

## The Influence of Soil Wetness on Near-Surface Atmospheric Variability

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### ABSTRACT

The influence of land surface processes on near-surface atmospheric variability on seasonal and interannual time scales is studied using output from two integrations of a general circulation model. In the first experiment, of 50 years duration, soil moisture is predicted, thereby taking into consideration interactions between the surface moisture budget and the atmosphere. In the second experiment, of 25 years duration, the seasonal cycle of soil moisture is prescribed at each grid point based upon the results of the first integration, thereby suppressing these interactions. The same seasonal cycle of soil moisture is prescribed for each year of the second integration. Differences in atmospheric variability between the two integrations are due to interactions between the surface moisture budget and the atmosphere.

Analyses of monthly data indicate that the surface moisture budget interacts with the atmosphere in such a way as to lengthen the time scales of fluctuations of near-surface relative humidity and temperature, as well as to increase the total variability of the atmosphere. During summer months at middle latitudes, the persistence of near-surface relative humidity, as measured by correlations of monthly mean relative humidity between successive months, increases from near zero in the experiment with prescribed soil moisture to as large as 0.6 in the experiment with interactive soil moisture, which corresponds to an  $e$ -folding time of approximately two months. The standard deviation of monthly mean relative humidity during summer is substantially larger in the experiment with interactive soil moisture than in the experiment with prescribed soil moisture. Surface air temperature exhibits similar changes, but of smaller magnitude.

Soil wetness influences the atmosphere by altering the partitioning of the outgoing energy flux at the surface into latent and sensible heat components. Fluctuations of soil moisture result in large variations in these fluxes, and thus significant variations in near surface relative humidity and temperature. Because anomalies of monthly mean soil moisture are characterized by seasonal and interannual time scales, they create persistent anomalous fluxes of latent and sensible heat, thereby increasing the persistence of near-surface atmospheric relative humidity and temperature.

### 1. Introduction

The influence of anomalous soil moisture conditions on the atmosphere has been the subject of research for some time. Namias (1958, 1963) was among the first to address the issue, noting that seasonal anomalies of soil wetness could have an impact on the seasonal cycle of the atmosphere. More recently, a number of modeling studies have explicitly examined the influence of anomalies of soil moisture on the atmosphere.

Two types of studies have been prevalent. The first is exemplified by Shukla and Mintz (1982) who examined the impact on the atmosphere of prescribed, constant anomalies of soil wetness. They demonstrated that negative (positive) anomalies of soil moisture decrease (increase) evaporation rates and increase (decrease) surface temperatures. In the second type of study, exemplified by Walker and Rowntree (1977), Rowntree and Bolton (1983), Rind (1982) and Yeh et al. (1984), soil moisture was computed interactively

in the model integrations after initial anomalies of soil moisture were prescribed. In particular, Walker and Rowntree (1977) and Yeh et al. (1984) demonstrated that, under certain conditions, positive anomalies of soil moisture interact with the atmosphere in such a way as to sustain themselves by enhancing evaporation, thereby increasing precipitation rates and prolonging the initial soil wetness anomaly. Yeh et al. (1984) also showed that the degree of this persistence has a latitudinal gradient, with anomalies of soil moisture persisting longer at higher latitudes.

A different approach was used by Delworth and Manabe (1988, hereafter referred to as DM), and Gordon and Hunt (1987). In these studies, the temporal variability of soil moisture in a multiple-year integration of a general circulation model of the atmosphere (GCM) was analyzed. Both studies explicitly examined the natural variability of soil moisture in a GCM, rather than the response of a GCM to initially prescribed anomalies of soil moisture. They show that, within a GCM, time series of soil moisture contain variance on seasonal to interannual time scales. DM demonstrated that the soil layer acts as an integrator of short time

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scale precipitation anomalies, transforming the almost white noise time series of monthly mean precipitation into the red noise time series of soil moisture. DM also showed that the time scales of soil moisture anomalies are primarily controlled by potential evaporation and the ratio of potential evaporation to the mean precipitation rate.

While the main focus of DM was the explicit study of the natural variability of soil moisture in a GCM, it is the purpose of this paper to identify the effect of soil moisture variability on atmospheric variability. DM briefly described how soil moisture variability can increase the variance of surface air temperature. In this paper, we focus on the influence of soil wetness fluctuations on the persistence of the near-surface atmosphere. To do this, we compare atmospheric variability in two integrations of a GCM. In the first integration, of 50 years duration, soil moisture is predicted. In the second integration, of 25 years duration, the seasonal cycle of soil moisture is prescribed at each grid point based upon the results of the first integration (the prescription of soil moisture is discussed more fully in section 2). In this manner, interactions between the soil layer and the atmosphere are not permitted in the second integration. Differences in atmospheric variability between the two experiments are therefore attributable solely to interactions between the soil layer and the atmosphere in the first experiment.

It should be emphasized that the model variability results presented here must be interpreted in light of the model's ability to simulate the current climate correctly. In regions where the simulation is poor, some of the quantitative aspects of the variability results cannot be taken too literally, but the mechanisms by which soil wetness influences the model atmosphere in various climatic regions will still be relevant to the real climate system.

## 2. Model description

The model and integrations used are the same as in DM and will be only briefly described here. The model consists of two parts: (i) a general circulation model of the atmosphere, and (ii) a heat and water balance model over the continents. The atmospheric GCM is very similar to that described by Manabe and Hahn (1981). The spectral computations employ the "rhomboidal 15" wavenumber truncation. The resultant transform grid has a resolution of  $7.5^\circ$  longitude by  $4.5^\circ$  latitude. There are nine finite-difference levels in the vertical. Zonal mean cloud cover is prescribed to be constant in time, depending only on latitude and height. A seasonal cycle of solar radiation at the top of the atmosphere is prescribed, with no diurnal variations. The seasonal cycles of sea surface temperature and sea ice are prescribed at all ocean grid points based upon observed monthly mean fields.

A heat and water balance is computed over land. Ground surface temperature is computed from the requirement that a balance exist at the surface between net radiation and the vertical fluxes of latent and sensible heat. No heat storage is allowed in the soil layer.

In the first integration, the surface moisture budget is computed by the "bucket method." Changes in soil moisture are computed from the rates of rainfall, evaporation, snowmelt, and runoff, as given by

$$dw(t)/dt = -E_p f(w(t)/w_{FC}) + \text{Rainfall} + \text{Snowmelt} - \text{Runoff} \quad (1)$$

where

$t$  time  
 $w(t)$  soil moisture (cm)  
 $w_{FC}$  field capacity (= 15 cm)  
 $E_p$  potential evaporation (cm d<sup>-1</sup>)

and

$$E_p = -\rho C_d |v_9| (q_9 - q_s(T_*)) \quad (2)$$

where

$\rho$  density of the air (g cm<sup>-3</sup>)  
 $C_d$  drag coefficient  
 $v_9$  wind speed (cm s<sup>-1</sup>) at the lowest model level (about 85 meters above the surface)  
 $q_9$  mixing ratio at the lowest model level  
 $q_s(T_*)$  saturation mixing ratio corresponding to the ground surface temperature  
 $T_*$  ground surface temperature

and

$$f(w(t)/w_{FC}) = \begin{cases} w(t)/(0.75w_{FC}), & \text{if } w \leq 0.75w_{FC} \\ 1, & \text{if } w > 0.75w_{FC}. \end{cases} \quad (3)$$

If the computed soil moisture exceeds the field capacity (15 cm), the excess moisture runs off and is no longer accounted for in the model. Evaporation from the soil [the first term on the right side of (1)] is determined as a product of the potential evaporation rate (2) and a function of soil wetness (3). This function incorporates the observation that evapotranspiration is at the potential rate when soil moisture is above some critical threshold (75% saturation in this parameterization), but it decreases when soil moisture is below that threshold.

In the second integration, the seasonal cycles of soil wetness and surface albedo are prescribed at all land points and are identical for each year. The prescribed soil wetness and surface albedo values were derived from the results of the first integration by the following procedure. At each land point, 5-year means for  $f(w(t)/w_{FC})$  and surface albedo were computed for each 5-day period of the year. The 5-day means, 73 in all, determine a Fourier series, which in turn determines

the daily values of  $f(w(t)/w_{FC})$  and surface albedo that are used in the second integration. The potential evaporation rate is computed in the same manner as in the first integration. This second experiment will be referred to as "SMP" (Soil Moisture Prescribed), while the first experiment will be referred to as "SMI" (Soil Moisture Interactive).

The model was integrated for several years from an isothermal atmosphere at rest to a state of statistical equilibrium. From that point, a 50-year integration was performed with soil moisture computed interactively. From that same starting point, the second integration was performed, of 25 years duration, with a prescribed seasonal cycle of soil wetness and surface albedo.

### 3. Persistence of soil wetness

Before examining the influence of the variability of soil moisture on the variability of the atmosphere, one should be familiar with the variability of soil moisture itself. The data used for analyses (for soil moisture and all other variables) are deviations of monthly mean values from the long-term mean for that month. One measure of the temporal variability of monthly mean soil moisture is the lag-one autocorrelation coefficient. At each grid point, the time series of soil moisture was correlated with itself, but lagged one month. A map of these coefficients, computed using data from the months of June, July, and August (JJA), is plotted in Fig. 1a for SMI. The autocorrelations are generally positive, demonstrating that anomalies of soil moisture persist on monthly time scales. There is a latitudinal gradient, with autocorrelations ranging from less than 0.4 at lower and middle latitudes of the Northern Hemisphere to greater than 0.7 at high latitudes of the Northern Hemisphere and portions of the Southern Hemisphere.

As discussed more fully in DM, the persistence of monthly mean soil moisture may be viewed as the red-noise response of the soil layer to the time series of monthly mean rainfall, which resembles white noise (lag-one autocorrelations near zero). The soil layer acts as an integrator of the time series of rainfall, producing a time series of soil moisture which is similar to red noise (lag-one autocorrelations greater than zero).

It is important to note, however, that the time series of monthly mean rainfall only *resembles* white noise; there is, in fact, some small persistence, and lag-one autocorrelations are as large as 0.3 in SMI for a few small regions. The spectra of daily rainfall (not shown) demonstrate that model precipitation resembles white noise at periods longer than about one week.

The degree of persistence of soil moisture anomalies depends on how rapidly anomalies of moisture are removed from the soil layer by evaporation. As shown in (1), potential evaporation is used in conjunction with soil wetness to determine the model evaporation rate. Consequently, the smaller (larger) the value of

potential evaporation, the smaller (larger) the evaporation rate, the more slowly (rapidly) anomalies of soil moisture are dissipated, and the larger (smaller) the autocorrelations of soil moisture. This can be seen by comparing Fig. 1a with Fig. 1b, a map of potential evaporation for JJA. Smaller potential evaporation values at higher latitudes where insolation is weak result in low evaporation rates and large autocorrelations of soil moisture.

One other factor strongly influencing the persistence of soil moisture is the ratio of the potential evaporation rate to the precipitation rate. Where this ratio is less than one, evaporation alone cannot balance precipitation and the soil is frequently saturated. From (1) and (3), evaporation is at the potential rate, and changes of soil moisture are chiefly governed by short time scale precipitation anomalies resulting in a low persistence of soil moisture. This explains the small autocorrelation values found in the extreme northern part of South America, the extreme northeastern region of Siberia, and the area to the east of the Tibetan plateau.

Using the analogy that the time series of soil moisture is similar to red noise (the spectral results presented later will show this to be a good assumption), the autocorrelation values in Fig. 1a can be translated into  $e$ -folding times. For a red-noise process (Jones 1975):

$$r(t) = \exp(-\lambda t) \quad (4)$$

where  $r(t)$  is the autocorrelation at lag  $t$  (the lag is one month for Fig. 1a) and  $(1/\lambda)$  is the  $e$ -folding time of anomalies in the absence of forcing. Using this relation, one-month lagged autocorrelation values of 0.8, 0.6, .04 and .02 correspond to  $e$ -folding times of 4.5, 2.0, 1.1 and .06 months respectively.

Soil moisture autocorrelations for December–January–February (DJF) are shown in Fig. 2a. Contrasting this map to Fig. 1a shows that there are seasonal variations of the persistence of soil moisture. In general, persistence is larger in winter than in summer, a result of smaller potential evaporation values in winter when insolation is weak. Fig. 2b shows potential evaporation values for DJF, which can be contrasted with Fig. 1b. Over the middle and high latitudes of the Northern Hemisphere, potential evaporation is very small in DJF, resulting in soil moisture autocorrelations larger than 0.9. For the Southern Hemisphere, potential evaporation values are smaller in JJA (Southern Hemisphere winter) than in DJF, resulting in larger soil moisture autocorrelations for JJA than DJF.

The dependence of soil moisture variability on potential evaporation implies that the quantitative aspects of the results presented here depend on the definition of potential evaporation used. It should be noted that, in the present model, potential evaporation is not independent of soil moisture. As a soil layer dries, the ground surface temperature tends to increase, leading to larger potential evaporation rates. If potential evap-

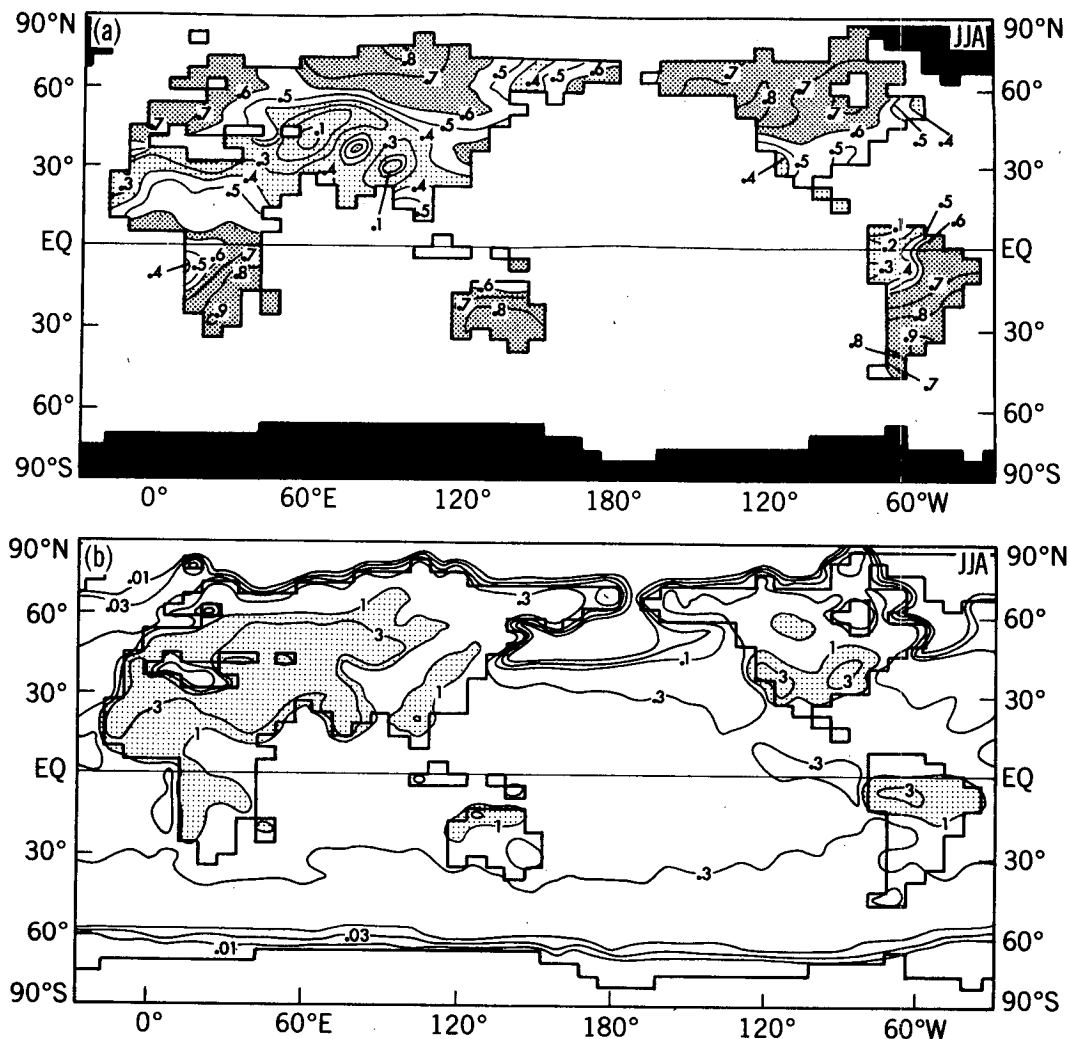


FIG. 1. (a) Lag one autocorrelation values of soil moisture for the months of June, July and August (JJA) for SMI. At each grid point, deviations of monthly mean soil moisture from the long-term mean for that month were correlated with data from the same grid point, but lagged one month. Coefficients greater than 0.16 (0.3) are significantly different from zero at the 95% (99.9%) confidence level (see Chatfield 1984, p. 63). Values greater than 0.6 are densely stippled, while values less than 0.4 are lightly stippled. Permanently ice-covered regions are black. (b) Potential evaporation ( $\text{cm d}^{-1}$ ) for JJA in SMI.

oration had been defined as the evaporation from a completely wetted surface (Budyko 1974), the quantitative aspects of the results would have differed somewhat, but the basic impact of potential evaporation on the persistence of soil moisture would be the same.

#### 4. Persistence of relative humidity

##### a. Geographical and seasonal variations

Soil moisture influences the near-surface atmospheric moisture content and temperature, and therefore near-surface relative humidity, by affecting the surface fluxes of latent and sensible heat. Differences in the variability of near-surface relative humidity between the two experiments will be examined in order to assess the influence of fluctuations of soil moisture

on the atmosphere. We concentrate on relative humidity because this field is strongly influenced by soil wetness. The term "near-surface" refers to the lowest finite-difference level of the model, which is approximately 85 meters above the surface of the earth.

For convenience, an index of relative humidity is defined as the monthly mean atmospheric mixing ratio divided by the saturation mixing ratio corresponding to the monthly mean temperature. This is not identical to the monthly mean relative humidity due to the non-linearity of the Clausius–Clapeyron equation, but it is nevertheless an adequate indicator of near-surface atmospheric relative humidity. Hereafter, the term "relative humidity" refers to this index. Time series of this index were computed for both experiments. Lag one autocorrelations of the deviations of monthly mean

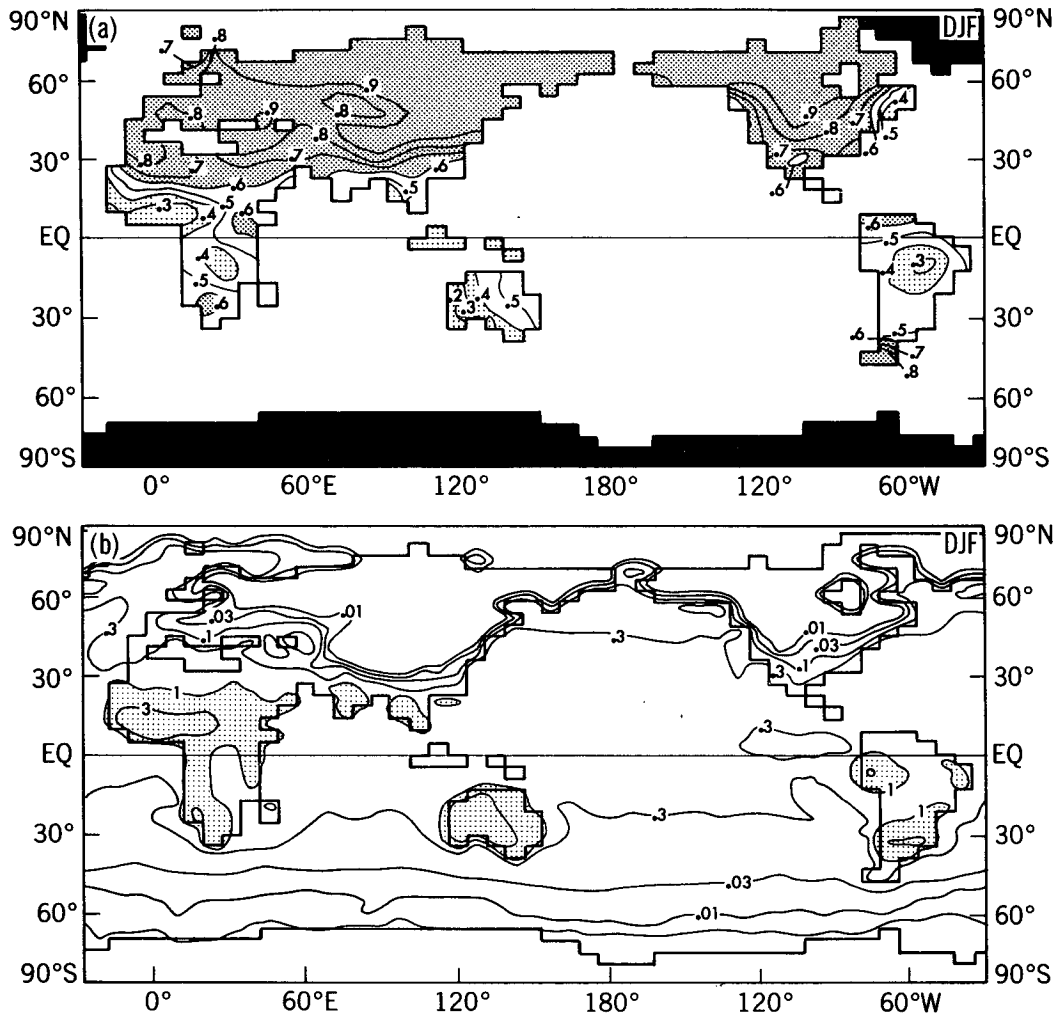


FIG. 2. (a) Same as Fig. 1a, for DJF. (b) Same as Fig. 1b, for DJF.

relative humidity from the long-term mean for that month were then computed using JJA data and are shown in Figs. 3a and 3b. Differences between these two maps indicate the effect of interactions between soil wetness and the atmosphere on the persistence of anomalies of relative humidity. As in Fig. 1a, coefficients in Fig. 3a greater than 0.16 (0.3) are significantly different from zero at the 95% (99%) confidence level. Due to the smaller number of points in the time series for SMP, coefficients in Fig. 3b greater than 0.22 are significantly different from zero at the 95% confidence level.

The differences between the two maps are striking. Autocorrelations in SMI are greater than 0.4 in many locations over land, suggesting that anomalies of relative humidity persist on the monthly time scale when interactions between the soil layer and the atmosphere occur. The largest values occur over continental regions, while there is virtually no persistence over the

oceans (note that sea surface temperatures are prescribed). Over land, small persistence is seen at very high latitudes of the Northern Hemisphere and in a wide band from northern Africa to central Asia. In contrast, anomalies of relative humidity have virtually no persistence in SMP.

There are also seasonal variations of the persistence of relative humidity. The autocorrelations for DJF are plotted in Fig. 4 for SMI. Over the middle and high latitudes of the Northern Hemisphere, persistence of relative humidity is near zero, in sharp contrast to JJA (Fig. 3a). Persistence is also smaller in DJF than in JJA for the Southern Hemisphere.

Another way to view the seasonal and latitudinal variations of the persistence of model soil moisture and relative humidity is through latitude-month plots of autocorrelation coefficients, as shown in Figs. 5a (soil moisture) and 5b (relative humidity). A coefficient plotted for January denotes the zonal mean over land

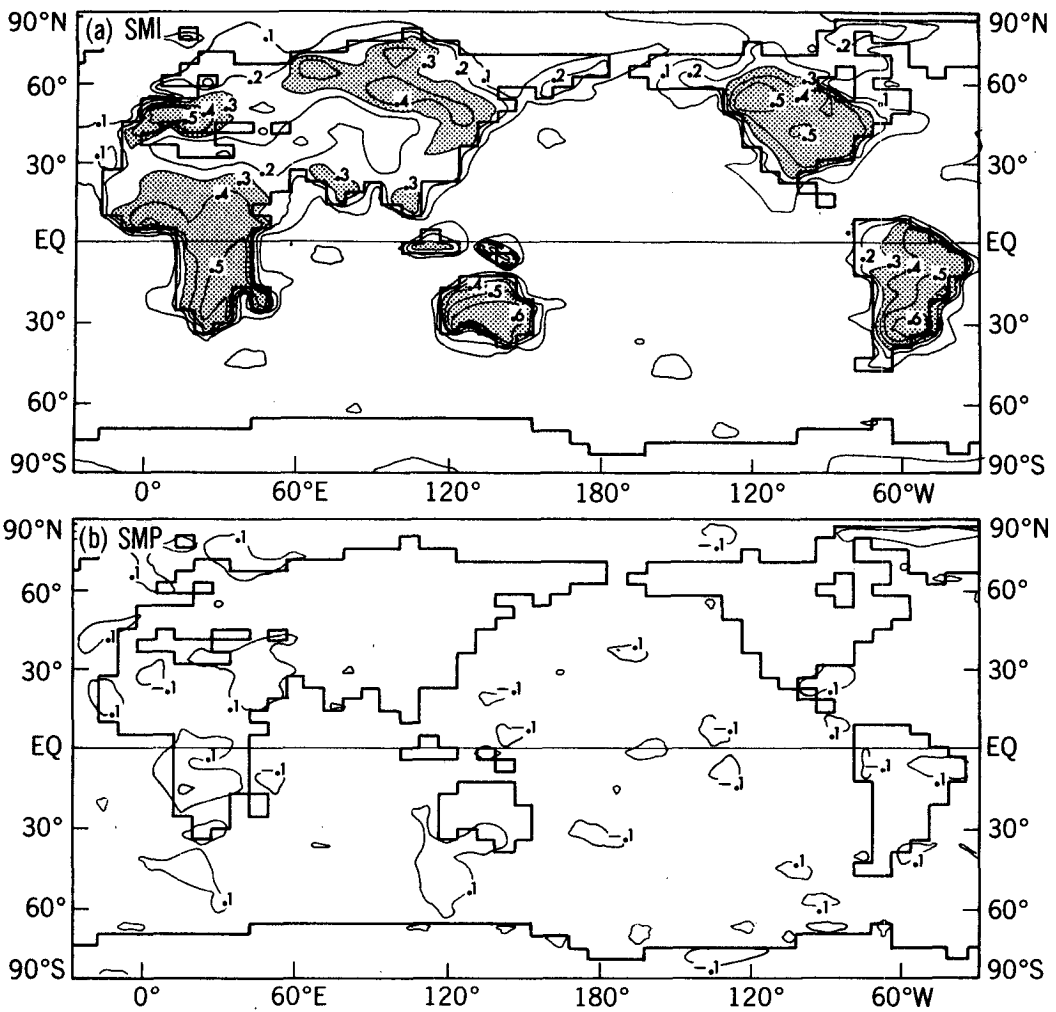


FIG. 3. Lag one autocorrelation values for the time series of the index of near-surface relative humidity using JJA data. (a) Results from SMI, values greater than 0.3 are stippled. Coefficients greater than 0.16 and 0.3 are significantly different from zero at the 95% and 99.9% confidence levels, respectively. (b) Results from SMP. Coefficients greater than 0.22 are significantly different from zero at the 95% level.

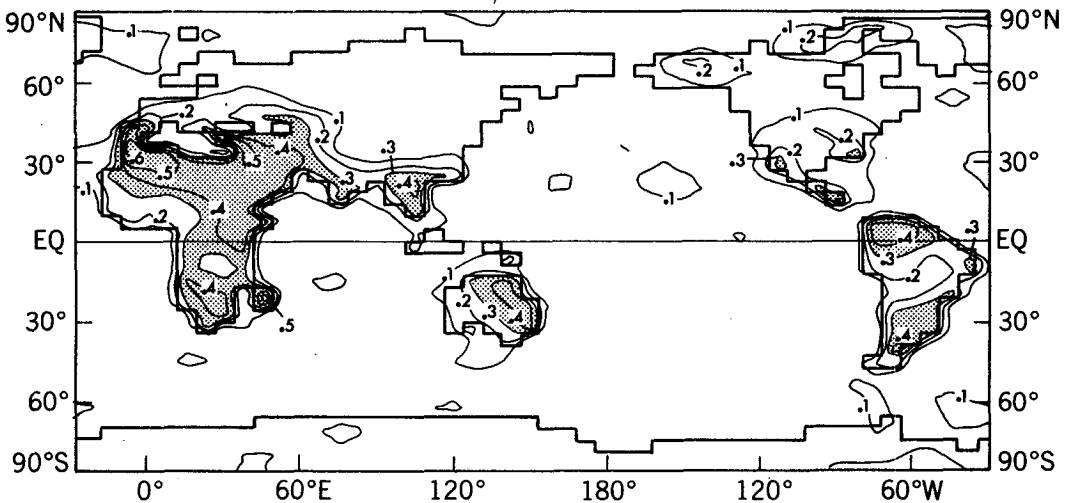


FIG. 4. Lag one autocorrelation values for the time series of the index of near-surface relative humidity using DJF data. Values greater than 0.3 are stippled. Statistical significance is the same as in Fig. 3a.

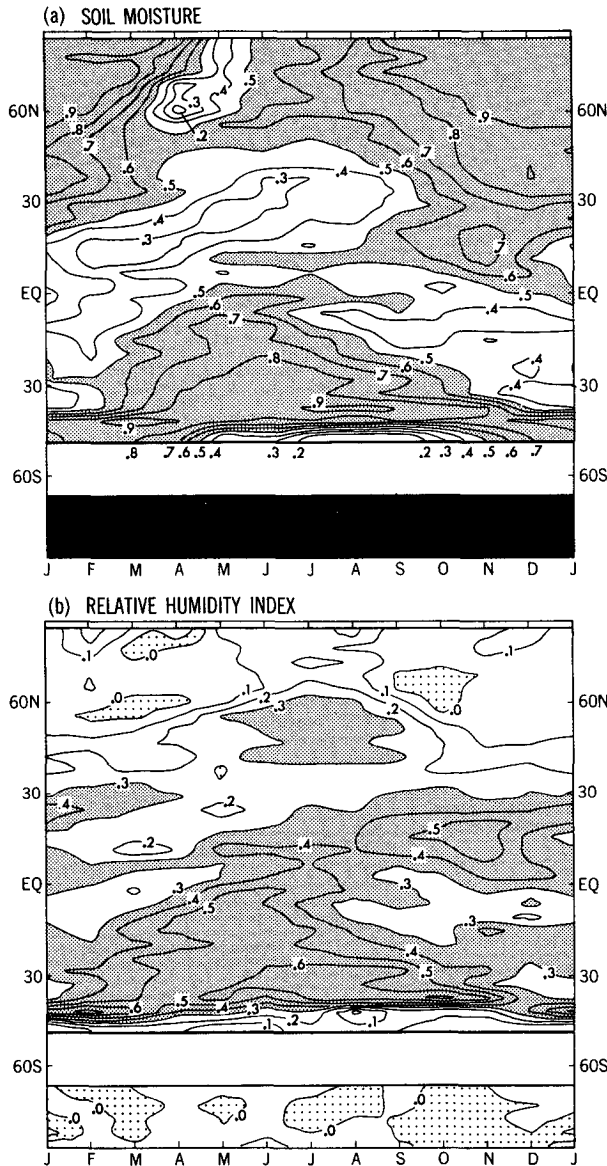


FIG. 5. Latitude/time plots of lag one autocorrelation values for experiment SMI. Autocorrelation values were computed at each grid point for each pair of months in the year (a value for January denotes the correlation between anomalies in January and anomalies in February for all years of the experiment). These values were then zonally averaged over land. No data are plotted between 50° and 67°S where there are very few or no land points in the model. (a) Soil moisture. Values greater than 0.5 are stippled. Latitudes for which all land points are permanently ice-covered are black. (b) Near-surface relative humidity. Values greater than 0.3 are densely stippled. Values less than 0.0 are lightly stippled.

of the correlation coefficient between anomalies in January and anomalies in February. In large part, Fig. 5a represents the effect of the seasonal and latitudinal variations of potential evaporation on the persistence of soil moisture, as discussed previously. One exception, however, is the minimum in autocorrelation near 60°N during April. This feature is caused by frequent satu-

ration of the soil layer from snowmelt, as discussed in DM. The results for relative humidity in SMI are shown in Fig. 5b. For middle latitudes of the Northern Hemisphere, there is a maximum of persistence in the summer months. At higher latitudes, persistence is generally low during the entire year. The pattern is somewhat more complicated at lower latitudes.

Differences in the temporal variability of relative humidity between the two experiments can also be seen by computing the spectrum of the time series of relative humidity anomalies at each grid point. These spectra were then areally averaged over a large region of central North America to arrive at a composite spectrum of relative humidity for each experiment. These spectra, along with the areal mean spectrum of soil moisture from SMI, are shown in Fig. 6. (Note that for SMP the spectrum of soil moisture anomalies would be zero at all frequencies, because the seasonal cycle of soil moisture is prescribed and there are no anomalies.) The areas used to construct the composites are defined in the figure caption. Two observations are clear: 1) the total variance of relative humidity (as measured by the area under the spectrum) is substantially larger in SMI than in SMP; and 2) most of the increase of variance is located at low frequencies, suggesting that fluctuations of relative humidity in SMI are characterized by much longer time scales than in SMP. By comparing these three spectra, it appears that the effect of interactive soil moisture is to both “redden” the spectrum of relative humidity (i.e., preferentially enhance the low frequency variance) and increase the total variance of relative humidity. The reasons for these effects are discussed below.

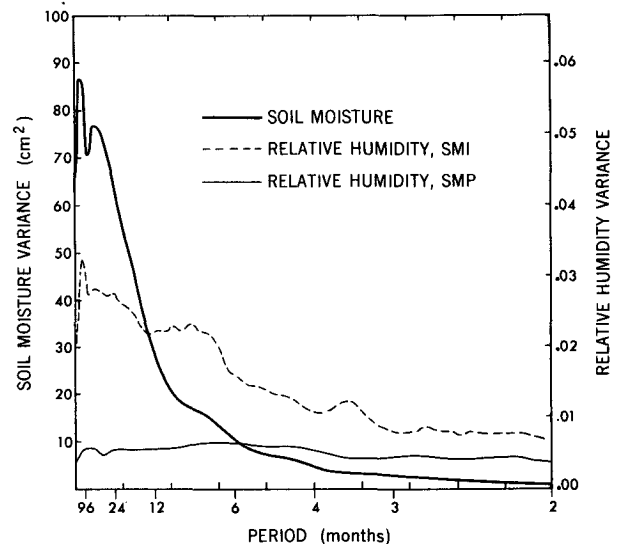


FIG. 6. Spectra of soil moisture and relative humidity areally averaged over the region of North America between 36° and 54°N, 79° and 116°W. Solid, heavy line is soil moisture for SMI. Solid, thin line is relative humidity for SMP. Dashed line is relative humidity for SMI.

*b. Mechanisms*

Variations in soil wetness influence the atmosphere by altering the model surface energy balance, which requires that the net radiative flux at the surface be balanced by the sum of the fluxes of latent and sensible heat (there is no storage of heat allowed in the model soil layer). The latent heat flux in the model depends strongly on soil wetness via the relation

$$LH = LE_p f(w(t)/w_{FC}) \quad (5)$$

where LH is the latent heat flux at the surface, L is the latent heat of evaporation,  $E_p$  is the potential evaporation rate given by (2), and  $f(w(t)/w_{FC})$  is given by (3). Anomalies of soil wetness create anomalies in the latent heat flux. For a given net radiative heat flux at the surface, the surface energy balance dictates that anomalies of latent heat are accompanied by anomalies of the sensible heat flux of opposite sign.

These anomalous fluxes of latent and sensible heat,

created by persistent soil moisture anomalies, are themselves quite persistent. Figures 7a and 7b show the lag one autocorrelations for experiment SMI of anomalies of the monthly mean latent and sensible heat fluxes during JJA (statistical significance limits are the same as for Fig. 3a). For both fluxes there is considerable persistence in SMI, while there is virtually no persistence in SMP (not shown). The geographical dependence of the persistence of these fluxes is similar to the geographical dependence of the persistence of relative humidity (Fig. 3a).

Large negative correlations between the latent and sensible heat fluxes are observed over land (see Fig. 8) due to the requirement that the sum of these fluxes balance the net radiation. Although other factors can influence the sensible heat flux, the importance of changes in the latent heat flux on the sensible heat flux is demonstrated by the large negative correlations. Because relative humidity depends oppositely on temperature and moisture content, these negatively cor-

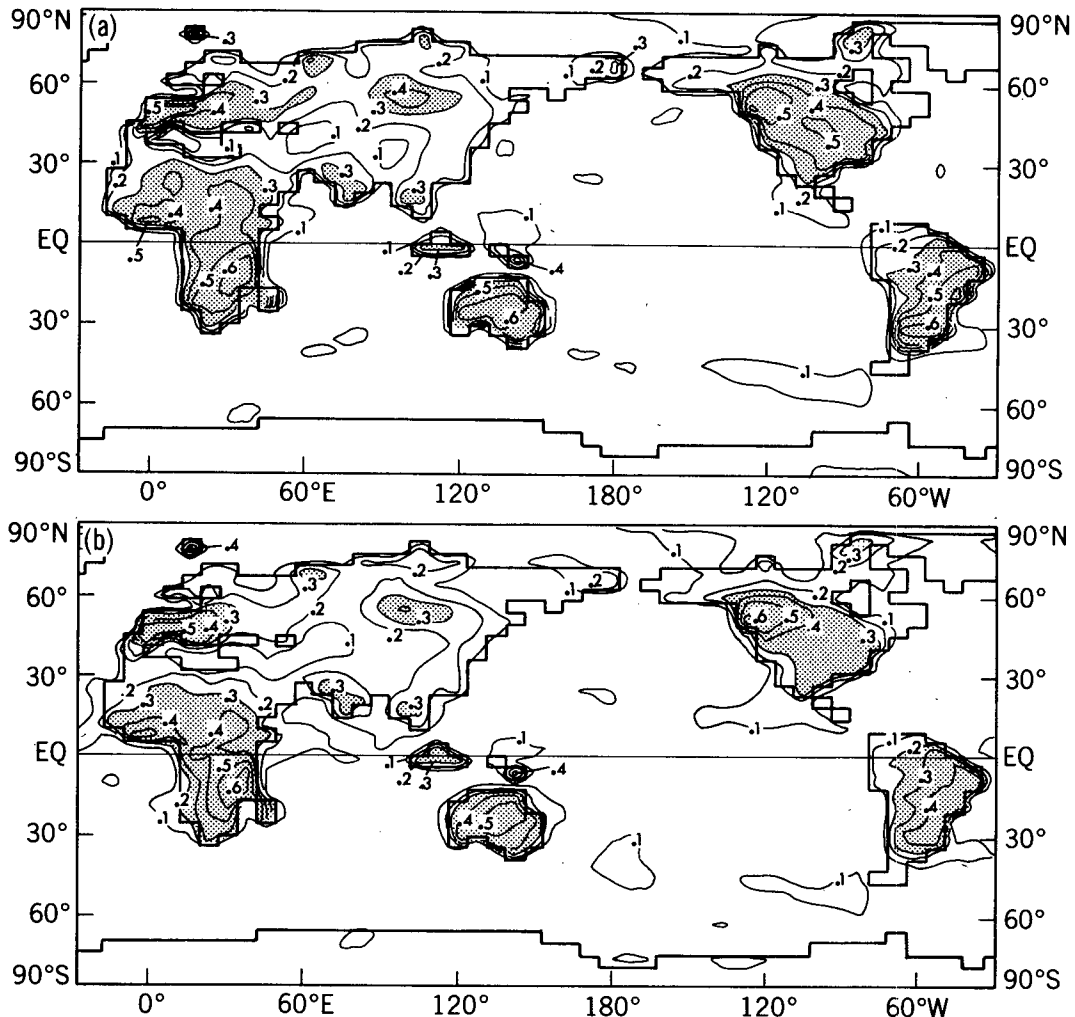


FIG. 7. Lag-one autocorrelation values for the months of JJA from SMI. Statistical significance is the same as in Fig. 3a. Values greater than 0.3 are stippled. (a) Latent heat flux at the surface. (b) Sensible heat flux at the surface.



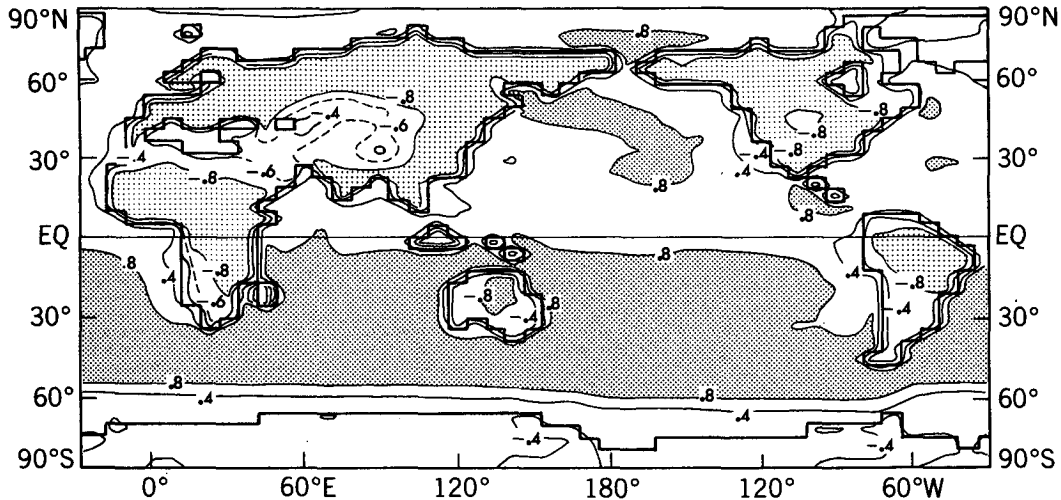


FIG. 8. Correlations between the latent heat flux and the sensible heat flux at the surface for data from the months of JJA. Values greater than 0.8 are densely stippled, while values less than  $-0.8$  are lightly stippled.

related fluxes work together to change relative humidity in the same direction. For example, an increase of soil moisture tends to increase the latent heat flux, and therefore atmospheric moisture, while decreasing the sensible heat flux, and therefore air temperature. The effect of both changes is to increase relative humidity.

As seen from (5), the impact of a change of soil moisture on the latent heat flux, and hence on relative humidity, is directly proportional to potential evaporation. Therefore, the degree to which fluctuations of soil moisture can influence the surface heat fluxes, and thus the atmosphere, depends on the magnitude of potential evaporation. Much of the seasonal and geographic dependence of the influence of soil wetness on the persistence of relative humidity is explained by the seasonal and geographic variations of potential evaporation. For example, potential evaporation rates for JJA are small at high latitudes (Fig. 1b). This means from (5) that fluctuations of soil moisture at high latitudes have little impact on the latent (and sensible) heat fluxes, and thus on the atmosphere. The persistent anomalies of soil wetness at high latitudes are unable to affect relative humidity, which is characterized by low persistence there. By contrast, larger values of potential evaporation during JJA at middle and low latitudes (Fig. 1b) mean that the soil layer can have a substantial effect on the atmosphere. This accounts for the maximum in the persistence of near-surface relative humidity during JJA at middle latitudes of the Northern Hemisphere (Figs. 3a and 5). The low persistence of relative humidity during DJF at middle latitudes of the Northern Hemisphere is similarly explained by the low potential evaporation values during winter.

In regions with extremely large values of potential evaporation, such as the arid interior of Asia in JJA, the persistence of soil wetness is low (see Fig. 1). This results in low persistence of the surface heat fluxes and

relative humidity (see Fig. 3a). Thus, in order for fluctuations of soil wetness to increase atmospheric persistence, potential evaporation must be large enough that fluctuations of soil wetness have an appreciable effect on the latent heat flux, but not so large that the persistence of soil wetness is low.

Fluctuations of soil wetness have little impact on the atmosphere in regions which are frequently saturated (typically regions where the ratio of potential evaporation to precipitation is less than 1, as discussed previously). When the soil layer is saturated, evaporation no longer depends directly on soil wetness, as seen from (1) and (3) [the function  $f(w(t)/w_{FC}) = \text{const} = 1$ ]. Evaporation is at the potential rate, and soil wetness anomalies have no influence on the latent heat flux or the atmosphere. This explains the small autocorrelation values of relative humidity found in regions of frequent runoff. Figure 9 shows the mean percentage saturation of the soil layer in SMI for JJA. A comparison of Figs. 9 and 3a shows that regions with mean saturation greater than 75%, as in southeast Asia, the far northeast of Siberia, and the extreme northern section of South America, have small relative humidity autocorrelations.

As measured by lagged autocorrelations of relative humidity, the strongest influence of the soil layer on atmospheric variability extends at least up to 800 mb, as shown in Fig. 10. Large-scale anomalies of soil moisture could also affect the large-scale circulation of the atmosphere. Results presented later support this possibility by showing that variability in the soil layer increases the persistence of precipitation.

### c. Comparison to observations

It is imperative to compare the model variability results presented here to observations. Unfortunately,

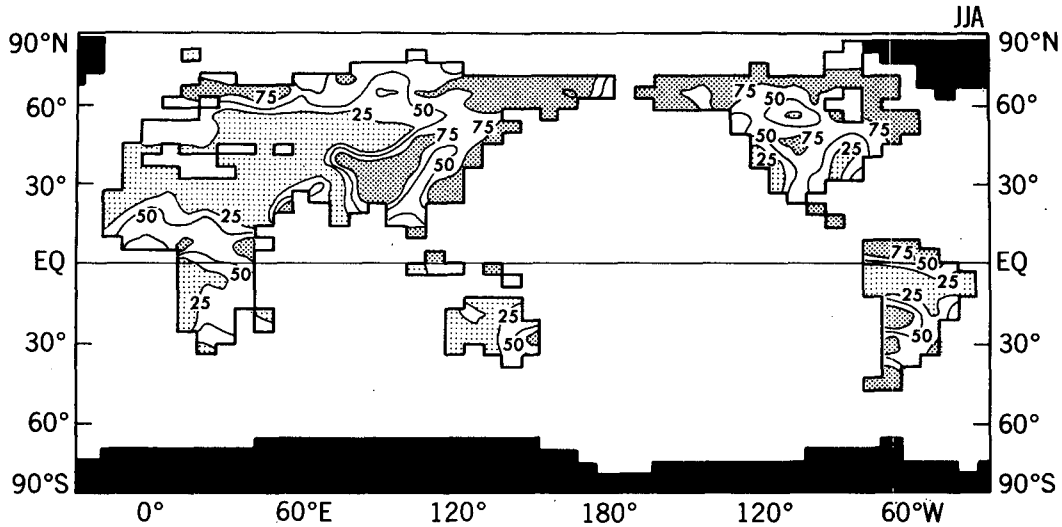


FIG. 9. Mean saturation of the soil layer for SMI during JJA. Values are % saturation  $[100.0 \cdot (\text{time-mean soil moisture} / \text{field capacity})]$ . Values greater than 75% are densely stippled, while values less than 25% are lightly stippled.

measurements of soil moisture suitable for variability analysis are not routinely available. Therefore, the variability of model relative humidity is compared to observations. Surface station data were analyzed over North America for the period 1968–86 using the NCAR World Monthly Surface Station Climatology. Values more than three standard deviations from the long-term monthly means were first removed from the time series. This arbitrary procedure was adopted in order to eliminate obviously erroneous data values from the time series (visual inspection of the time series before and after removal suggest that this procedure had the desired effect). Monthly mean surface pressure, temperature and mixing ratio were then used to compute an index of monthly mean relative humidity in the same manner as previously defined for the model

output. The autocorrelation coefficients at lag-one month were computed using data for the months of April–September and are shown in Fig. 11a. Six months of the year were used instead of the three summer months in order to increase the number of data points.

A comparison of Fig. 11a with the persistence of model relative humidity for April–September in Fig. 11b reveals some broad similarities. Autocorrelations are positive and of comparable magnitude in both the model output and the observations. In both cases there is a general poleward decrease of persistence, consistent with the poleward decline of model potential evaporation. There are two major areas with substantial differences: first, over western Alaska, autocorrelations are positive in the observations, but near zero in the

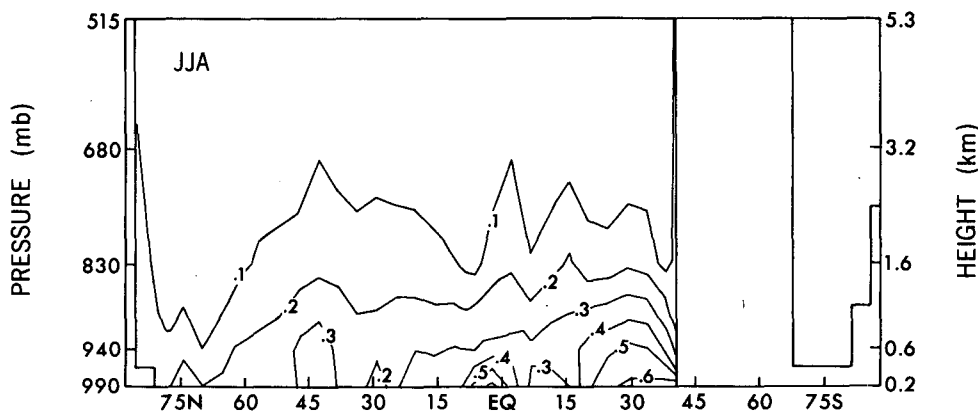


FIG. 10. Zonal means over land of the one-month lagged autocorrelations of relative humidity for SMI during JJA. No data are plotted between approximately 40° and 67°S where there are very few or no land points in the model.

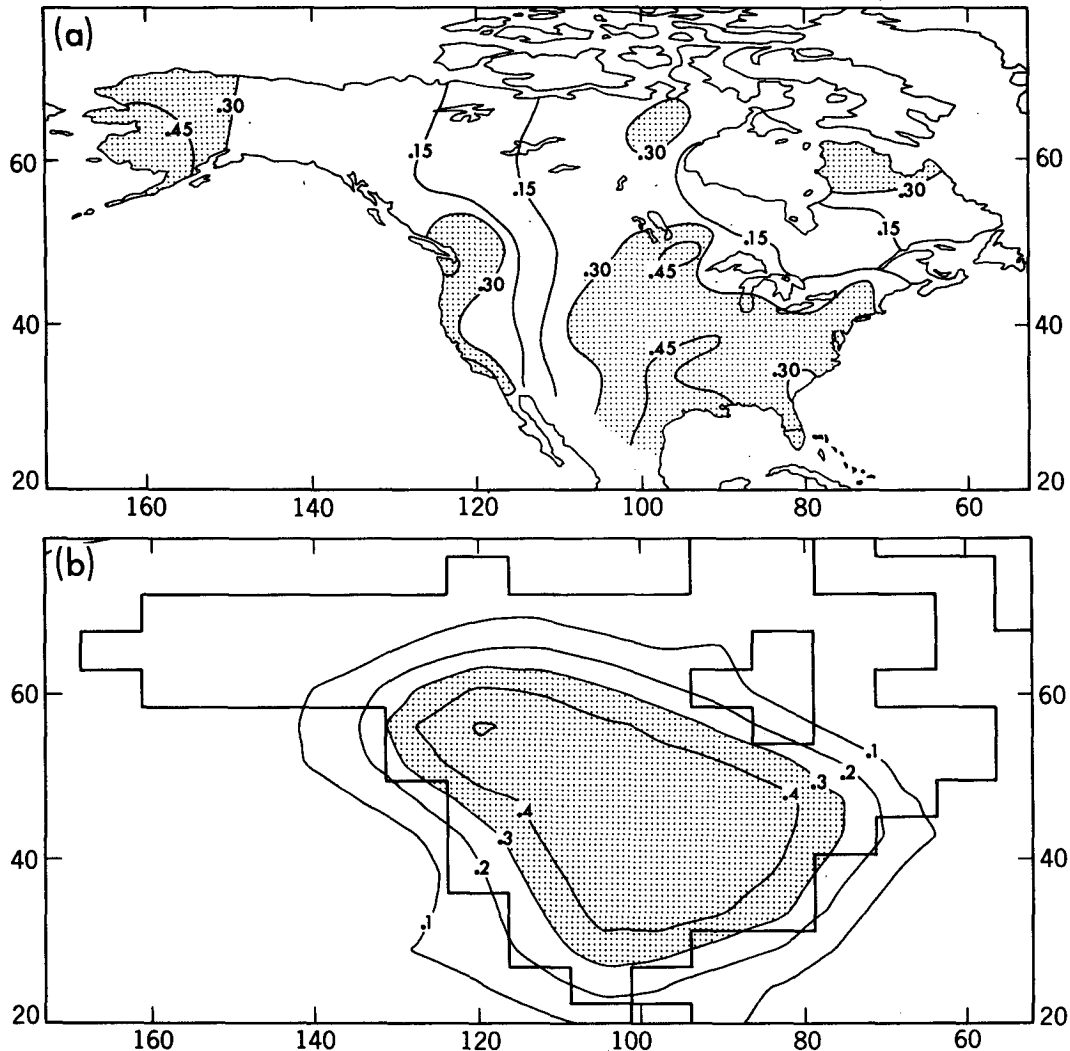


FIG. 11. (a) Lag-one autocorrelations of an index of observed monthly mean relative humidity (index defined in text), April–September for 1968–86. Data are from the NCAR World Monthly Surface Station Climatology. Coefficients greater than 0.18 (0.3) are significantly different from zero at the 95% (99.8%) confidence level. (b) Lag-one autocorrelations of model relative humidity for JJA from SMI. Coefficients greater than 0.11 (0.19) are significantly different from zero at the 95% (99.9%) confidence level.

model output. Second, over most of the western United States, autocorrelations are small in the observations, but large in the model output.

The broad similarities between the observed and simulated variabilities of relative humidity are encouraging. However, in disagreement with the model, substantial autocorrelations were found in the observed data during winter at middle latitudes, suggesting that other factors, such as the influence of the ocean, may also be important.

##### 5. Standard deviation of relative humidity

Fluctuations of soil wetness increase not only the persistence of the surface heat fluxes but their total variability as well. The standard deviations of the surface heat fluxes for JJA are shown in Figs. 12 and 13

for both experiments. The stippling indicates regions where the ratio of the variance of the latent (sensible) heat flux in SMI to the variance of the latent (sensible) heat flux in SMP is greater than one with a 99% confidence level [see Bendat and Piersol (1971) for details of the testing; as a conservative estimate, 50 degrees of freedom were assumed for SMI]. Clearly, fluctuations of soil moisture significantly increase the variability of the surface heat fluxes over most land areas. The only exceptions are regions with small potential evaporation values (high latitudes) and regions which are frequently saturated, such as the extreme northern area of South America. As previously discussed, in regions of frequent saturation evaporation is usually at the potential rate [see (1) and (3)], and anomalies of soil wetness have little impact on the surface heat fluxes.

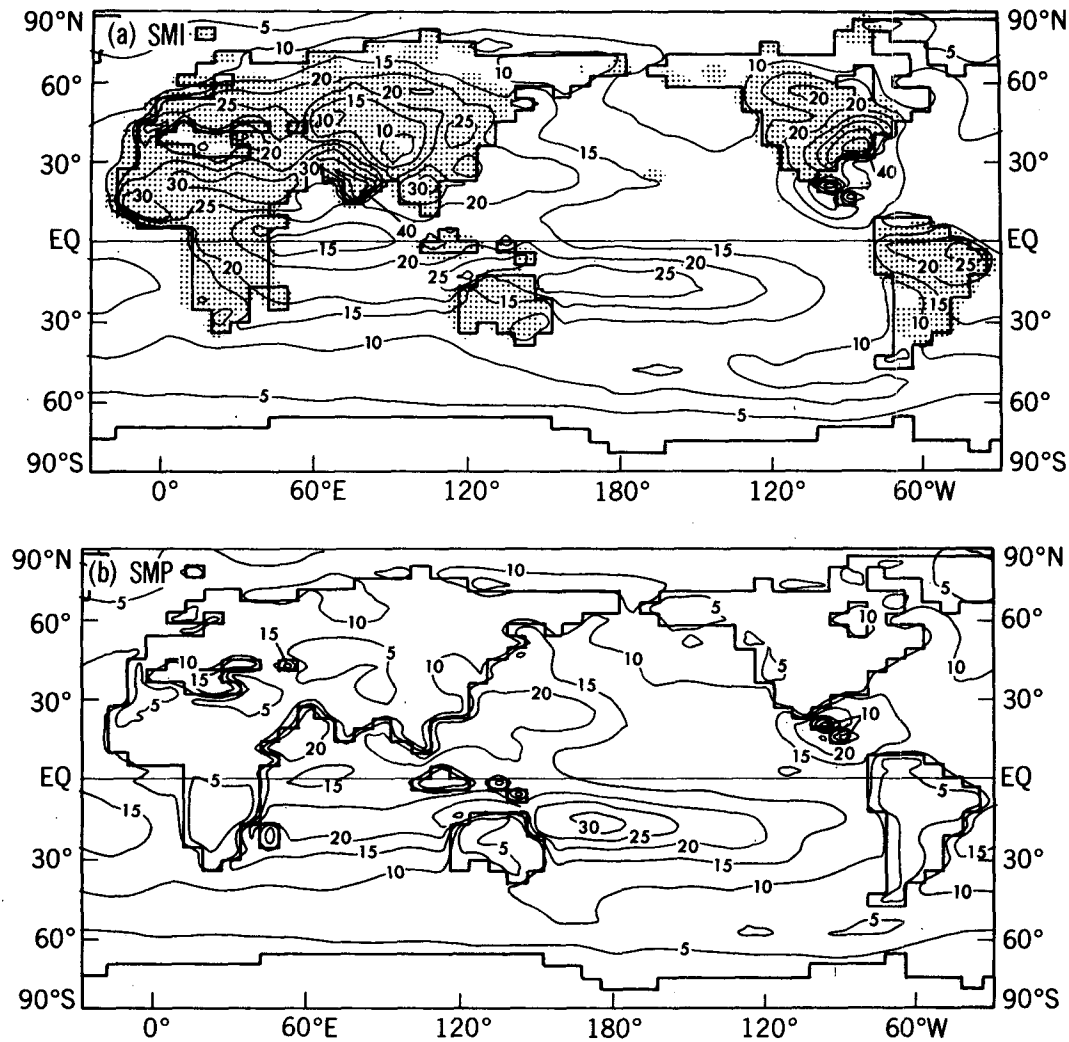


FIG. 12. Standard deviation of the latent heat flux at the surface for JJA. Units are  $\text{W m}^{-2}$ . (a) SMI. Stippling indicates regions where the ratio of the variance of the latent heat flux in SMI to that in SMP is significantly greater than 1 at the 99% confidence level. (b) SMP.

The increased fluctuations of the surface heat fluxes in SMI result in significantly larger fluctuations of relative humidity in SMI relative to SMP over most land areas, as shown in Fig. 14. The only regions where this does not hold are those where the standard deviations of the surface fluxes are not significantly greater in SMI than in SMP, as discussed above.

Over the oceans the standard deviations of the surface heat fluxes and relative humidity are virtually identical in the two experiments. Because sea surface temperatures are the same in SMI and SMP, the surface fluxes and their variations are similar in the two experiments.

## 6. Effect of interactive soil moisture on other variables

### a. Temperature

Fluctuations in soil wetness can also influence surface air temperature. DM demonstrated that interac-

tions between the soil layer and the atmosphere can substantially increase the variance of surface air temperature. These interactions also affect the persistence of surface air temperature. Figure 15 shows the lag-one autocorrelation coefficients of surface air temperature for JJA for both experiments. The statistical significance of individual grid points is the same as in Fig. 3. To test whether the number of points which are significant at the 95% level arose by chance, field significance testing (Livezey and Chen 1983) was performed. For SMP, the results are ambiguous; whether or not we can reject the null hypothesis that the field of autocorrelations arose by chance depends on the number of degrees of freedom in the field, which is a function of the spatial correlations (9.7% of the total land area is significant at the 95% level). Such marginal significance, however, can be contrasted with the results for SMI, where we can reject at the 95% level the null hypothesis that the field of autocorrelations arose by

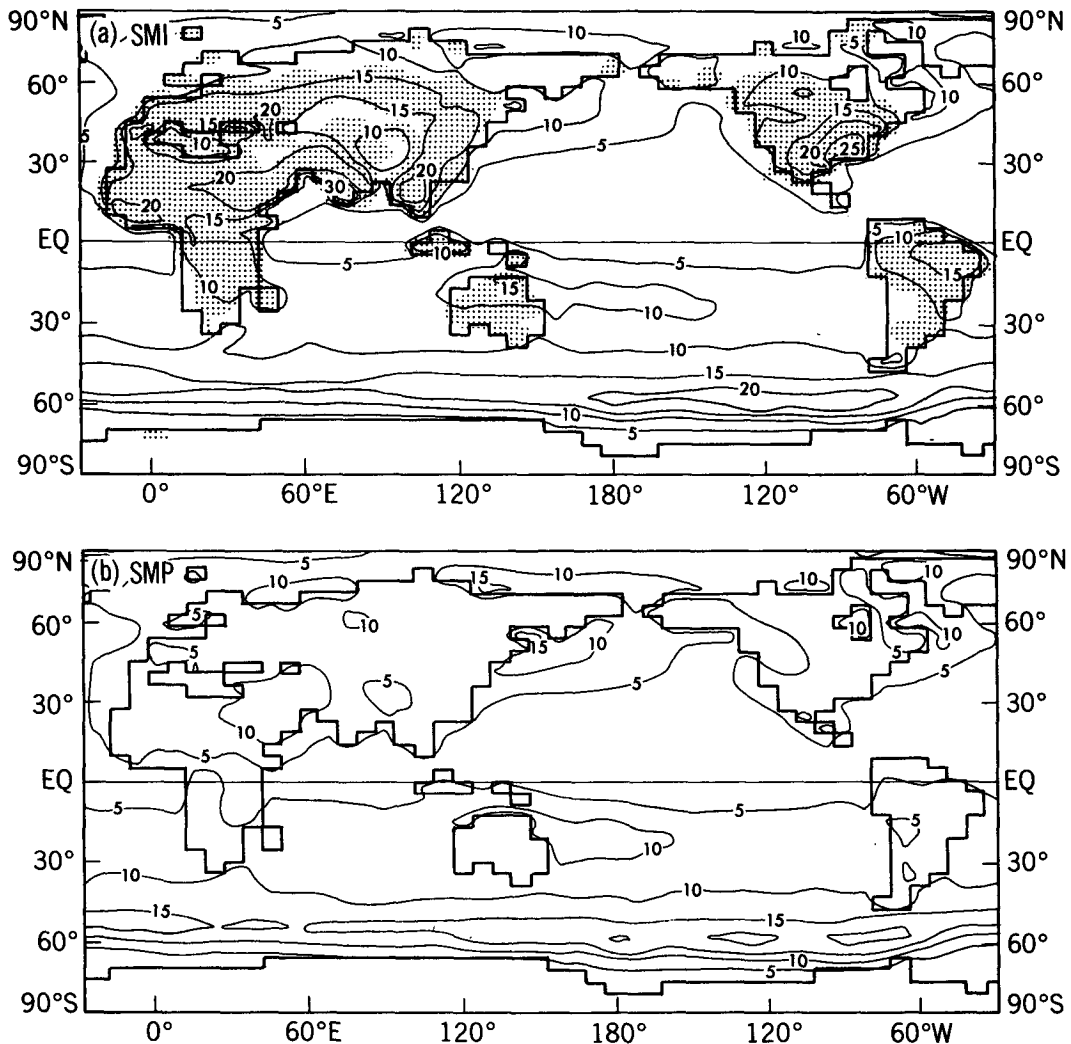


FIG. 13. Standard deviation of the sensible heat flux at the surface for JJA. Units are  $W m^{-2}$ . (a) SMI. Stippling indicates regions where the ratio of the variance of the sensible heat flux in SMI to that in SMP is significantly greater than 1 at the 99% confidence level. (b) SMP.

chance (39.9% of the total land area is significant at the 95% level). Thus, fluctuations of soil moisture increase the persistence of surface air temperature.

#### b. Precipitation

Variations in soil moisture and evaporation also affect the amount and distribution of water vapor in the atmosphere, and hence may affect precipitation. To examine this, time series of precipitation anomalies were analyzed. The time series were first spatially smoothed to remove some of the small-scale spatial variability contained in the precipitation field. A nine-point filter was used, reducing by more than 75% the amplitude of features with spatial scales less than approximately 1500 kilometers. This spatially smoothed time series was then used to compute the lag-one autocorrelation values of precipitation in both experiments for JJA.

The results from SMP show virtually no persistence of precipitation (not shown). As shown in Fig. 16, however, there is a small but clear persistence of precipitation in SMI. With 46.0% of the total land area statistically significant at the 95% level (autocorrelation coefficients greater than 0.16), field significance testing allows us to reject at the 95% level the null hypothesis that the field of autocorrelations arose by chance. Physically, positive (negative) anomalies of precipitation can produce positive (negative) anomalies of soil moisture which increase (decrease) the latent heat flux and decrease (increase) the sensible heat flux. The changes in these fluxes work together, as discussed previously, to increase (decrease) relative humidity, making precipitation more (less) likely. In this manner, the initial anomaly in precipitation is sustained. This process appears to be more prevalent on larger spatial scales. The autocorrelation coefficients computed

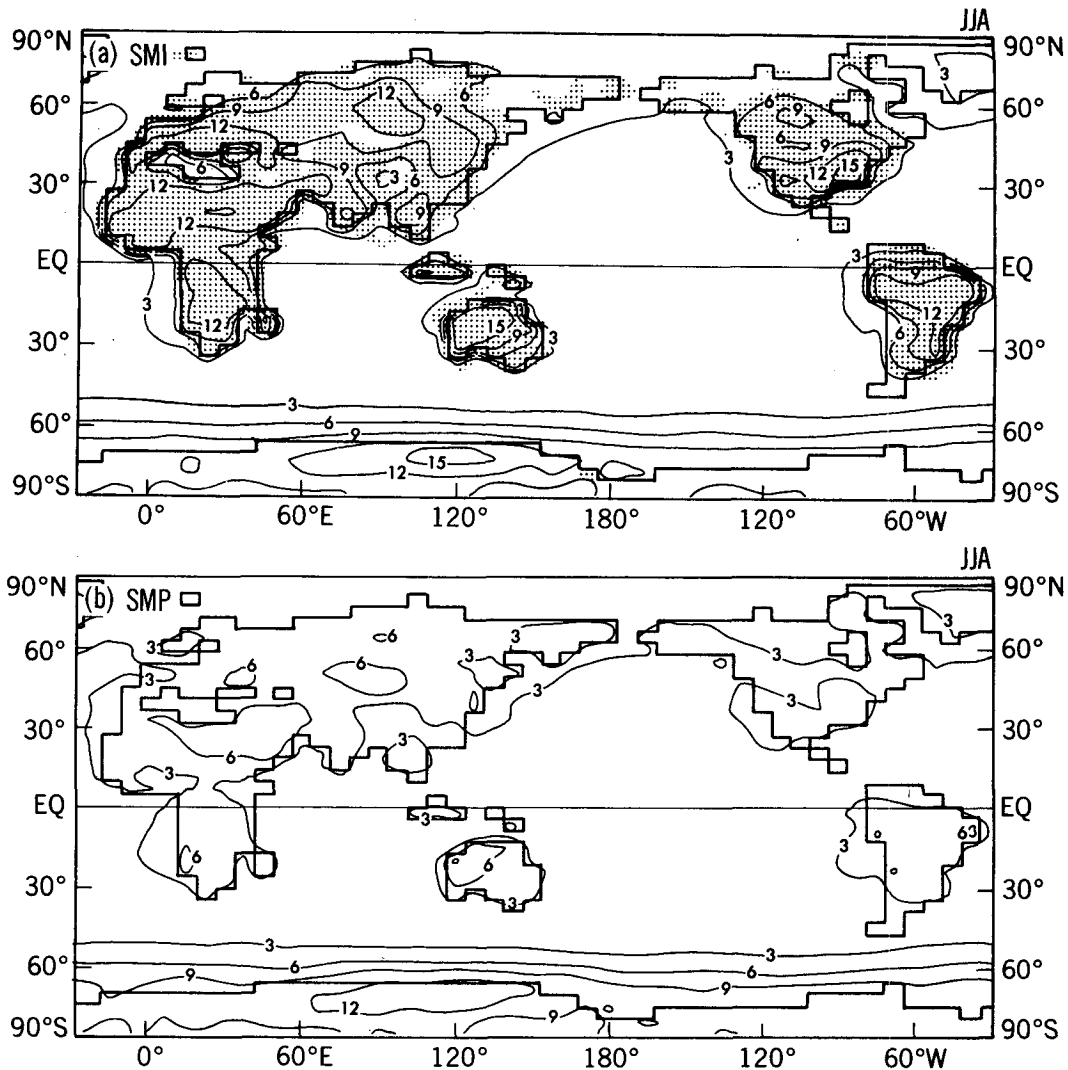


FIG. 14. Standard deviation of the relative humidity index for JJA. Units are percent saturation. (a) SMI. Stippling indicates regions where the ratio of the variance of relative humidity in SMI to that in SMP is significantly greater than 1 at the 99% confidence level. (b) SMP.

without spatial smoothing of the precipitation data were smaller (not shown).

## 7. Summary and discussion

Time series of monthly mean soil moisture computed in a general circulation model contain fluctuations on seasonal and interannual time scales. The soil layer acts as an integrator of monthly mean rainfall. The time scales for fluctuations of soil moisture are largely governed by values of potential evaporation and the ratio of potential evaporation to precipitation. The smaller the value of potential evaporation (which typically decreases poleward), the more slowly anomalies of soil moisture are evaporated, and the longer the time scales of soil moisture anomalies. Where the ratio of potential evaporation to precipitation is less than 1,

the soil is frequently saturated and the time scales of soil wetness are substantially reduced.

The persistence of soil wetness anomalies has a substantial impact on the variability of the lower model troposphere. Persistent anomalies of soil wetness increase both the persistence and total variability of near-surface relative humidity and temperature by altering the surface fluxes of latent and sensible heat. The seasonal and geographical dependence of this influence is strongly determined by the value of potential evaporation. Because the latent heat flux is directly proportional to the value of potential evaporation, the effect of a change in soil moisture on the latent heat flux, and thus on the atmosphere, is proportional to potential evaporation. In winter and at high latitudes, potential evaporation values are small due to weak insolation. Fluctuations of soil moisture thus have little impact

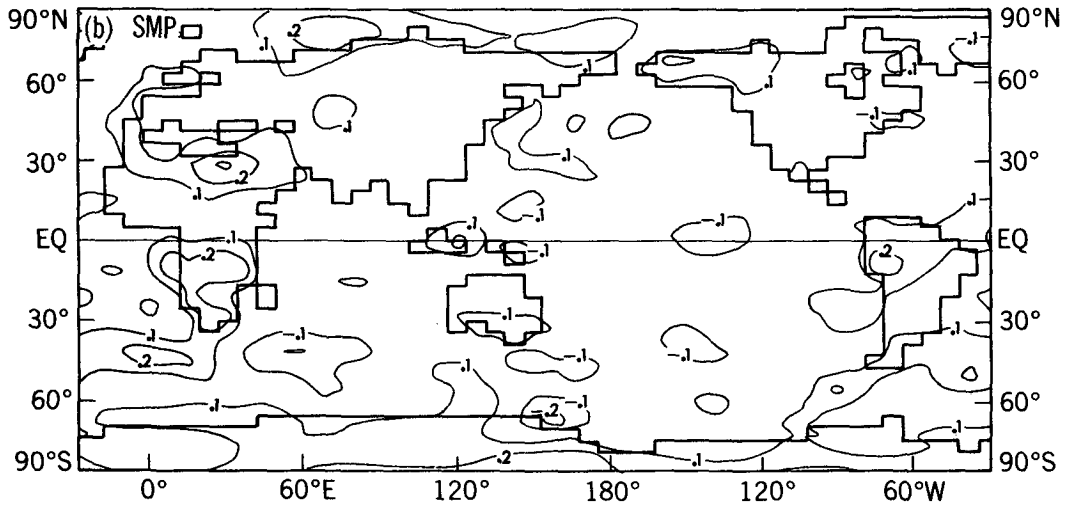
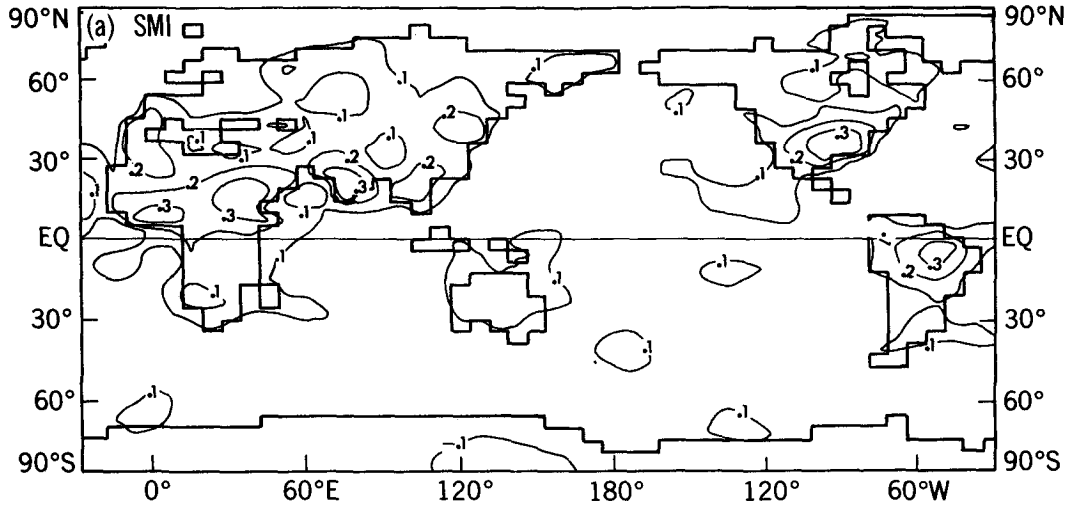


FIG. 15. One-month lagged autocorrelation of surface air temperature for JJA. (a) SMI. Statistical significance the same as in Fig. 3a. (b) SMP. Statistical significance the same as in Fig. 3b.

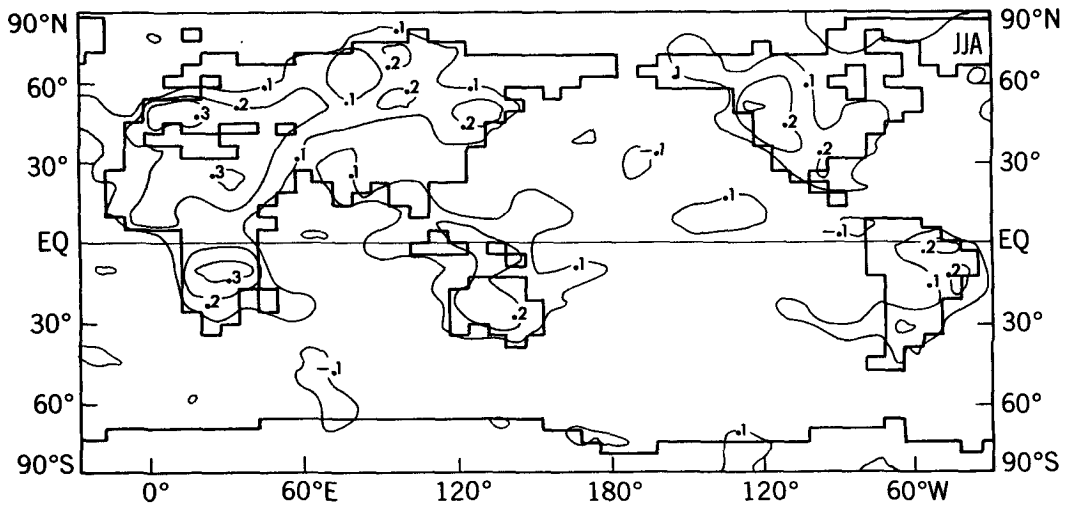


FIG. 16. One-month lagged autocorrelations of precipitation (JJA) for SMI. Statistical significance the same as in Fig. 3a.

on the latent heat flux, and consequently on the atmosphere, in these seasons and regions. In the tropics and during the summer season, however, larger values of potential evaporation allow fluctuations of soil moisture to have a substantial effect on the variability of the lower atmosphere.

The influence of the soil layer on atmospheric variability depends not only on potential evaporation, but on the variability of soil wetness as well. In regions with extremely large values of potential evaporation, such as the arid interior of central Asia during JJA, the persistence of soil wetness is low. Consequently, the persistence of the surface heat fluxes and near-surface relative humidity is also low.

At the other extreme are regions that are frequently saturated. Such regions are characterized by potential evaporation values less than the mean precipitation rate, resulting in frequent saturation and runoff. Under such conditions, evaporation is almost always at the potential rate, independent of soil wetness. Fluctuations in soil wetness in such regions have little impact on evaporation, the latent heat flux, or on atmospheric variability.

Thus, land surface processes have the potential to make a critical impact on atmospheric variability. Such an influence has many implications, one of which is for seasonal climate forecasts. As discussed by Rind (1982) and extended by results presented here, soil wetness conditions in the spring can have a substantial influence on the summer climate at middle latitudes.

It should be noted that in the present study the seasonal cycle of sea surface temperatures was prescribed. Interactions between the oceanic mixed layer and the atmosphere, which can also have a substantial influence on atmospheric variability, are thus not present in this model. Another model simplification is that cloudiness is fixed. Furthermore, the simulation of the global distribution of climate obtained from the present model is far from satisfactory. Thus, the geographic distribution of climate variability obtained from the present study should not be taken too literally.

The formulation of land surface processes in the present model is highly idealized, and this should also be kept in mind. Despite these idealizations, the present model at least contains the most fundamental processes controlling the heat and moisture budget at the continental surface. Indeed, we believe that the simplicity of the land surface formulation facilitates the elucidation of some of the basic mechanisms involved in

the variability of the land surface and overlying atmosphere.

The persistence and variability of near-surface climate obtained from the present study should be thoroughly compared with observations. Such a comparison is essential for establishing the credibility of the present study.

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