NOTES AND CORRESPONDENCE

Further Experiments on the Effect of Tropical Atlantic Heating Anomalies upon GCM Rain Forecasts over the Americas

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25 March 1994 and 15 August 1994

ABSTRACT

A series of real-data experiments is performed with a general circulation model to study the sensitivity of extended range rain forecasts over the Americas to the structure and magnitude of tropical heating anomalies. The emphasis is upon heat inputs over the tropical Atlantic, which have shown significant drying influences over North America in the author's prior simulations. The heating imposed in the prior experiments, that is, shown to be excessive by a factor of 2, is compared with the condensation heating rates that naturally occur in the forecast model. Present experiments reduce the imposed anomaly by a factor of 3 and also impose sea surface temperature decreases over the eastern tropical Pacific Ocean. The new experimental results are in many ways consistent with the author's prior results. The dry North American response is statistically more significant than the South American response and occurs at least as frequently in the different members of the experimental ensembles as in our prior experiments. The drying effect is accentuated by the presence of east Pacific cooling, but this does not appear to be the dominant influence. Over tropical South America, the Pacific and Atlantic modifications produce compensating influences, with the former dominating, and allow increased rainfall over the Amazon Basin.

1. Introduction

The purpose of this paper is to describe the sensitivity of the response of a general circulation model to variations in the structure and magnitude of tropical forcing. The study builds upon a series of real-data experiments imposing tropical heating modifications within the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM). Zhang (1985), Buchmann et al. (1986), Paegle et al. (1987), Buchmann et al. (1989), and Buja (1989) describe experiments in which the tropical heating of the east Pacific is modified within integrations of 10–36 d duration. Buchmann et al. (1990) and Buja (1989) study the effect of varying tropical Atlantic heating in 30-d predictions.

The principal conclusions of those investigations were that strong tropical latent heating anomalies influence regional tropical and subtropical circulations within 5 days and modify extratropical circulations after approximately 5–10 days. Global responses are evident after 30 days. In particular, modifications of tropical Atlantic heating have surprisingly strong and repeatable effects upon rainfall forecasts over North America (Buchmann et al. 1990). Those experiments used somewhat unrealistically strong tropical heating modifications. Two of our present goals are to investigate the natural variability of tropical precipitation and latent heating rates within the NCAR CCM and to study the model response to more realistic tropical heating anomalies suggested by the model variability.

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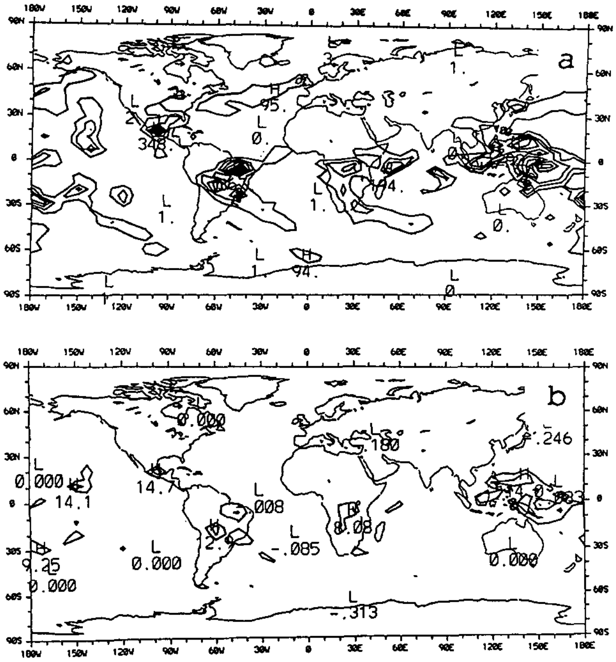

FIG. 1. (a) Precipitation rate (in m s^{-1}) averaged over days 620–630 in a perpetual January, 1200-d integration of the NCAR CCM. Contour interval is 5×10^{-8} m s^{-1}. (b) Condensation heating rate (in °C d^{-1}) at 500 mb averaged over days 620–630 in a perpetual January, 1200-d integration of the NCAR CCM. Contour interval is 5°C d^{-1}.

Selection of different real-data cases allows statistical evaluation of response in forecasts that retain some of the variability characterizing the actual atmosphere. The 36-d forecasts made by Buja (1989) are sufficiently short that their temporal average has not drifted completely into the model climate, and sufficiently long that they may be relevant to outlooks of monthly trends. The relatively short durations of the integrations may have justified the imposition of large heating anomalies, peaking at 8°C d^{-1} in the midtroposphere. Such heating rates are probably excessive with respect to latent heating occurring naturally in the NCAR CCM on seasonal timescales, and it is not clear how relevant they may be for shorter periods. This complicates the interpretation of the earlier results. The first goal of the present study is to analyze the precipitation and latent heating that occur naturally in the NCAR CCM (section 2). The peak monthly averaged rainfall and heating rates in perpetual January integrations are on the order of 2.5 cm d^{-1} and about 11°C d^{-1}, respectively. These values are larger than those imposed in our prior experiments.

The second goal of the present investigation is to study the sensitivity of prior conclusions to the strength of the tropical forcing. This is done for cases of tropical Atlantic heating, which have produced a particularly interesting remote response. Buchmann et al. (1990) find that increased tropical Atlantic heating reduces model rainfall in 12 out of a total of 16 experimental cases over eastern North America and increases rainfall here in only two of the cases. The drying effect is statistically significant at the 95%–99% level over much of the eastern United States. These results are consistent with observations during 1986 (Fig. 1 of Buchmann et al. 1990, and references therein) and 1988 (Fig. 2b of Trenberth and Branstator 1992, and references therein), which show decreased precipitation over sections of eastern North America and above-normal convection over the tropical Atlantic Ocean. Section 3 describes the response of the NCAR CCM-I when the magnitude of the imposed tropical heating is reduced to one-third of the values used by Buchmann et al. (1990). The resulting heat input is less than the values produced naturally by the model. The integrations also include modifications of east Pacific SST to determine whether east Pacific events modify the Atlantic influence significantly.

Section 4 describes the response of the vertical circulation to the heating modifications, section 5 presents the variability of the response between different cases, and section 6 summarizes conclusions.

2. Model rainfall and condensation heating

The present version of the NCAR CCM-I is a spectral global model, using 11 sigma levels initialized with real data, and running at a horizontal rhomboidal spectral

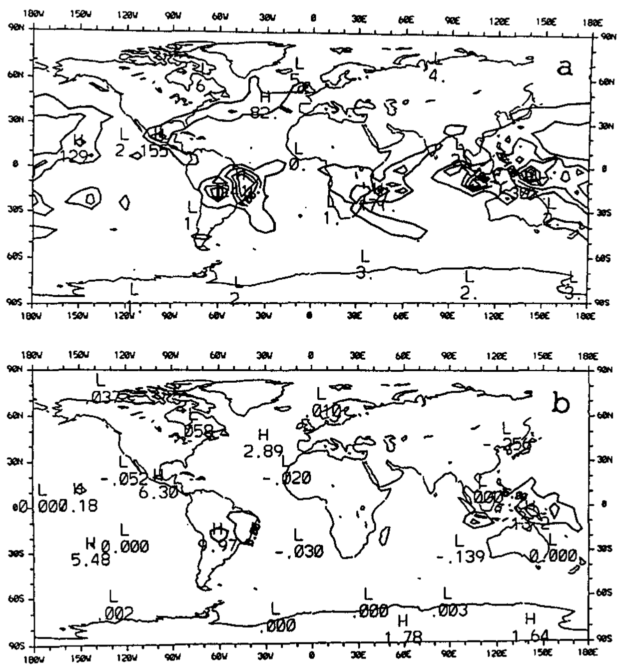

FIG. 2. (a) As in Fig. 1a for an average over days 600–630. Contour interval is 5×10^{-8} m d^{-1}. (b) As in Fig. 1b for an average over days 600–630. Contour interval is 5°C d^{-1}.

truncation retaining 15 waves. Other details of the model are outlined by Buchmann et al. (1990) and further elaborated by Williamson et al. (1987). Williamson and Williamson (1987) describe the model climatology for a perpetual January integration of 1200-d duration. This run does not include anomalous heating.

Figure 1a displays precipitation averaged over a 10-d period from days 620 to 630 of this 1200-d integration. The maximum value of 3.6×10^{-7} m s^{-1} corresponds to a 10-d rainfall of approximately 30 cm. Peaks of this order occur over the western Pacific, the Amazon Basin, and Central America. The maxima in the midtroposphere latent heating are on the order of 10°C d^{-1} in these regions (Fig. 1b). Comparing these heating rates with the anomalies imposed by Zhang (1985), Buchmann et al. (1986), and Paegle et al. (1987) in 10-d integrations indicates that the latter are about 50% larger than heating rates that occur naturally in the model. These studies imposed peak midtropospheric heating rates of 8°C d^{-1} and also produced similar latent heating response in the vicinity of the heat input, leading to total maximum heating rates on the order of 15°C d^{-1}.

Figures 2a and 2b display 30-d averages of rainfall and midtroposphere heating rates, respectively. The maxima in these monthly averages are approximately 70% of the maxima in the 10-d averages. The 30-d integrations of Buchmann et al. (1990) impose total tropical heating modifications that are about twice this magnitude. The next section describes the sensitivity of the model response to the magnitude of the imposed tropical heating.

3. Rainfall response to tropical heating

The experiments of the present study are composed of four 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 in each of the eight separate cases. In the second, third, and fourth ensembles, a heating term is added to the thermodynamic equation during the forecast but does not affect the initialization. In the case of the second and third ensembles, this heating maximizes at 6.6°N, 30°W, and in the case of the fourth ensemble, the heating is strongest at 6.6°S,

TABLE 1.

Experiment	1	2	3	4	5
Heating location, strength	none (control)	30°W, 6.6°N 5°C/d^{-1}	30°W, 6.6°N 1.7°C/d^{-1}	30°W, 6.6°S 1.7°C/d^{-1}	none
SST anomaly	none (control)	E. Pacific −10°C	E. Pacific −3.3°C	E. Pacific −3.3°C	.E. Pacific −3.3°C

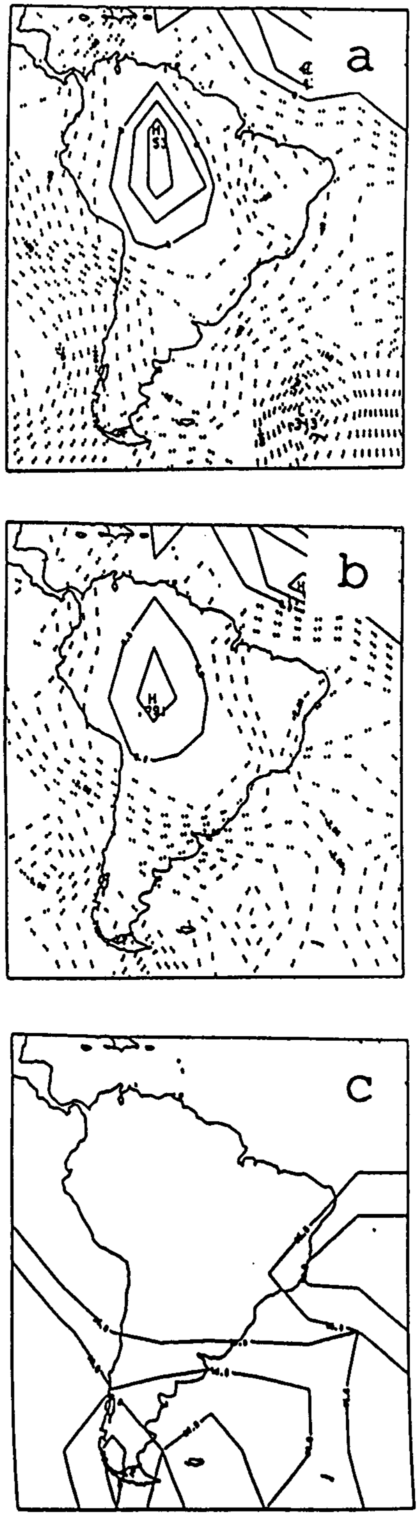

FIG. 3. (a) Ensemble and time-averaged precipitation response for ensemble 2. Contour interval is 0.2×10^{-8} m s^{-1}. (b) The t test statistic for response depicted in Fig. 3a. Contour interval is 1/2. (c) Probability that the response depicted in Fig. 3a is statistically significant. Contour interval is 3%, and only values of 95% and higher are analyzed.

30°W. The heating decreases radially outward with a Gaussian profile that has a half-width of approximately 8° lat and 16° long. It is strongest at 400 mb in both cases, with a local maximum of $8°C\ d^{-1}$ in ensemble 2 and $2.6°C\ d^{-1}$ in ensembles 3 and 4. The column-averaged heating rate is $5°C\ d^{-1}$ in ensemble 2 and $1.7°C\ d^{-1}$ in ensembles 3 and 4. The vertical distribution of the heating is shown in Fig. 12 of Paegle et al. (1987).

The experiment ensembles (2, 3, and 4) contain additional modifications in the form of lowered SST, as also done by Buchmann et al. (1989). In the case of ensemble 2, the SST decrease is rather extreme, peaking at −10°C (used by Buchmann et al. 1989), and in ensembles 3 and 4 it peaks at a more realistic value of −3.3°C. The SST modification is centered near the equator at 135°W and has Gaussian dependence on radial distance, decreasing to half its peak value at 3000 km. Another experiment, ensemble 5, designed to study the influence of the east Pacific heating, acting alone was also run for each of the eight cases. These results will not be emphasized here but are useful to substantiate claims of the relative influence of Atlantic and Pacific modifications. Experiments are summarized in Table 1.

Figure 3a displays the time-averaged precipitation response of the second ensemble in the vicinity of South America. This can be compared with Fig. 14a of Buchmann et al. (1990), which shows a similar case that lacks Pacific cooling, and Fig. 4 of Buchmann et al. (1989), which is similar to the present case but lacks Atlantic heating. The increased Amazon Basin rainfall of Fig. 3a is more in agreement with Buchmann et al. (1989) than with Buchmann et al. (1990), suggesting that the Pacific influence dominates the rainfall of the Amazon Basin. However, in contrast to the situation with only Pacific cooling, the *t* test statistic (Fig. 3b) of the present case is not significant at the 95% level anywhere in the vicinity of the Amazon Basin (Fig. 3c). Reduction of the Atlantic heating and the Pacific cooling to one-third reduces the response substantially (see Fig. 4), and the maximum response shifts southeastward. In summary, the presence of Pacific cooling is sufficient to strongly modify the South American response to tropical Atlantic heating, and reduction of both forcings shrinks the region of statistically significant response.

Figure 5 displays the significance analysis over North America for ensembles 2 and 3. Both strong (Fig. 5) and weak (Fig. 6) forcings produce statistically significant rainfall reductions over North America. The patterns of these rainfall reductions are similar to that in Fig. 13a of Buchmann et al. (1990) and are not shown here. The drying effects are about twice as pronounced in the strongly forced case than in the weakly forced case, but each produces responses that are significant at more than a 95% confidence level over eastern sections of the United States. The responses for ensemble

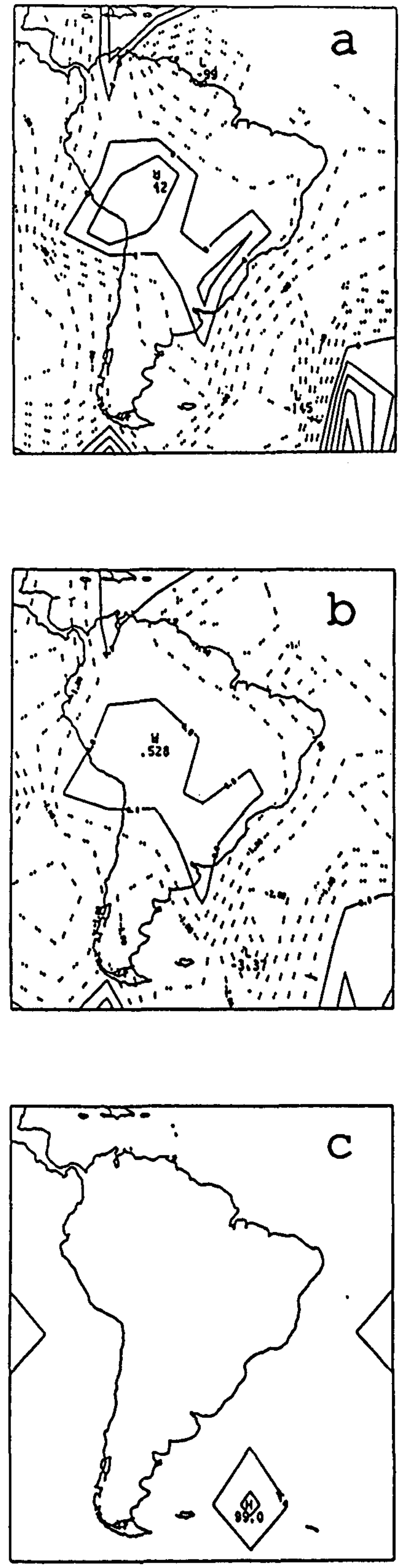

FIG. 4. As in Fig. 3 for weaker tropical forcing, ensemble 3.

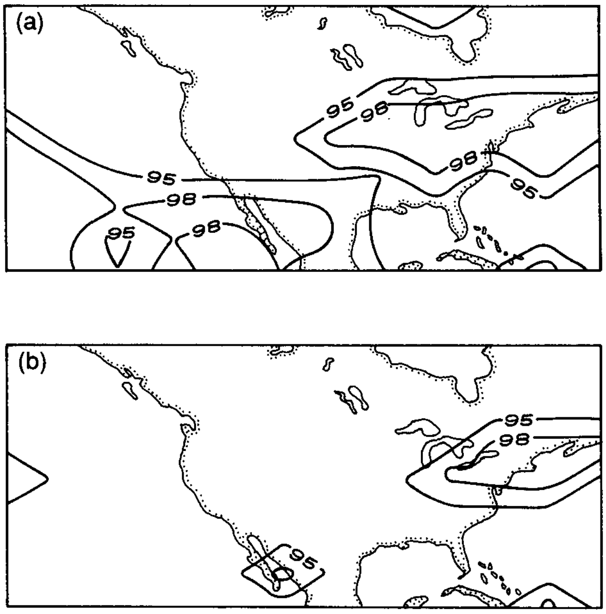

FIG. 5. (a) As in Fig. 3c for North America. (b) As in Fig. 4c for North America.

4 are similar to those for ensemble 3 and are not shown. Comparison of these cases with the experiment in which only the east Pacific tropical SST was changed (not shown) suggests that the tropical Atlantic heating is the principal influence on the eastern North American rainfall response of Fig. 5. The dependence of this conclusion on resolution remains to be studied.

4. Circulation response

Precipitation is affected by those flow modifications that change the ascent rate or modify the air trajectories with respect to low-level moisture sources. The purpose of this section is to describe the vertical motion response.

Figures 6 and 7 display the vertical motion response, its t test statistic, and significance analyses over South America for ensembles 2 and 3, respectively. Both ensembles display increased rising motion (negative values) over central and southern sections of South America and increased subsidence (positive values) outside these regions. These patterns agree with the rainfall response as depicted in Figs. 3 and 4.

Figures 8 and 9 display the same fields over North America for ensembles 2 and 3, respectively. Regions of increased subsidence accompany the areas of decreased rainfall displayed in Figs. 5 and 6. Similar results characterize ensemble 4 (not shown).

5. Case to case variability

The response of the atmosphere to a given forcing is variable from one initial state to another because

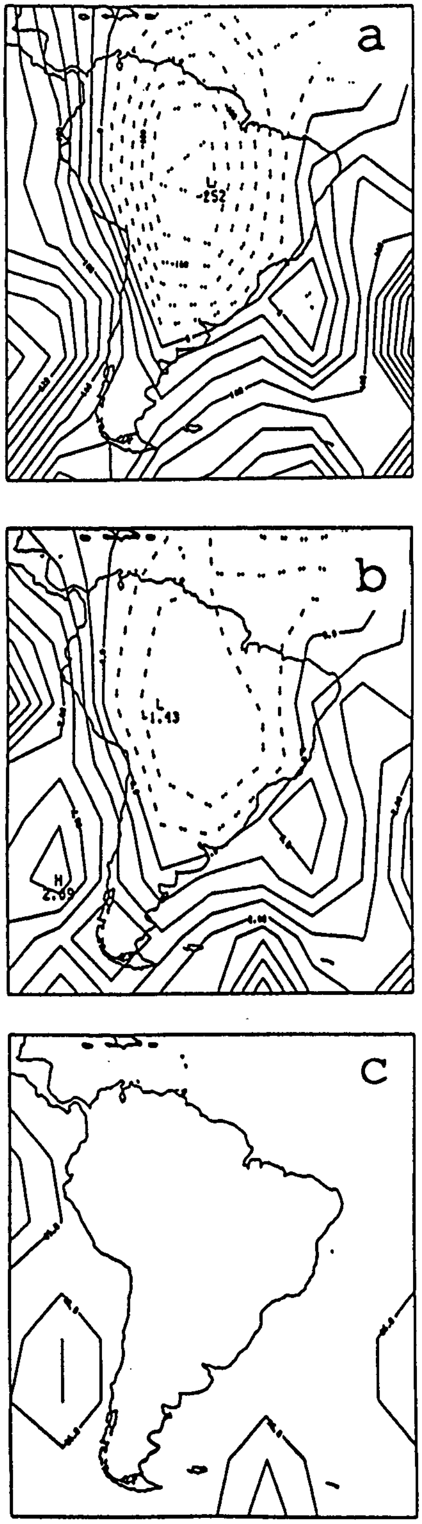

FIG. 6. (a) Ensemble and time-averaged vertical motion response for ensemble 2. Contour interval is 0.4×10^{-2} Pa s^{-1}. (b) The t test statistic for response depicted in Fig. 6a. Contour interval is 1/2. (c) Probability that the response depicted in Fig. 6a is statistically significant. Contour interval is 3%, and only values of 95% and higher are analyzed.

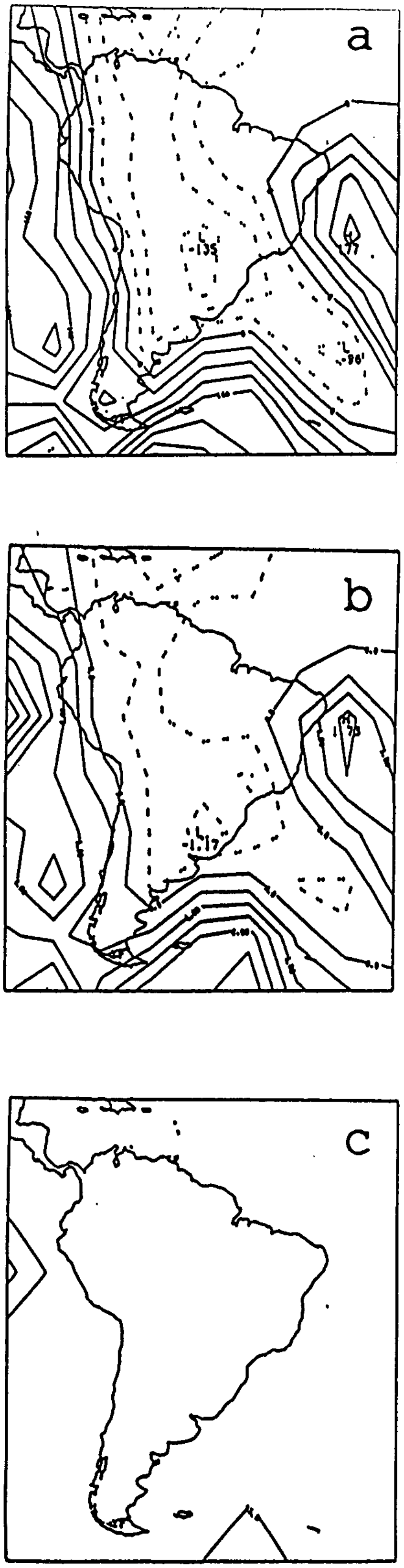

FIG. 7. As in Fig. 6 for ensemble 3.

different initial states represent backgrounds with different refractive indices for propagation of the wave response [see Fig. 7 and accompanying discussion in Trenberth and Branstator (1992)]. Furthermore, the prediction of atmospheric states on the medium to extended range is influenced by the limiting barrier to deterministic predictability that exists within this time interval. A fundamental question regards the frequency with which a foreknown tropical heating anomaly produces a fixed remote response of the same sign. The t test provides one measure of the predictability of a particular response. A more easily communicated measure is the number of times a prespecified anomaly produces a response of the same sign. This can be easily translated into a forecast probability of above- or below-normal conditions.

The studies by Buchmann et al. (1986) and Buchmann et al. (1990) have attempted to attach such measures of variability to the average forecast signal. A total of 16 different forecasts investigating rainfall response over the Americas to heating anomalies of the tropical Atlantic were reported in the second of these studies. Twelve of the 16 forecasts reduced rainfall rel-

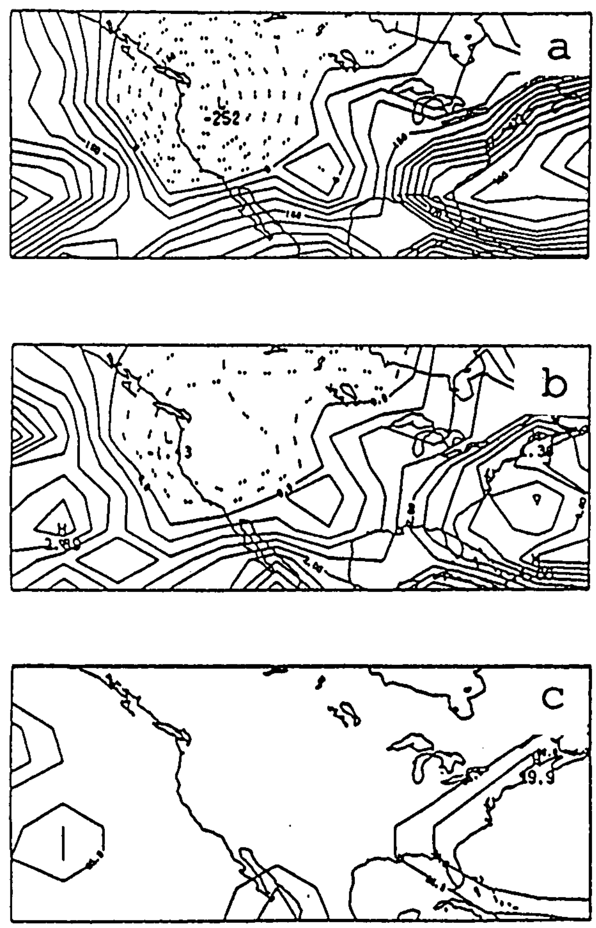

FIG. 8. As in Fig. 6 for North America.

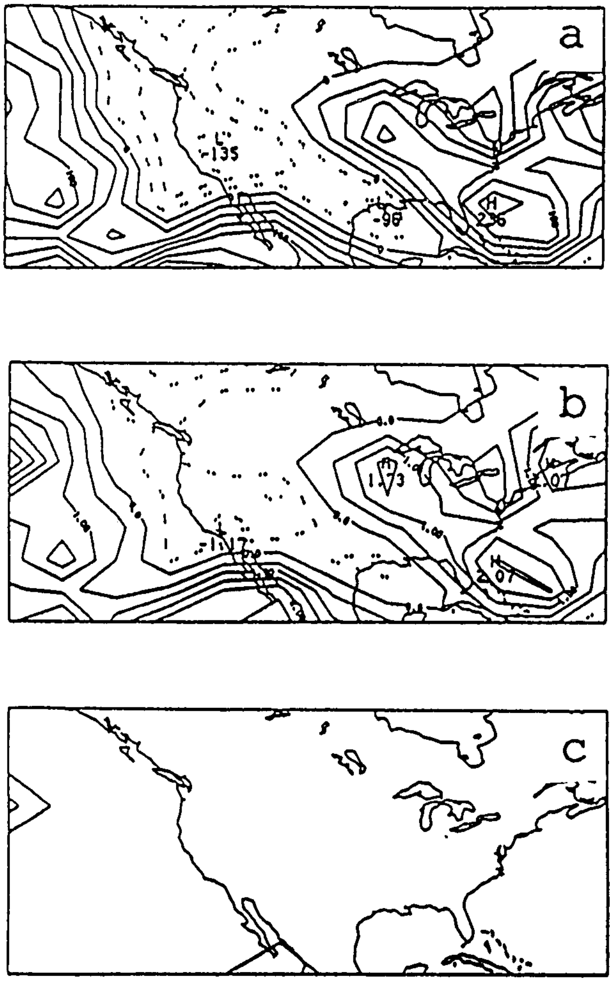

FIG. 9. As in Fig. 8 for North America and the weaker tropical forcing of ensemble 3.

ative to the unperturbed control over eastern sections of North America. Two of the forecasts produced little response there, and only two others predicted likelihood that this anomaly would contribute to relatively wet conditions.

We currently describe the sensitivity of this conclusion to the variation of the tropical heating. In particular, two of the current ensembles (three and four) incorporate more realistic, reduced heating anomalies in addition to SST perturbations over the east Pacific. The time-averaged response of the North American precipitation in each of these cases is presented in Figs. 10 and 11. Subjective evaluation of these diagrams shows that 12 of the 16 cases experience distinct drying on the east coast of the United States, another two have some local drier regions in this vicinity, and two others have variable response, including some rainfall increases. All 8 of the strongly forced cases of ensemble 2 display drying on the east coast of North America (results not shown).

We conclude that the tendency for relatively dry forecasts by this model over eastern North America in conditions of warm tropical Atlantic anomalies is relatively independent of both the magnitude of the forcing and the presence of eastern Pacific cooling. The latter effect appears to promote aridity but is not the primary reason for suppressed precipitation over North America.

6. Discussion and summary

The present study of tropical Atlantic influences upon North America was motivated by our last investigation (Buchmann et al. 1990), which demonstrated that this teleconnection is one of the strongest tropical–extratropical teleconnections produced in a rather extensive series of real-data, medium to extended range forecasts we have performed with the NCAR CCM. Other investigators have studied teleconnections over the Atlantic sector (e.g., Namias 1972; and Moura and Shukla 1981), but most of these studies emphasized rainfall response over South America. Interest in real-data deterministic forecasting over South America prompted our earlier studies of the influence of Atlantic anomalies upon rainfall in the Western Hemisphere. Unexpectedly, these showed that the most significant response to tropical Atlantic heating anomalies resides in eastern North America.

The major simplification in the earlier studies was imposition of rather strong heating anomalies relative to those that may occur in nature, or to those that characterize the climate of the NCAR CCM. The results of section 2 suggest that these prior heating rates were about 50% larger than the naturally occurring rates within the Tropics over similar time intervals in extended range integrations. To cover a broader range of anomalies, we have repeated the integrations of Buchmann et al. (1990) using their unrealistically strong heating, as well as values that are approximately one-third of those large rates. The latter produce weaker net heating than peak values that characterize the unmodified model Tropics, and the new experiments together with the prior studies should bracket the range of likely response to positive tropical Atlantic heating anomalies. To further broaden the range of experimental conditions we have also lowered the SST of the eastern tropical Pacific Ocean.

A total of 24 new 30-d experiments were run. One set of eight cases used strong tropical Atlantic forcing centered north of the equator, another set of eight cases specified moderate tropical forcing here, and the final set of eight cases imposed moderate tropical Atlantic forcing centered south of the equator. Twenty of these cases predicted reduced rainfall over eastern sections of the United States. The t test statistic suggests less than a 5% chance that the drying is a sampling fluctuation unrelated to the tropical heating modifications. A separate set of eight cases in which only the east

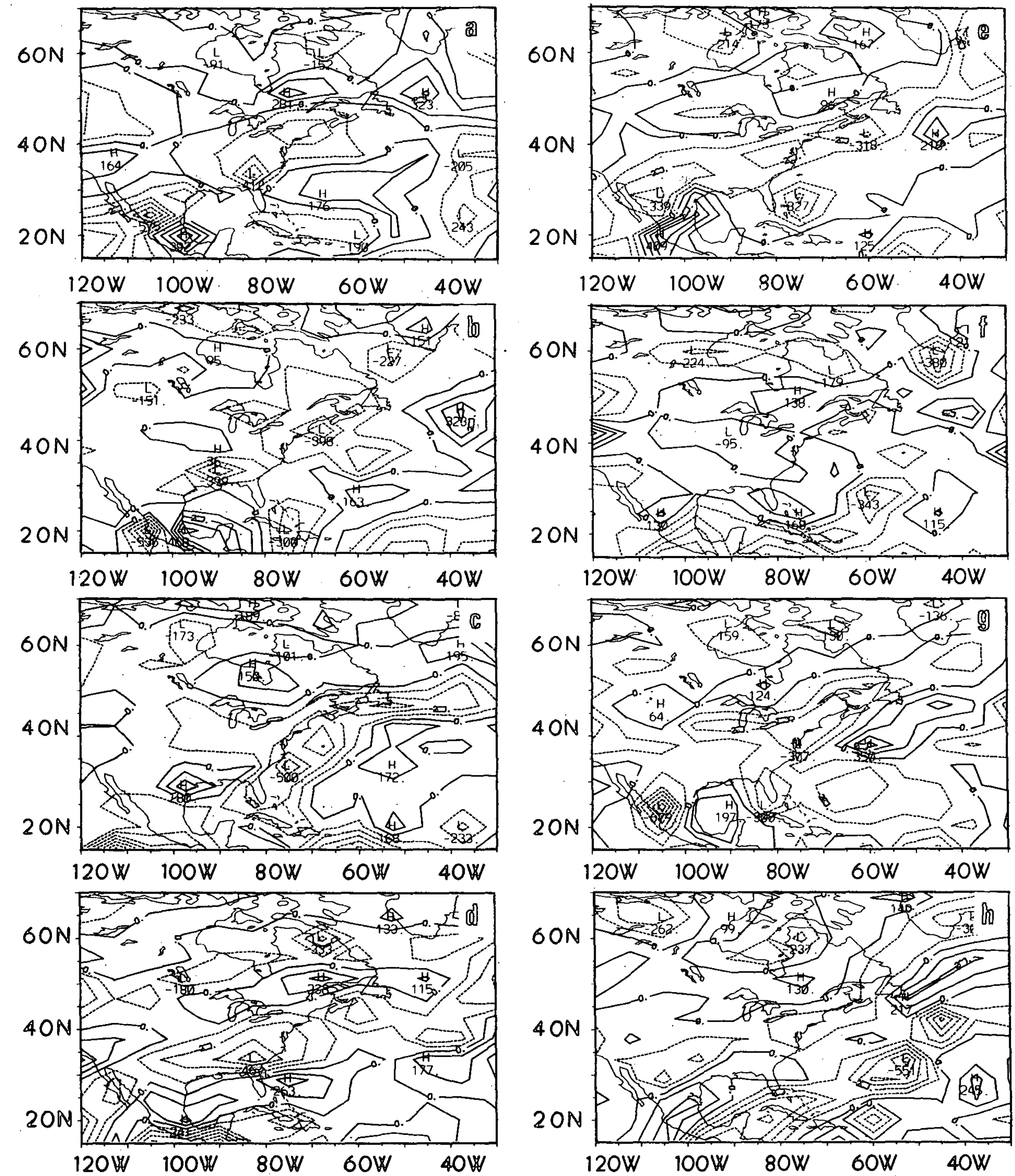

FIG. 10. Time-averaged precipitation response for ensemble 3 (experiment − control) for (a) 1977, (b) 1978, (c) 1979, (d) 1980, (e) 1981, (f) 1982, (g) 1983, and (h) 1984. Contour interval is 10^{-8} m s^{-1}.

Pacific SST cooling was imposed indicates that this supports the drying influence over eastern North America, but is not the dominant effect.

Although some of the deficiencies of the prior simulations have been addressed in the present study, others persist. The crude model resolution of rhomboidal 15 may be the most serious of the remaining simplifications. Such resolution carries information about rainfall only on a resolution of approximately 5° lat and long. Finer resolution would be desirable

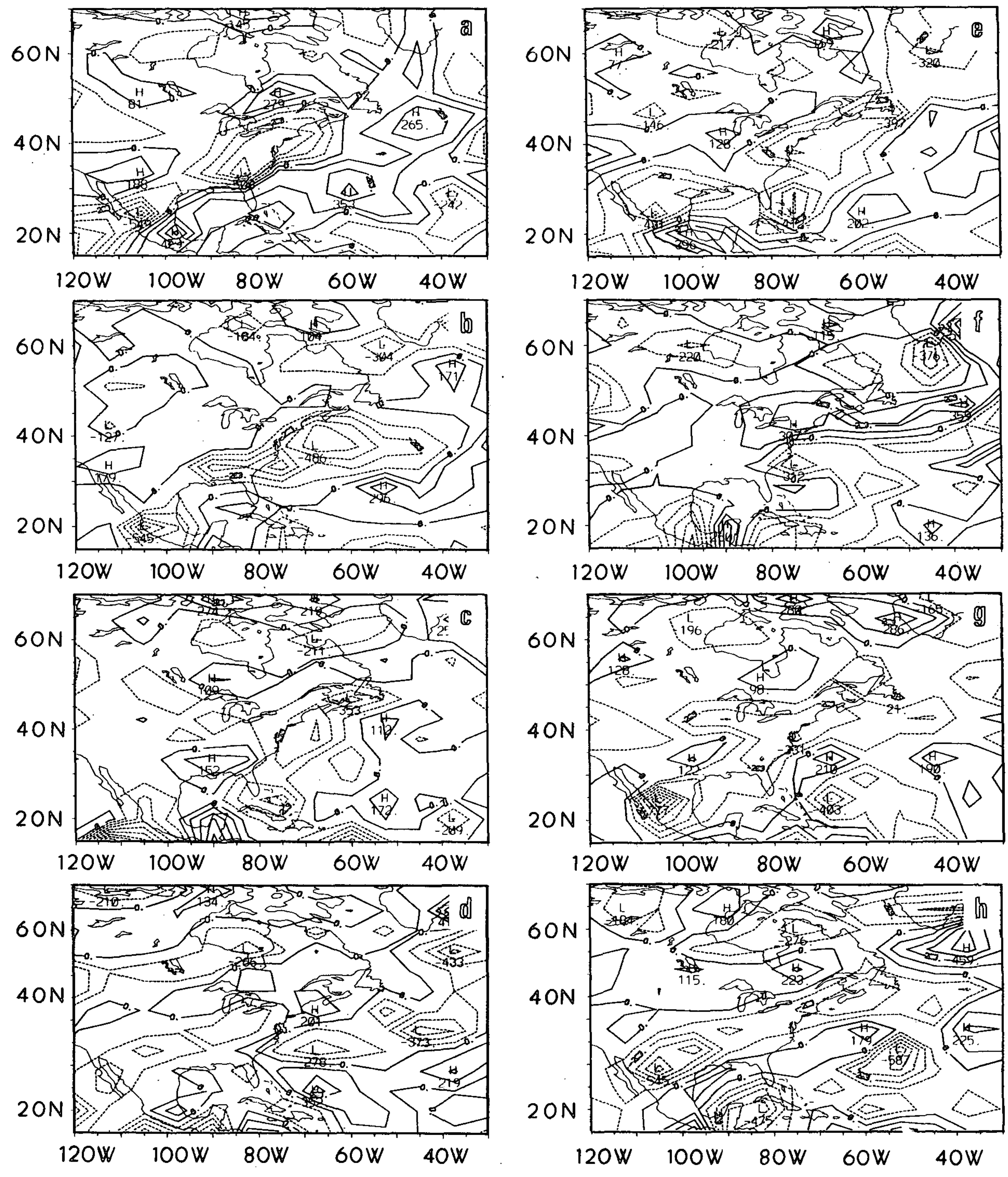

FIG. 11. As in Fig. 10 but for ensemble 4.

for rainfall prediction. Better resolution of surface transfer processes, possibly including the diurnal cycle would also be useful for present integrations that span the typical residence time of atmospheric water vapor.

The present results appear to be more relevant to the North American drought of 1986 than to the North American drought of 1988, because the former commenced during the northern winter, while the latter commenced in the northern spring (Trenberth and Branstator

1992). It would be useful to repeat the present set of experiments for conditions characteristic of spring 1988, when both the tropical Atlantic heating and North American precipitation anomalies became well established. The connection between the Atlantic anomaly and the 1988 drought requires further analysis in view of Trenberth and Branstator's (1992) findings and the current knowledge of energy propagation in an atmosphere with zonally and meridionally varying basic state.

In view of these simplifications, the present study and recent related work by Buja (1989), which investigates the linearity and rate of response, are of a preliminary nature; they are intended to provide relatively inexpensive guidelines regarding the design of experiments intended to clarify the effect of tropical heating anomalies upon the extratropics with a more complete and focused set of higher resolution experiments. The first group of such experiments is presented in a related study (Buchmann et al. 1994, unpublished manuscript).

Acknowledgments. This research was partly supported by NSF Grants INT 8602690, ATM 8905369, and ATM 9014650 to the University of Utah. J. Buchmann's participation was partly supported by CNPQ, Brazil.

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