contour interval is 1%, and only values of 95% and higher are analyzed. Other plotted values give local maxima and minima.

Fig. 15. As in Fig. 14, for ensemble 2 and Northern Hemisphere. Contour interval is $5 \times 10^{-4}$ m s$^{-1}$ in (15a).

The averaged response of ensemble 3 over North America is rather similar to the averaged response of ensemble 2. Both show substantial reductions over the eastern portions of the United States, (Fig. 17a). Here the $T$-statistic exceeds 3, and the number of degrees of freedom is sufficient to produce an extensive area wherein the significance level exceeds 99%. We con-
clude that large anomalies in the tropical Atlantic latent heating are significantly correlated with the rainfall over eastern portions of the United States in 30-day integrations of the NCAR GCM. Furthermore, this correlation is not particularly sensitive to detailed placement of the tropical heating anomaly.

6. Conclusions

The original motivation of this study was to expand the experimental information base regarding external
influences upon the rainfall of South America. Our prior studies (Buchmann et al. 1986, 1988; Paegle et al. 1987) had emphasized the influence of heating anomalies of the tropical Pacific Ocean upon the Amazon Basin, and concluded that positive (negative) heating anomalies decrease (increase) the rainfall over northern sections of tropical South America and increase (decrease) it over southern portions of this region. However, the most recent study (Buchmann et al. 1988), which included statistically quantified measures of reliability for rainfall response, suggested that these conclusions were barely significant at the 95% level of confidence, even in the rather strong modifications of the experimental conditions imposed in those studies.

This provided partial motivation for the present investigation, which was designed to study the proposal that rainfall in the central sections of eastern South America is reduced by excess heating in the tropical Atlantic Ocean. This suggestion has apparently not been previously advanced within modeling studies. However, the strong correlations found by Namias (1972) between South American rainfall and the Atlantic circulation, the corroboration of those results by Moura and Shukla (1981), and the correspondence of above normal tropical Atlantic heating with the droughts of central regions of South America during 1985–1986 and 1988 are suggestive of this possibility.

One of the reasons to investigate North American as well as South American responses was that the lower-tropospheric Atlantic circulation anomalies of each of the last two droughts appeared to be rather similar in both Northern and Southern Hemispheres. In the case of 1988, this similarity is also present in the upper troposphere (Fig. 2).

Unexpectedly, in the present experiments, the response of the rainfall predictions over eastern North America is statistically more significant than it is over South America in both experiment ensembles (section 5), even though the latter region is much closer to the area of greatest experimental forcing. Furthermore, the North and South American signals in both the precipitation (section 3) and the circulation (section 4) are fully apparent by day 10 of the experiment. This suggests, in conformity with our prior tropical Pacific heating experiments, that tropical heating affects the weather in the extratropics in both hemispheres on a time scale that is clearly relevant to medium-range forecasting. It follows that accurate medium-term, midlatitude forecasts with the present model may depend upon accurate tropical forecasts. Time-averaged circulation statistics suggest that subsidence and reduced water vapor transport both contribute to the rainfall suppression.

Perhaps the most important circulation responses to the heating changes are the enhanced equatorward flows predicted at low levels around eastern sections of both North and South America. These modifications would allow relatively less moisture to penetrate the United States and east-central South America from the principal sources located over the Gulf of Mexico and Amazon Basin, respectively.

Of course, each of these conclusions depends upon various model assumptions. Foremost of these may be that the assumed tropical Atlantic heating is well above that suggested from the seasonally averaged anomalies of 1986 and 1988. As mentioned in section 2, this large heating was used partly because the current 30-day forecasts have subseasonal duration and partly for conformity with the overestimations of tropical heating anomalies in previous experiments.

The relatively crude resolution at R15 implies further limitations. However, even a modest doubling of resolution in the horizontal direction would require an eightfold increase in the calculation time, because the number of horizontal modes would increase by a factor of four, and numerical stability would probably require a time step about half of that needed at R15. Much higher resolution is difficult to afford in the present experimental design, which required a total of 24 different 30-day integrations.

The possibility, nevertheless, remains that the present conclusions, as well as those presented in the previously quoted studies, may not represent the actual atmospheric response to the important American droughts of the current decade. Recently, there have been suggestions that global warming due to greenhouse effects associated with increases in the atmosphere of certain industrially produced effluents may increase the probability of North American drought. Another recent study (Trenberth et al. 1988) suggests that SST anomalies in the eastern tropical Pacific may be responsible for the North American drought of 1988. It is also likely that droughts may have positive local feedback effects that require more careful simulation of surface processes and hydrology. The relative roles of those processes compared to presently studied mechanisms remain to be delineated in further observational and modeling investigations.

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