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Climate variability and land-surface processes

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The coupled ocean–atmosphere–land climate system is characterized by substantial amounts of variability on a wide range of spatial and temporal scales. This natural variability of climate increases the difficulty of detecting climate change attributable to increasing greenhouse gas concentrations. A key issue in climate research is obtaining a better description of this variability and the physical mechanisms responsible for it. One of the important physical processes contributing to this variability is the interaction between the land surface and the atmosphere. Through its effect on the surface energy flux components, the land surface can exert a pronounced effect on the variability of the atmosphere. The potential importance of such interactions for climate variability is examined through the use of numerical modeling studies. The physical mechanisms governing the time scales of soil moisture variability in the model are outlined, and observational evidence is presented supporting this analysis. In addition, it is shown that interactions between soil wetness and the atmosphere can both increase the total variability of the atmosphere and lengthen the time scales of near-surface atmospheric fluctuations.

Key words: Climate model, atmospheric variability, soil moisture (influence of soil moisture on atmospheric variability), atmosphere, potential evaporation, land surface.

1 INTRODUCTION

The coupled ocean–atmosphere–land climate system is characterized by substantial amounts of variability on a wide range of spatial and temporal scales. This natural variability increases the difficulty of detecting climate change attributable to increasing greenhouse gas concentrations. A key goal of climate research is to obtain a better description of this variability and the physical mechanisms responsible for it. A wide variety of physical processes contributes to the observed natural variability. Interactions between the ocean and the atmosphere have a very important impact on atmospheric variability on a wide range of temporal scales. Interactions between patterns of sea surface temperature anomalies and the atmosphere can have a pronounced effect on the structure of atmospheric planetary waves. For example, the El Niño/Southern Oscillation phenomenon in the tropical Pacific contributes substantially to atmospheric variability on interannual time scales, and has been shown to have a pronounced influence on hemispheric and global mean temperatures.[10,18] At interdecadal time scales, interactions between the deep ocean circulation and the atmosphere can be very important.[6] In particular, modeling studies have shown fluctuations in the ocean's thermohaline circulation on decadal and longer time scales,[5,24] which may be quite important for climate variability.

Over continental regions, interactions between the land surface and the atmosphere are potentially important to climate variability. Through its influence on the outgoing latent and sensible heat fluxes, the state of the land surface can have a profound influence on the heat flux that the atmosphere feels, and thus on the state of the atmosphere. Variations in the state of the land surface may therefore contribute substantially to atmospheric variability. Within this context, one of the key questions in climate research is to achieve an accurate representation of the hydrologic state of the land surface, its variations in time and space, and how those variations influence the atmosphere and its variability. It will be shown in this paper that variations in soil wetness in a numerical model can have substantial persistence on seasonal and interannual time scales, and that these variations have a profound impact on near-surface atmospheric variability in the model.

The low frequency nature of soil wetness variability has been recognized for some time. Namias,[16,17] in studies of persistence of atmospheric circulation, noted that seasonal anomalies of soil wetness could have an impact on the seasonal cycle of the atmosphere. Subsequent studies of soil wetness, frequently utilizing general circulation models (GCMs), have examined the interactions between persistent anomalous soil wetness conditions and the atmosphere.

Shukla and Mintz[21] demonstrated the substantial impact of soil wetness on climate by examining the differences in climate simulated with a completely wet land surface and a completely dry land surface. Substantial differences were seen in surface air temperature, surface pressure and precipitation, especially in the tropics and summer hemisphere extratropics. They also suggested that time-varying soil wetness has an influence on the annual cycle of the atmosphere. Shukla and Mintz[21] hypothesize that 'in the extratropics, with its large seasonal changes, the soil plays a role analogous to that of the ocean. The ocean stores some of the radiational energy it receives in summer and uses it to heat the atmosphere over the ocean in winter. The soil stores some of the precipitation it receives in winter and uses it to humidify the atmosphere in summer.'

Walker and Rowntree[23] and Rowntree and Bolton[20] performed model studies over Africa and Europe respectively. Their results demonstrated not only the time-mean effect of persistent soil wetness anomalies on the atmosphere, but also showed that the atmosphere may respond to these anomalies in such a way as to perpetuate the initial soil wetness anomalies. The time scale for such persistence appears to be at least on the order of weeks.

Rind[19] examined the importance of soil wetness anomalies on summertime model predictability over North America. Given an early summer anomaly of soil wetness over the entire United States, he found the subsequent summer model conditions to be statistically different from a control run. He concluded that '. . . a knowledge of the ground moisture at the beginning of summer might allow for improved summer temperature forecasts. . .'.

Yeh *et al.*,[25] using a model with idealized geography, also discussed the atmospheric response to initially prescribed soil wetness anomalies. In addition to describing a feedback process which may have helped to prolong the initial soil wetness anomalies, they noted that the persistence of soil wetness anomalies depends significantly upon the latitude.

Gordon and Hunt[7] were among the first to explicitly study the variability of soil moisture as simulated by an atmosphere–land model. They found that in their model with a prescribed seasonal cycle of sea surface temperatures, droughts occurred on an interannual time scale. On much shorter time scales, Hammarstrand[8] has shown a sensitivity of the atmmosphere to soil wetness.

In this paper, some of the results of a previous study of the variability of soil wetness and its impact on climate[3,4,13] are reviewed and placed within the more general context of climate variability. These results highlight the potential importance of land-surface processes to the variability of climate, and stress the importance of obtaining an adequate representation of land-surface processes in climate models in order to obtain a realistic simulation of atmospheric variability. In this study, output from two integrations of a GCM are analyzed, one of length 50 years in which soil moisture is computed interactively, and one of length 25 years in which the seasonal cycle of soil moisture is prescribed. The analysis of the variability of soil wetness in these models, the physical processes governing that variability, and the impact of that variability on the atmosphere, serves as an important indicator of the potential impact of the land surface on climate variability. Differences in atmospheric variability between the two experiments are a measure of the impact of interactions between the land surface and the atmosphere on atmospheric variability. While the model land surface formulation is extremely simple, general conclusions may be drawn about the relevance of land-surface processes for climate variability, and the physical mechanisms governing the land surface and its interactions with the atmosphere.

2 MODEL DESCRIPTION AND EXPERIMENTAL DESIGN

The model consists of two parts: (i) a general circulation model (GCM) of the atmosphere, and (ii) a heat and water balance model over the continents. The atmospheric GCM is described by Delworth and Manabe.[3] The spectral computations employ the 'rhomboidal 15' wavenumber truncation, such that the resultant transform grid has an approximate horizontal resolution of 7·5° longitude by 4·5° latitude. There are nine finite difference levels in the vertical. The seasonal cycles of sea surface temperature and sea ice are prescribed at all ocean grid points based upon observed monthly mean fields.

The distribution of incoming solar radiation at the top of the atmosphere is prescribed. There is a seasonal cycle of incoming solar radiation, but no diurnal cycle. The mixing ratio of carbon dioxide is assumed to be constant everywhere, whereas ozone is specified as a function of latitude, height and season. Cloud cover is prescribed to be zonally uniform and invariant with respect to season, depending only on latitude and height.

While the model employs the spectral transform method for the mass, temperature, and momentum fields, a separate set of finite difference computations was employed for atmospheric water vapor. This methodology partially alleviates the problems inherent in the spectral representation of atmospheric water vapor, and results in an improved simulation of precipitation. For a more complete discussion of this technique, see appendix A of Ref. 3.

Precipitation is predicted whenever supersaturation occurs in the model. This supersaturation can be the result of either large-scale condensation or convective adjustment (see Ref. 14 for details of the convective adjustment scheme). If precipitation occurs while the air temperature just above the surface is below freezing, the precipitation falls as snow; otherwise, any precipitation is assumed to fall as rain.

A snow budget is computed at the land surface in which a change of snow depth is predicted as the net contributions from snowfall, sublimation, and snowmelt, with the last two determined from the surface heat budget. The surface heat budget, in turn, is strongly influenced by snow cover, which reflects a large fraction of the insolation.

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.'[12] Changes in soil moisture are computed from the rates of rainfall, evaporation, snowmelt, and runoff from:

$$dw(t)/dt = r_a - E + S_m - r_f \qquad (1)$$

where t is time, w is soil moisture (cm), r_a is rainfall (cm s^{-1}), E is evaporation (cm s^{-1}), S_m is snowmelt (cm s^{-1}), and r_f is runoff (cm s^{-1}). The evaporation is determined

$$E = E_p f(w(t)/w_F) \qquad (2)$$

where w_F is the constant field capacity, assumed to be 15 cm everywhere, and E_p is potential evaporation (cm day^{-1}), determined as

$$E_p = -\rho C_d \mid v_9 \mid (q_9 - q_s(T_*)) \qquad (3)$$

where ρ is the density of the air (g cm^{-3}), C_d is the drag coefficient (constant), v_9 is the wind speed (cm s^{-1}) at the lowest model level (about 85 m above the surface), q_9 is the mixing ratio at the lowest model level (g/g), and $q_s(T_*)$ is the saturation mixing ratio (g/g) corresponding to the ground surface temperature (T_*, units are °C). The function f has the form:

$$f(w(t)/w_F) = w(t)/(0{\cdot}75 w_F) \quad \text{if } w(t) \le 0{\cdot}75 w_F$$
$$= 1 \quad \text{if } w(t) > 0{\cdot}75 w_F \qquad (4)$$

If the computed soil moisture exceeds the field capacity (15 cm), the difference between the computed soil moisture and field capacity is defined as runoff, and the soil moisture is then set equal to the field capacity. As specified by eqn (2), evaporation from the soil is determined as a product of the potential evaporation rate and a function of soil wetness (eqn (4)). This function incorporates the observation that evapotranspiration is at the potential rate (i.e. not limited by soil moisture, with the result that $E = E_p$) when soil moisture is above some critical threshold (75% saturation in this parameterization), but 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. Although the interactions between the soil layer and the atmosphere are present in the first integration, they are not present in the second integration. Differences in atmospheric variability between the two experiments therefore indicate the effect on atmospheric variability of interactions between soil wetness and the atmosphere.

The prescribed soil wetness and surface albedo values used in the second integration were derived from the results of the first integration by the following procedure. At each land point, 5-year means for $f(w(t)/w_F)$ and surface albedo were computed for each 5-day period of the year. From the 73 5-day means, Fourier coefficients were computed and used to derive daily values of $f(w(t)/w_F)$ and surface albedo. These daily values are then used to prescribe soil wetness and surface albedo in the second integration. The potential evaporation rate is computed in the same manner as in the first integration (see eqn (3)). 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 SURFACE VARIABILITY

3.1 Soil wetness spectra

Before examining the influence of the interactions between soil moisture and the atmosphere on atmospheric variability, one should be familiar with the variability of soil moisture itself. We first use spectral analysis to describe the temporal fluctuations of model-computed soil moisture. Anomaly time series of monthly mean soil moisture were generated at each grid point by subtracting the appropriate ensemble monthly mean soil moisture values from the individual monthly mean soil moisture values. The same procedure was performed for precipitation (P), defined as the sum of rainfall and snowfall, as well as for the time series of the sum of rainfall and snowmelt (rs). This rs time series may be viewed as the actual forcing term of the soil moisture variations as seen in eqn (1). Spectral analysis was performed on these time series at each grid point.

Table 1. Definition of latitudinal bands used to average spectra of soil moisture, rainfall and snowmelt

Latitude band	Range

The annual cycle and its harmonics were removed from each spectrum, after which the spectra were normalized by their respective total variances; this process allows spectra from different regions to be averaged without the spectra from high variance regions overwhelming the low variance spectra in the averages. The soil moisture, P and rs spectra were then zonally averaged over all land points. The zonally averaged spectra were further averaged into four bands defined in Table 1. The bands were defined based on similarities of spectral shape, mean precipitation values and soil moisture values. These band-averaged spectra, representative of the large-scale latitudinal spectral variations, are shown in Fig. 1.

The most basic feature of the soil moisture spectra is their resemblance to red noise. In contrast, the P and rs spectra bear a resemblance to white noise. Further, the 'redness' of the soil moisture spectra increases with latitude, while there is little variation with latitude in the P or rs spectra (although there appears to be a slight suggestion of redder spectra in the subtropical and midlatitude bands). One interesting point in the high latitude spectra is the difference between the P and the rs spectra. Because snowmelt is concentrated in a 2- to 3-month period in the spring, more of the total variance of the rs spectrum is located at higher frequencies relative to the P spectrum.

A very prominent feature is the long time scale associated with all the soil moisture spectra. Large amounts of variance are located at periods of one year or more, suggesting that soil moisture may play a role in low-frequency atmospheric variability. With this possibility, it is desirable to understand the mechanisms by which this low-frequency soil moisture variability is generated, and to assess its contribution to the overall climatic spectrum.

3.2 Relation to first order Markov process

The persistence of monthly mean soil moisture may be viewed as the red-noise response of the soil layer to the forcing time series of monthly mean rainfall plus snowmelt, which resembles white noise (since the autocorrelations of the forcing time series are close to zero). Red noise can be generated by a first-order Markov process $y(t)$, which is determined by:

$$dy(t)/dt = -\lambda y(t) + z(t) \qquad (5)$$

where λ is a damping constant and $z(t)$ is a white noise forcing. Physically, this represents a system possessing an inherent exponential damping, but which is continually forced by some random (white noise) process. It is important to note that the characteristic time scale of the input time series $z(t)$ is much shorter than the characteristic time scale of the response $y(t)$[9]. The spectral response $Y(\omega)$ of such a system is given by

$$Y(\omega) = F/(\omega^2 + \lambda^2) \qquad (6)$$

where F denotes the variance spectrum of the white noise forcing, ω is angular frequency, and λ is the damping constant from eqn (5). For a red noise process, the larger (smaller) the value of the damping constant λ, the shorter (longer) the time scales of variation. The autocorrelations of a first-order Markov process are given by

$$r(t) = \exp(-\lambda t) \qquad (7)$$

where $r(t)$ is the autocorrelation at lag t and $(1/\lambda)$ is the e-folding time of anomalies in the absence of forcing.

Where the time-mean potential evaporation ($\bar{E}_p$) is greater than the time-mean precipitation rate $\bar{P}$ (i.e. $\bar{E}_p/\bar{P} > 1$), the soil is seldom near saturation, runoff is infrequent, and evaporation can be approximated from eqns (2) and (4) as

$$E = [E_p/0{\cdot}75w_F)]\, w(t) \qquad (8)$$

In such regions, the soil moisture parameterization (1) can be approximated as:

$$dw(t)dt = -(E_p/0{\cdot}75w_F)\, w(t) + rs \qquad (9)$$

There is a clear analogy between eqns (5) and (9). The damping constant λ in eqn (5) corresponds to the $[E_p/(0{\cdot}75w_F)]$ term in eqn (9). The larger the value of $[E_p/(0{\cdot}75w_F)]$, the more rapidly anomalies of soil moisture are removed from the land surface by evaporation, and the shorter the time scales of soil moisture variations. The time series of the surface water supply (i.e. the total contribution from both rainfall and snowmelt) in eqn (9) corresponds approximately to the white noise forcing term in eqn (5). The soil layer acts as an integrator of the time series of the surface moisture input, producing a time series of soil moisture that is similar to red noise (lag-one autocorrelations greater than zero). The time scales of soil moisture variations are thus determined by the values of potential evaporation and field capacity.

In contrast, where $\bar{E}_p/\bar{P} < 1$, the soil is frequently saturated, and there is considerable runoff. Under such conditions, evaporation is at the potential rate (see eqns (2) and (4)), and there is no feedback between the evaporation rate and soil moisture. Under such conditions, soil moisture does not behave as a red noise process. The time series of soil moisture resembles the time series of the input water supply at the surface (rainfall plus snowmelt), and is characterized by very short time scales.

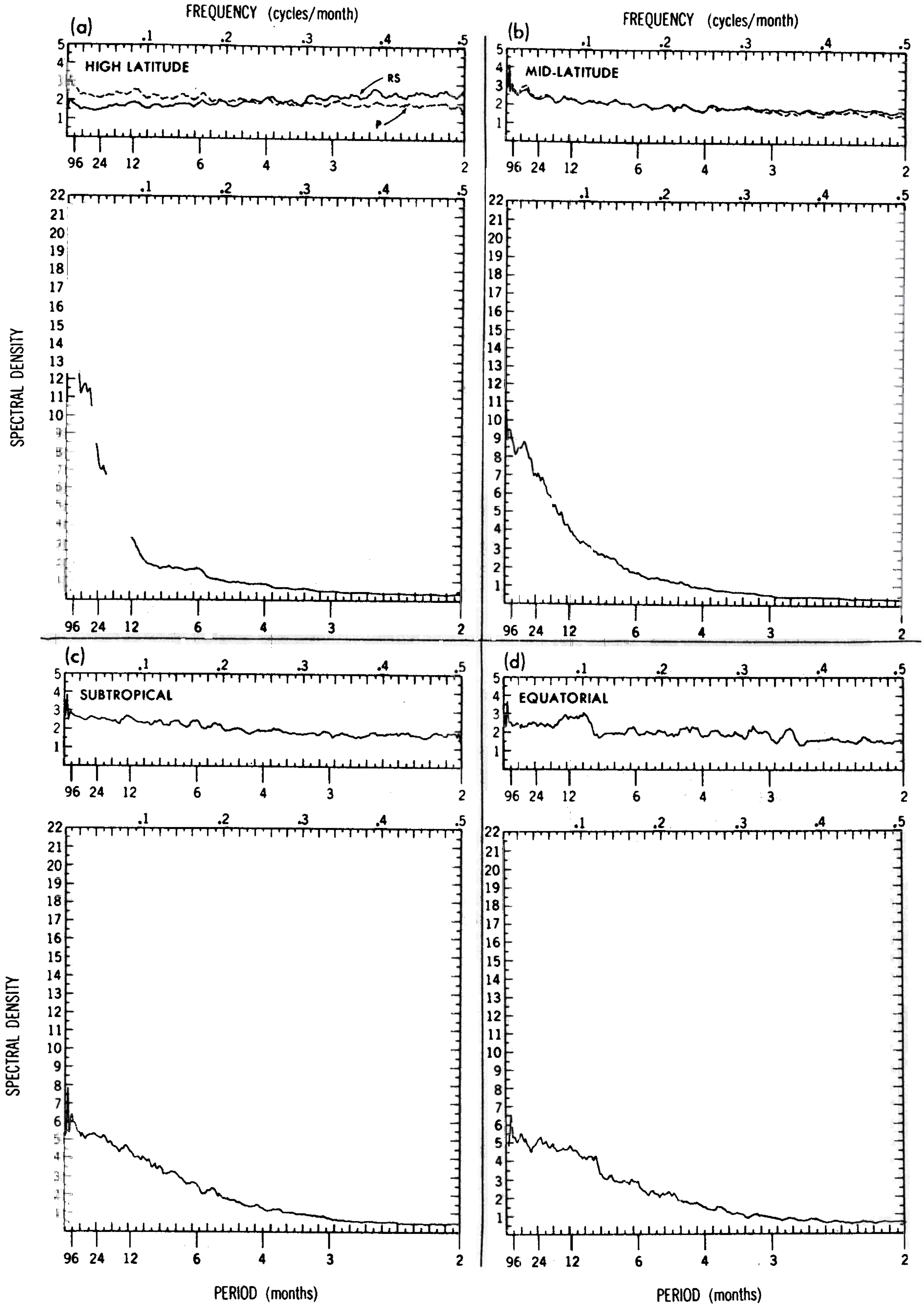

Fig 1. Composite spectra of soil moisture (solid lines in large boxes), rainfall plus snowmelt (solid lines in small boxes), and rainfall plus snowfall (dashed lines in small boxes).[3] See text for details of compositing. (a) High latitude band (see Table 1 for definition); (b) middle latitude band; (c) subtropical band; (d) equatorial band.

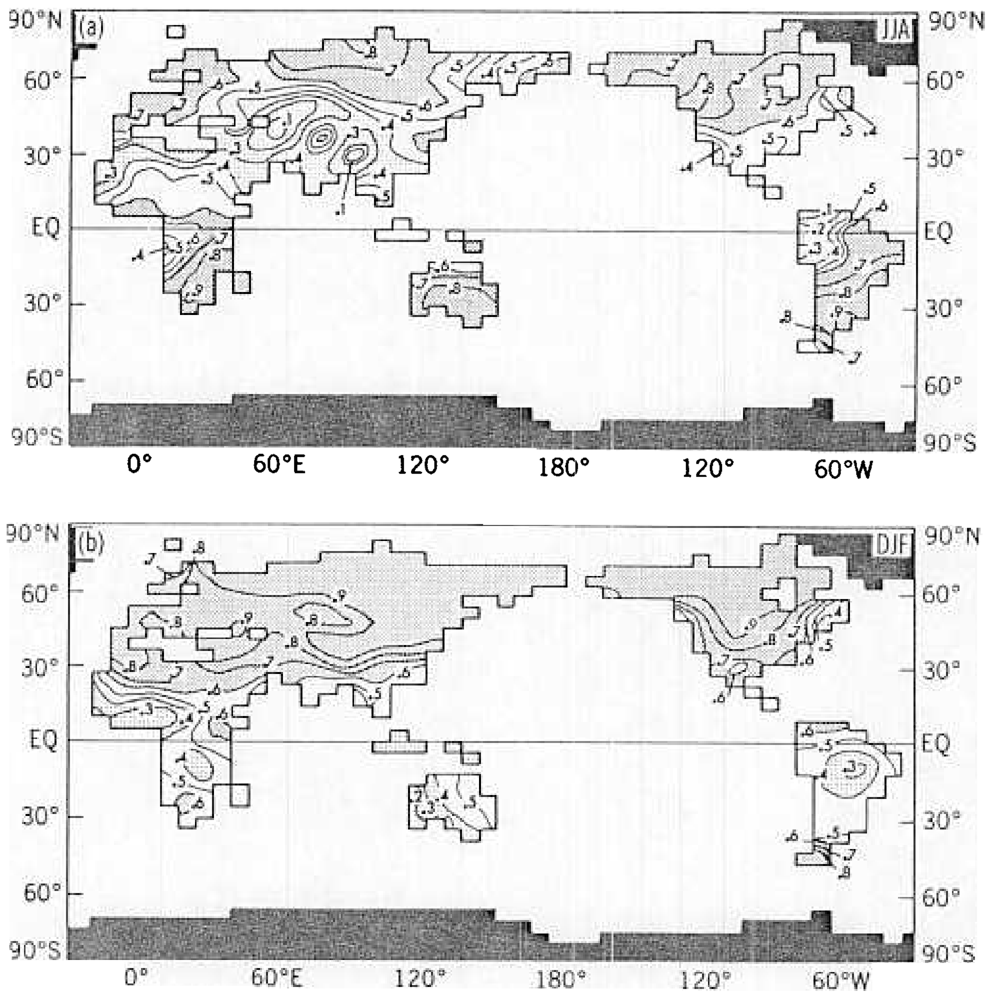

Fig. 2. (a) Lag-one autocorrelation values of soil moisture for the months of June, July and August (JJA) for SMI.[4] At each grid point, deviations of monthly mean soil moisture from the long-term mean for that month were correlated with data from the same grid point, but lagged one month. Coefficients greater than 0·16(0·3) are significantly different from zero at the 95% (99·9%) confidence level (see Ref. 2, 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) As for (a), but using values from December, January and February (DJF).

It is important to note that the resemblance between precipitation and white noise is only valid for time scales longer than about one week. It will be shown later that while there is some persistence of monthly mean anomalies of precipitation, this persistence is relatively small and primarily due to the influence of soil wetness anomalies.

The spatial dependence of the time scales of soil moisture variations can be seen by computing the lag-one autocorrelation coefficient. At each grid point, the time series of monthly mean soil moisture was serially correlated with a one month lag. A map of the autocorrelation coefficients, computed using model output from the months of June, July and August (JJA), is plotted in Fig. 2(a) for experiment SMI. The autocorrelations are generally positive, demonstrating that anomalies of soil moisture persist on monthly time scales. Autocorrelations range 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. Soil moisture autocorrelations for December–January–February (DJF) are shown in Fig. 2(b). A comparison between this map and Fig. 2(a) indicates that, in general, the persistence of soil moisture is larger in winter than in summer, a result of smaller potential evaporation values in winter when insolation is weak (shown below).

The inverse relationship between potential evaporation and the time scales of soil moisture variations discussed above can be seen by comparing Fig. 2 with Fig. 3, which shows maps of potential evaporation derived from the model output using the months of June, July and August (Fig. 3(a)) and December, January and February (Fig. 3(b)). There is a clear inverse relationship between the values of potential evaporation in Fig. 3 and the autocorrelations of soil moisture in Fig. 2 (this relationship is expected from the above discussion of the resemblance of soil moisture to a red noise process). Smaller potential evaporation values

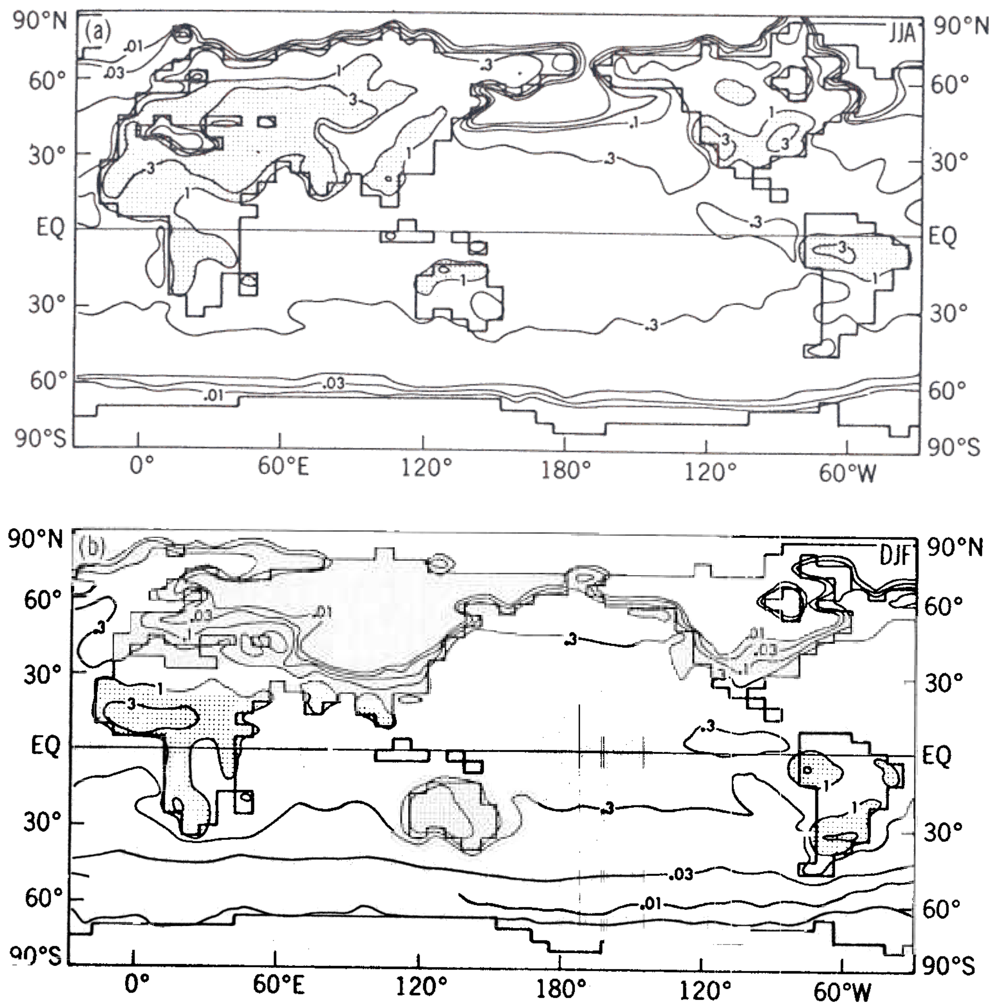

Fig. 3. (a) Potential evaporation (cm day^{-1}) for JJA in experiment SMI.[4] (b) Potential evaporation (cm day^{-1}) for DJF in experiment SMI.

at higher latitudes and in the winter season (i.e. regions and seasons with weak insolation) result in low evaporation rates and large autocorrelations of soil moisture. Hot, arid regions in the continental interiors during summer have large potential evaporation values, and thus very small persistence of soil moisture. The potential evaporation values for DJF, shown in Fig. 3(b), can be contrasted with Fig. 3(a). 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 for DJF. Assuming that the time series of soil moisture is similar to red noise, the autocorrelation values in Fig. 2 can be translated into *e*-folding times. Using eqn (7), one-month lagged autocorrelation values of 0·8, 0·6 and 0·4 correspond to *e*-folding times of 4·5, 2·0 and 1·1 months, respectively. It should be noted that using monthly-averaged soil moisture to estimate the autocorrelations and *e*-folding times typically yields larger autocorrelations and longer *e*-folding times than would have been computed from daily data. The monthly averaging acts as a low pass filter. This is appropriate since the focus of this study is on seasonal to interannual variability.

As discussed above, one other factor strongly influencing the persistence of soil moisture is the ratio of the time-mean potential evaporation to the time-mean precipitation rate, shown in Fig. 4 for June–August. Where this ratio is less than one, evaporation alone cannot balance precipitation and the soil is frequently saturated, resulting in the runoff of excess water. From eqns (2) and (4), 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 during JJA in the extreme northern part of South America and the extreme northeastern region of Siberia (the value of soil moisture in these regions is frequently near saturation in the model).

Recent observational evidence has lent support to the

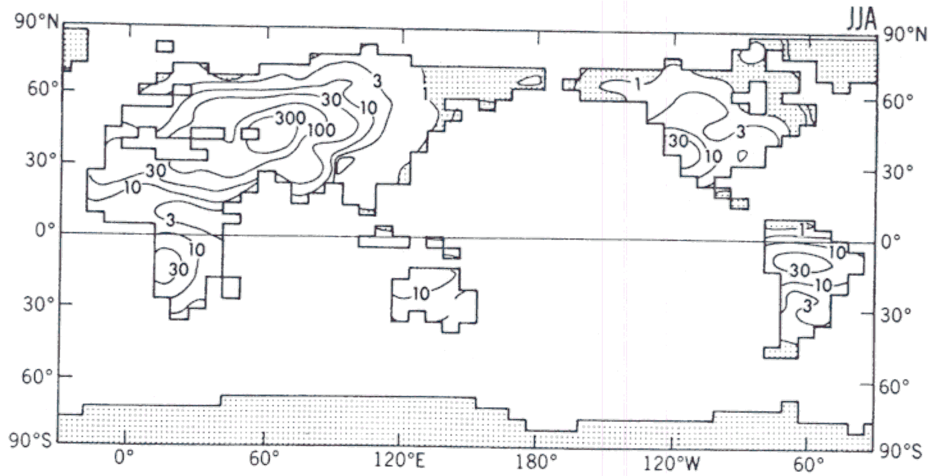

Fig. 4. Ratio of potential evaporation to precipitation at each grid point for the months of June–August.[3]

above analysis of the mechanisms controlling the time scales of soil wetness variations. Vinnikov and Yeserkepova[22] have analyzed observed soil wetness from several decades over the former Soviet Union. They state that 'Empirical datasets are found to confirm the theoretical conclusions of Delworth and Manabe (1988) that the spectrum of the temporal variations of soil moisture

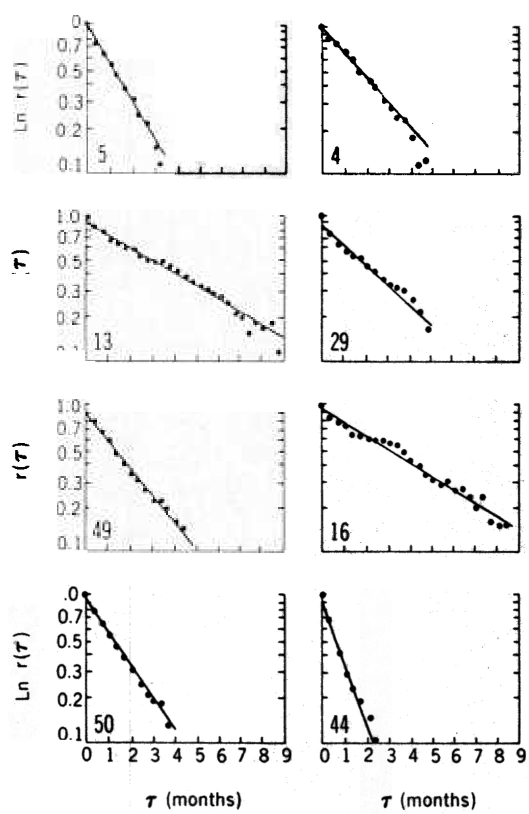

Fig. 5. Autocorrelations ($r(\tau)$) of observed soil wetness in a 1-m deep layer as a function of lag (in months).[22] Each panel represents results from a different station in the former Soviet Union. Stations 4 and 5 are in a forest zone, 29 is in the forest–steppe zone, 13 and 16 are in the steppe zone, 49 is in the semiarid zone, and 44 and 50 are in the desert. For details of the station locations and measurements, see Ref. 22.

corresponds to a first order Markov process with the decay time scale being equal to the ratio of field capacity to potential evaporation.' Shown in Fig. 5 (adapted from Fig. 6 of Ref. 22) are the computed autocorrelation functions for soil wetness measured at a number of stations in the former Soviet Union. The linear fits suggest that the temporal behavior of observed soil wetness corresponds quite well to a first-order Markov process. Additional observational results (not reproduced here) have demonstrated that the decay time scales of measured soil wetness for these regions are indeed influenced by the ratio of field capacity to potential evaporation. This observational evidence is an important indicator that the essential physical mechanisms controlling soil moisture variability are contained in eqn (1).

4 IMPACT OF SOIL MOISTURE VARIATIONS ON THE ATMOSPHERE

4.1 Relative humidity

Soil moisture influences the near-surface atmospheric temperature and moisture content by affecting the surface fluxes of latent and sensible heat. Differences in the variability of near-surface relative humidity and temperature between the two experiments will be examined in order to assess the impact on atmospheric variability of interactions between soil moisture and the atmosphere. These interactions are present only in experiment SMI, since soil moisture is computed interactively in SMI but prescribed in SMP.

We concentrate on relative humidity and temperature because these fields are strongly influenced by the surface heat fluxes, and hence by soil wetness. The term 'near-surface' refers to the lowest finite-difference level of the model, which is approximately 85 m above the surface of the earth.

Relative humidity is strongly affected by soil moisture

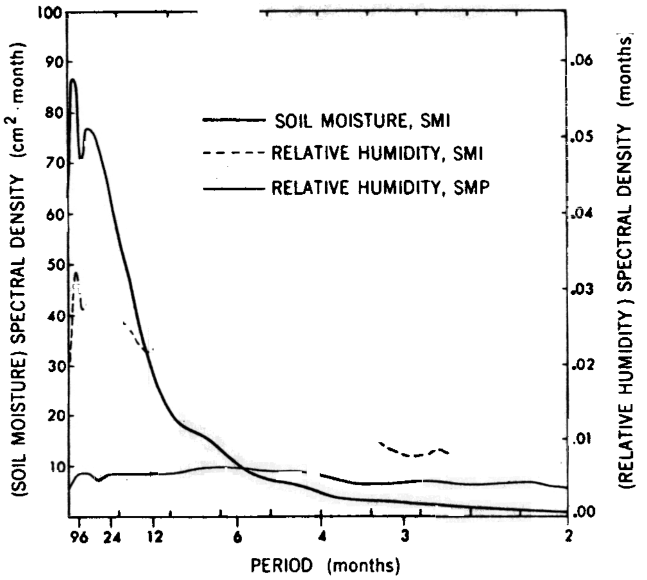

Fig. 6. Spectra of soil moisture and relative humidity.[4] These spectra were areally averaged over the region of North America between 36°N and 54°N, and between 79°W and 116°W.

variations since relative humidity is influenced by both atmospheric temperature and moisture (and therefore by the sensible and latent heat fluxes). An index of relative humidity is defined at each grid point 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.

In order to assess the impact on near-surface relative humidity of interactions between the soil layer and the atmosphere, the spectra of relative humidity from both experiments are shown in Fig. 6. These spectra are averaged over a large region of North America (defined in the caption to Fig. 6). The spectrum of soil moisture from this region is included for reference. The effect of soil moisture variations on the spectrum of near-surface

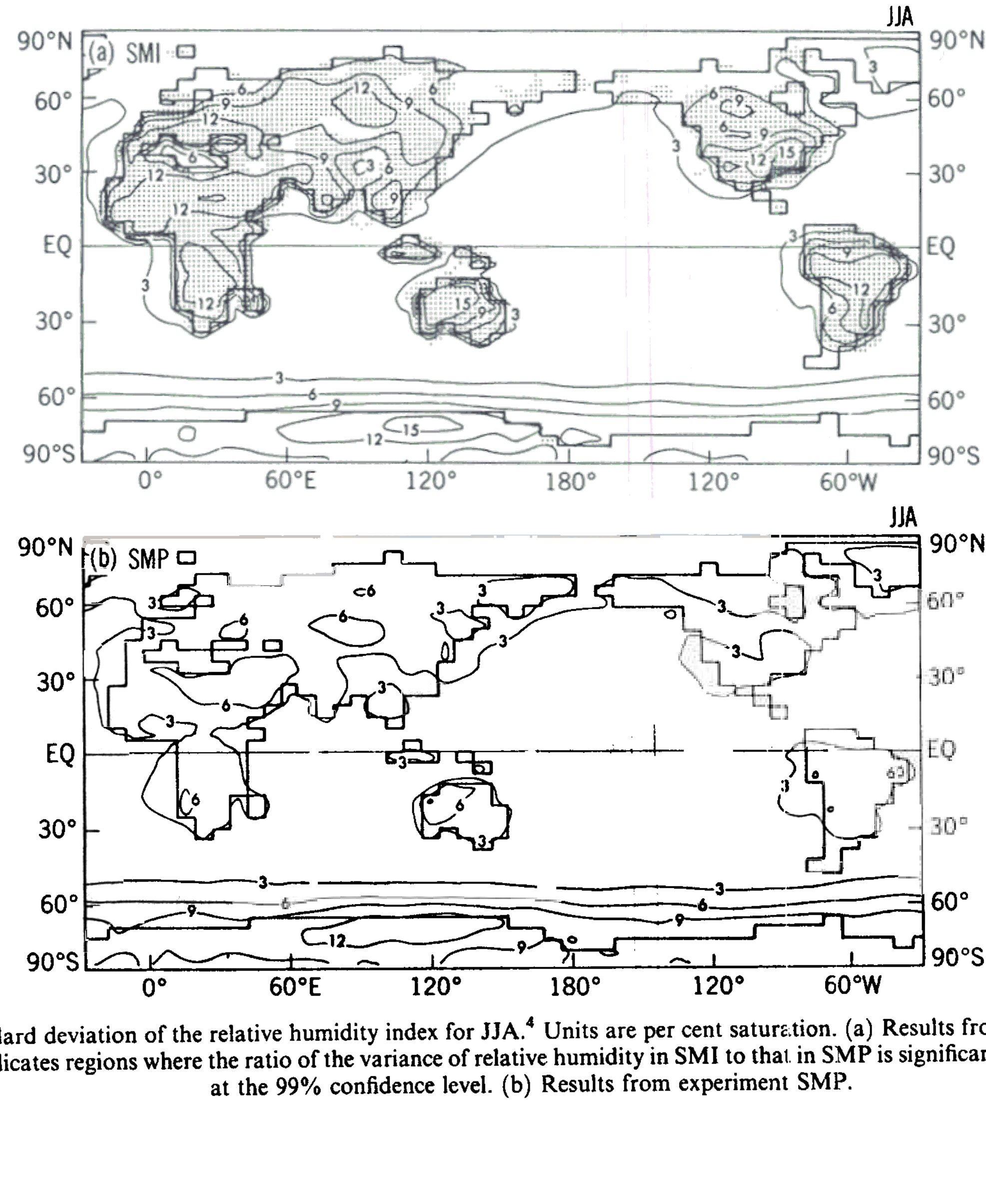

Fig. 7. Standard deviation of the relative humidity index for JJA.[4] Units are per cent saturation. (a) Results from experiment SMI. Stippling indicates regions where the ratio of the variance of relative humidity in SMI to that in SMP is significantly greater than one at the 99% confidence level. (b) Results from experiment SMP.

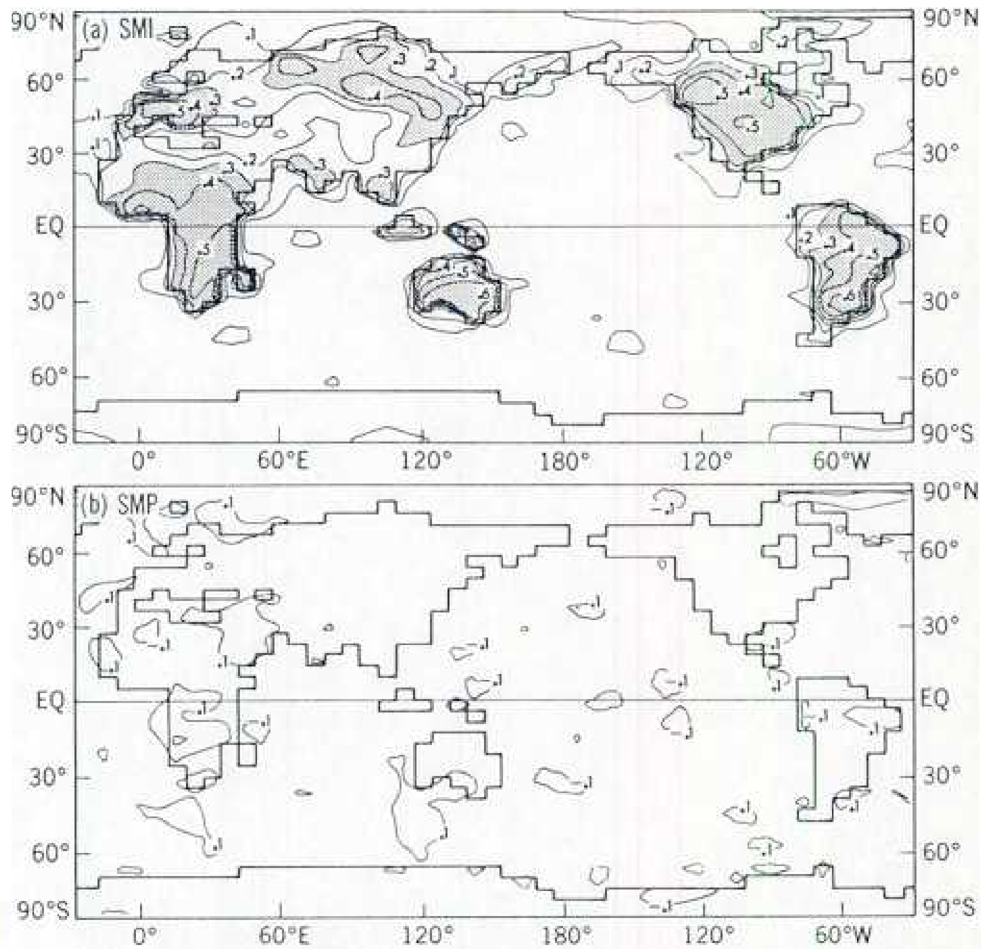

Fig. 8. Lag-one autocorrelation values for the time series of the index of near-surface relative humidity using JJA data.[4] (a) Results from SMI; values greater than 0·3 are stippled. Statistical significance is the same as in Fig. 2(a). (b) Results from SMP. Coefficients greater than 0·22 are significantly different from zero at the 95% level (the statistical significance criteria are different in experiment SMP relative to experiment SMI because of the shorter integration length for experiment SMP).

relative humidity is dramatic. In the first experiment, with interactive soil moisture, the spectrum of relative humidity has substantial variance at seasonal and longer time scales, whereas in the second experiment the spectrum of relative humidity is close to white. Variations of relative humidity in the second experiment have very short time scales. It is apparent that the long time scales and variability inherent in soil moisture fluctuations serve to lengthen the time scales of near-surface relative humidity anomalies, and to increase the variance of relative humidity (equal to the area beneath the spectrum).

The difference in the variability of relative humidity between the two experiments can be seen directly in Fig. 7, in which the standard deviations of near-surface relative humidity, computed using data from the months of June, July and August, are plotted for both experiments. The variability is substantially larger in SMI than in SMP over most continental regions, demonstrating the effect of interactions between the soil layer and the atmosphere.

The effect of interactive soil wetness on the time scales of the variations of relative humidity can be seen clearly in Fig. 8, which shows the lag-one autocorrelations of relative humidity computed using data from June, July and August for both experiments SMI and SMP. 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 seasonal variations of the persistence of relative humidity. The autocorrelations for DJF are plotted in Fig. 9 for SMI. Over the middle and high latitudes of the northern hemisphere, persistence of relative humidity is near zero, in sharp contrast to JJA

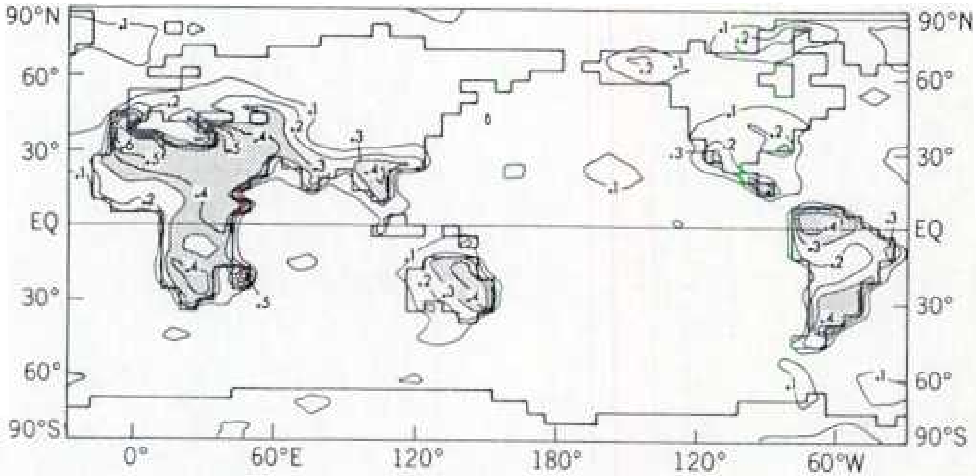

Fig. 9. Lag-one autocorrelation values for the time series of the index of near-surface relative humidity using DJF data from experiment SMI.[4] Values greater than 0·3 are stippled. Statistical significance is the same as in Fig. 2(a).

(Fig. 8(a)). This lack of month-to-month persistence in winter occurs despite the existence of interseasonal and interannual time scales of variations of relative humidity anomalies, as shown in Fig. 6. This issue will be discussed in the next section. Persistence is also smaller in DJF for the southern hemisphere extratropics.

The above-described differences in the variability of near-surface relative humidity between the two experiments demonstrate the profound effect that variations in the state of the land surface can have on atmospheric variability. The next section discusses the mechanism whereby soil moisture flucturations influence the atmosphere.

4.2 Mechanism

The state of the land surface is communicated to the atmosphere through the surface heat fluxes. Therefore, variability in these fluxes is of considerable importance to atmospheric variability. In the model, the net radiation at the surface is balanced by the sum of the latent and sensible heat fluxes, with no heat storage in the soil. Referring to eqn (2), the latent heat flux is determined as

$$\text{Latent heat flux} = LE_p f(w(t)/w_F) \qquad (9)$$

where L is the latent heat of evaporation, and E_p and $f(w(t)/w_F)$ are as defined in Section 2. Fluctuations in soil wetness alter the latent heat flux, and therefore the partitioning of the net heat flux from the surface into sensible and latent heat flux components. A positive anomaly of soil wetness will tend to increase evaporation and the latent heat flux, thereby increasing the specific humidity of the near-surface atmosphere (as well as the relative humidity for a fixed temperature). Additionally, since the sum of the latent and sensible heat fluxes is constrained to balance the net radiation at the surface, an increase of the latent heat flux is accompanied by a decrease of the sensible heat flux. The decreased sensible heat flux decreases the near-surface air temperature, thereby lowering the saturation vapor pressure and increasing the relative humidity. The changes in these two fluxes, created by a positive soil wetness anomaly, act in concert to increase relative humidity. A negative soil wetness anomaly has the opposite effect, with both the sensible and latent heat fluxes acting in concert to decrease relative humidity. Relative humidity is strongly affected by soil wetness anomalies because the changes in both the sensible and latent heat fluxes induced by a soil wetness anomaly have similar effects on relative humidity. Soil wetness anomalies can therefore influence the near-surface atmosphere by altering the Bowen ratio (ratio of sensible heat flux to latent heat flux).

The framework of this surface heat balance and the perturbations to the surface heat fluxes can be used to explain the effect of soil moisture variations on atmospheric variability. From eqn (9), fluctuations in soil moisture increase the fluctuations of both the latent and sensible heat fluxes. Since soil moisture is free to vary in experiment SMI, but is prescribed in SMP, we might expect that the variability of these surface heat fluxes would be greater in SMI than SMP. Computation of the standard deviations of the latent and sensible heat fluxes from the two experiments confirms this (not shown). There is a statistically significant increase in the variability of these fluxes over most continental regions when soil wetness is computed interactively. It is the increased variability of these fluxes in experiment SMI relative to SMP—and their impact on temperature and specific humidity—which accounts for the increased total variability of relative humidity in SMI relative to SMP.

Since soil wetness anomalies are characterized by their persistence, fluctuations in soil wetness also have a large impact on the persistence of the surface fluxes, and therefore on the persistence of near-surface atmospheric variations. The lag-one autocorrelations of the sensible and latent heat fluxes from experiment SMI are shown

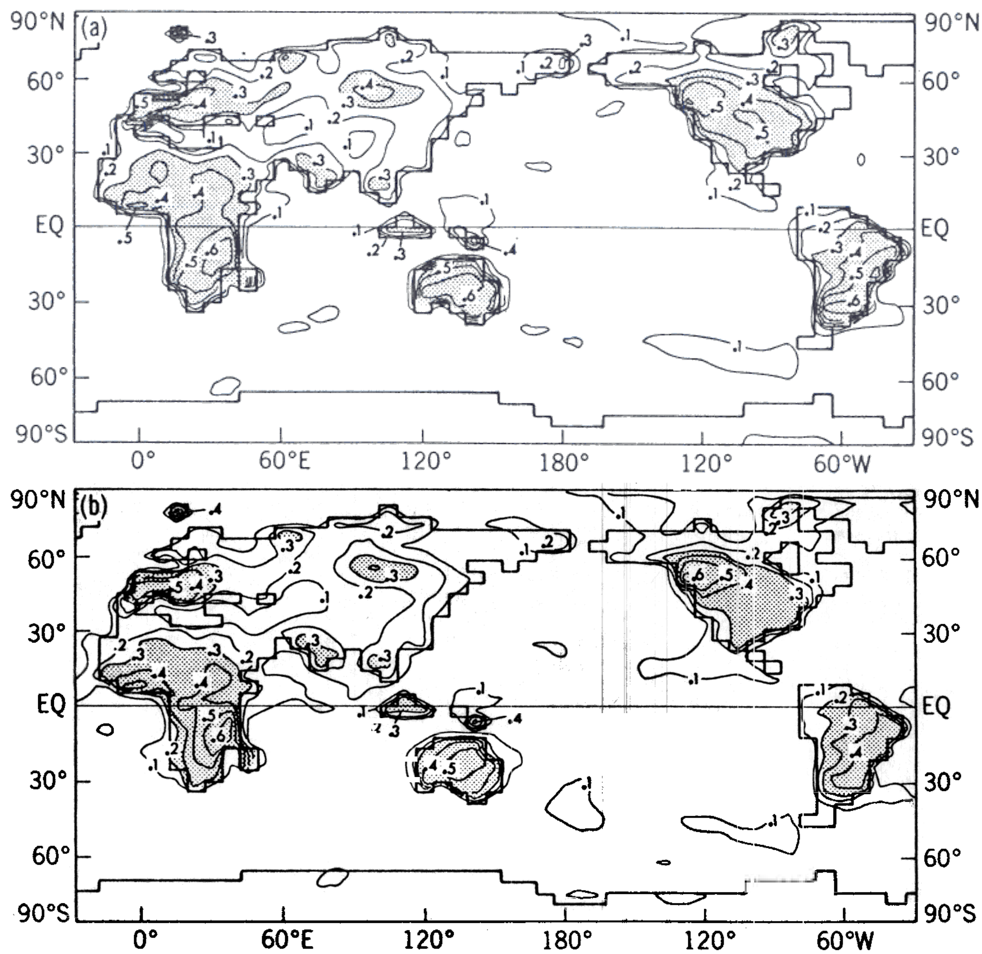

Fig. 10. Lag-one autocorrelation values for the months of JJA from SMI.[4] Statistical significance is the same as in Fig. 2(a). Values greater than 0·3 are stippled. (a) Latent heat flux at the surface; (b) sensible heat flux at the surface.

in Fig. 10. Note that the patterns bear a strong resemblance to the pattern of soil moisture persistence shown in Fig. 2(a). The lag-one autocorrelations of the surface fluxes from experiment SMP are all close to zero (not shown), indicating that the persistence of these fluxes as shown in Fig. 10 is attributable to the persistence of soil moisture fluctuations. The relatively greater persistence of these surface heat fluxes in experiment SMI relative to SMP is responsible for the longer time scales of fluctuations of atmospheric relative humidity in experiment SMI relative to SMP, as described above.

In light of this mechanism, the geographic and seasonal dependence of the increase of the time scales of relative humidity between the two experiments, as previously identified by the maps of the lag-one autocorrelation coefficients, bears further examination. The greater persistence of the sensible and latent heat fluxes in SMI relative to SMP demonstrated above leads to a greater persistence of near-surface relative humidity, as previously shown by the lag-one autocorrelations in Figs 8 and 9. The winter minimum in the persistence of relative humidity at middle and high latitudes, however, is at first perplexing, since the persistence of soil moisture is quite high during winter. One contributing factor to this is that fluctuations of relative humidity induced by cyclone waves may be larger in winter than summer. Another important factor is that the latent heat flux is proportional to potential evaporation, which is very small in winter (a result of weak insolation). Consequently, variations in the latent heat flux are quite small and do not appreciably affect atmospheric variability. Therefore, despite the substantial persistence of soil wetness anomalies in winter, there is very little persistence of relative humidity.

There is, however, substantial interseasonal and interannual variability of relative humidity as indicated in Fig. 6. While there is a lack of month-to-month persistence of relative humidity in winter, soil wetness anomalies are very persistent through the winter (since potential evaporation is quite small). These anomalies of soil wetness can persist through the winter, subsequently affecting near-surface relative humidity by perturbing the surface heat fluxes as discussed above. In this manner, the interseasonal persistence of soil moisture can induce interseasonal persistence of relative humidity, despite the very low persistence of relative humidity in winter.

4.3 Temperature and precipitation

In order to assess the impact of soil wetness variations on surface air temperature, the variances of monthly mean northern hemisphere summer (JJA) surface air temperature have been computed for the two experiments, and are shown in Fig. 11. The zonal means of the variances of surface air temperature over land are shown in Fig. 12. It is clear that the variance of the experiment with interactive soil moisture is larger than the noninteractive case in certain regions, indicating the substantial role interactive soil moisture plays in summer surface air temperature variability. The magnitude of this increase is latitudinally dependent. The zonal means of the surface air temperature variance show that the increase of variance for the interactive case is largest at low and middle latitudes of the northern (summer) hemisphere, and is very small at high latitudes. As a fraction of the variance of the noninteractive case, the changes in variance are largest at low latitudes.

There is a seasonal dependence of the increase in surface air temperature variance between the two experiments. Figure 13 is similar to Fig. 11, but computed for the northern hemisphere winter season (DJF). Possibly as a result of a sharper latitudinal temperature gradient and increased baroclinic activity, the overall magnitude of the variance of surface air temperature is larger for the winter months in both experiments. However, the relative increase in variance of the interactive experiment over the noninteractive experiment is much smaller in winter than in summer (not shown). While there are shifts in the spatial patterns of the variance between the two experiments, the areal mean northern hemisphere surface air temperature variance over land is very similar for the two experiments. Interactive soil moisture does not appear to make a substantial contribution to winter surface air temperature variability.

As discussed in the previous section, anomalies of soil wetness influence the atmosphere by perturbing the latent and sensible heat fluxes. Soil wetness is free to vary in experiment SMI, thus creating variability in the

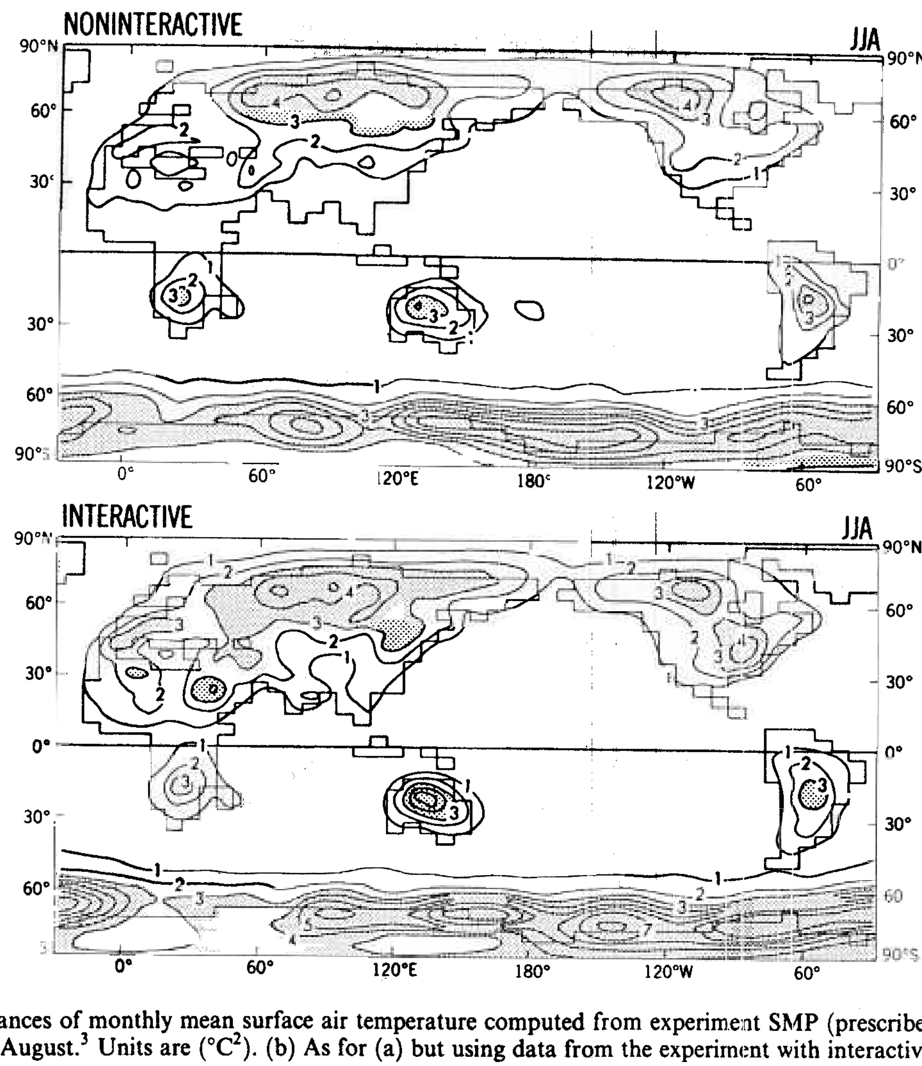

Fig. 11. (a) Variances of monthly mean surface air temperature computed from experiment SMP (prescribed soil moisture) using data from June–August.[3] Units are (°C^2). (b) As for (a) but using data from the experiment with interactive soil moisture (SMI).

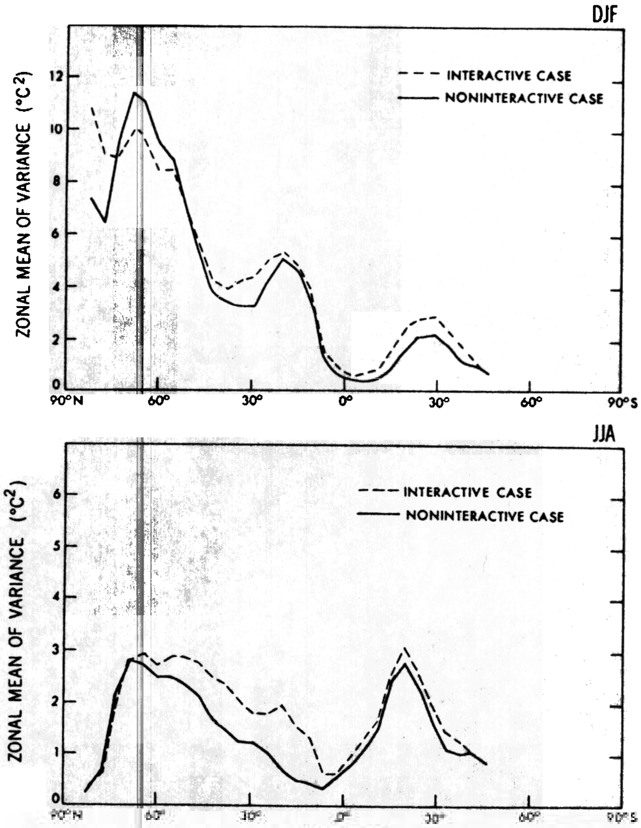

Fig. 12. Zonal means of surface air temperature variance over land.[3] Units are (°C²). (a) December–February (DJF) and (b) June–August (JJA).

Bowen ratio. In contrast, the prescription of soil wetness in experiment SMP substantially reduces the variability in the Bowen ratio. The increased variability of the Bowen ratio in experiment SMI relative to SMP—and thus in the sensible heat flux—accounts for the increased variance of surface air temperature in experiment SMI.

The latitudinal and seasonal dependence of differences in surface air temperature variance between the two experiments can be explained by variations in potential evaporation. Since the flux of latent heat depends strongly on the potential evaporation value (see eqn (9)), the magnitude of the influence of soil moisture anomalies on low level atmospheric variables also depends on the value of potential evaporation. For a fixed change in soil wetness, regions with large potential evaporation values will experience larger changes in their latent and sensible heat fluxes, and thus in surface air temperature variability, than regions with small potential evaporation values. The winter and summer mean maps of potential evaporation were shown in Fig. 3. It can be seen from a comparison of Figs 3 and 11 that the increase of surface air temperature variance in the interactive run relative to the noninteractive run is larger for regions and seasons with large potential evaporation values (tropics and the summer hemisphere extratropics). The very small changes in surface air temperature variance between the two experiments at high latitudes in summer and at middle and high latitudes in winter apparently result from the small potential evaporation values found there.

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 km. 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). However, as shown in Fig. 14, 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[11] allows us to reject at the 95% level the null hypothesis that the field of autocorrelations arose by chance. Physically, the persistent positive anomalies of soil moisture tend to increase evaporation (latent heat flux) and decrease the sensible heat flux, thereby enhancing relative humidity in the lower troposphere and making precipitation more likely.

It should be noted that the time-mean climates of the two experiments are very similar, but not identical. Differences in the climatologies occur in arid regions, and are pronounced (not shown) in the fields of potential evaporation, actual evaporation and, to a lesser degree, surface air temperature. These differences arise physically because of the non-linear dependence of potential evaporation on the saturation mixing ratio of water vapor (see eqn (3)). This may be seen from the following: as discussed above, negative (positive) anomalies of soil wetness result in an increase (decrease) of ground surface temperature. Because of the non-linearity of the Clausius-Clapeyron equation, the increase in the saturation mixing ratio (and thus potential evaporation) resulting from a positive temperature anomaly will be larger in magnitude than the decrease in saturation mixing ratio (and thus potential evaporation) resulting from a negative temperature anomaly of equal magnitude. Consequently, there is a tendency for the time-mean potential evaporation to be greater than the potential evaporation corresponding to the time-mean soil wetness. This is the essential difference between the climatologies of SMI and SMP. In SMP, the potential evaporation is computed using ground surface temperatures consistent with the prescribed soil wetness. In SMI, however the soil wetness is free to vary, resulting in a much larger variability of ground surface temperature. Therefore, the time-mean

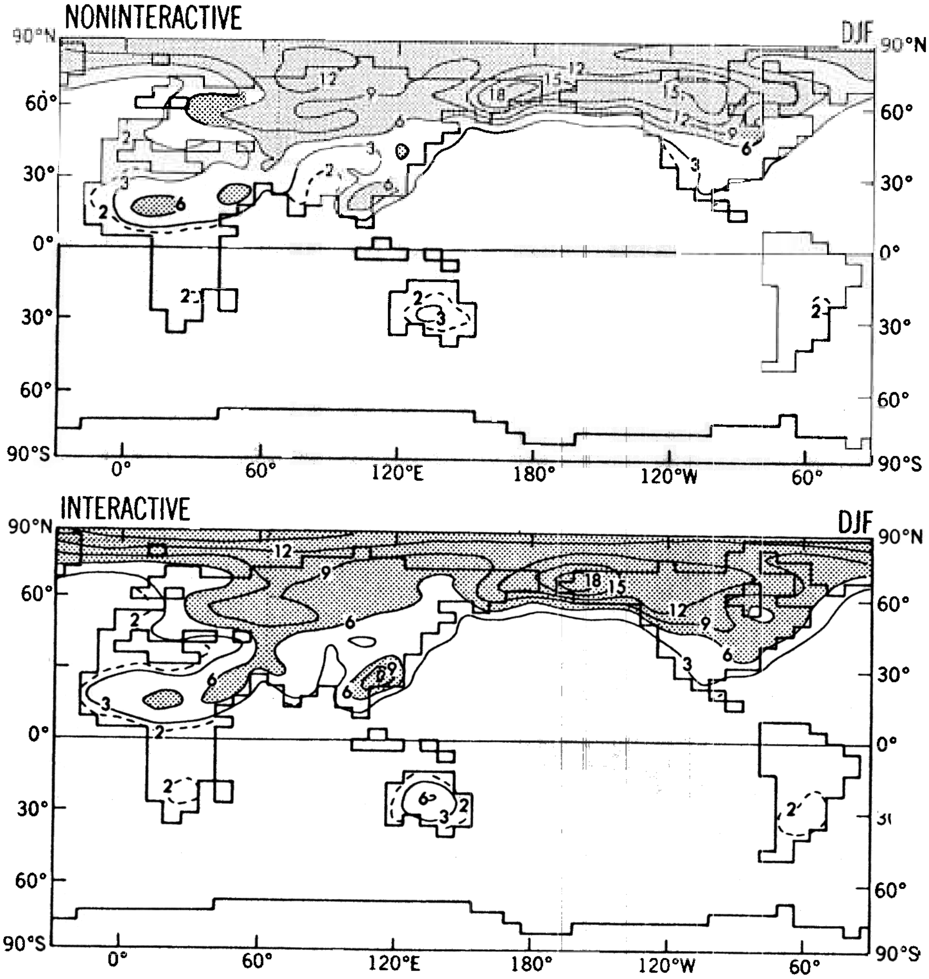

Fig. 13. (a) Variance of monthly mean surface air temperature computed from experiment SMP (prescribed soil moisture) using data from December–February.[3] Units are (°C^2). (b) As for (a) but computed from experiment SMI (the experiment with interactive soil moisture).

potential evaporation in SMI will tend to be larger than in SMP. This effect is most pronounced in hot climates (such as central Asia in summer, and northern Africa), since the non-linearity of the Clausius-Clapeyron equation increases with temperature. The larger potential evaporation values in arid regions (with large insolation) for SMI relative to SMP are consistent with the somewhat higher evaporation and temperatures in SMI relative to SMP for those regions. In relatively temperate climates, the time-mean climatologies of SMI and SMP are quite similar.

This dependence of potential evaporation on soil

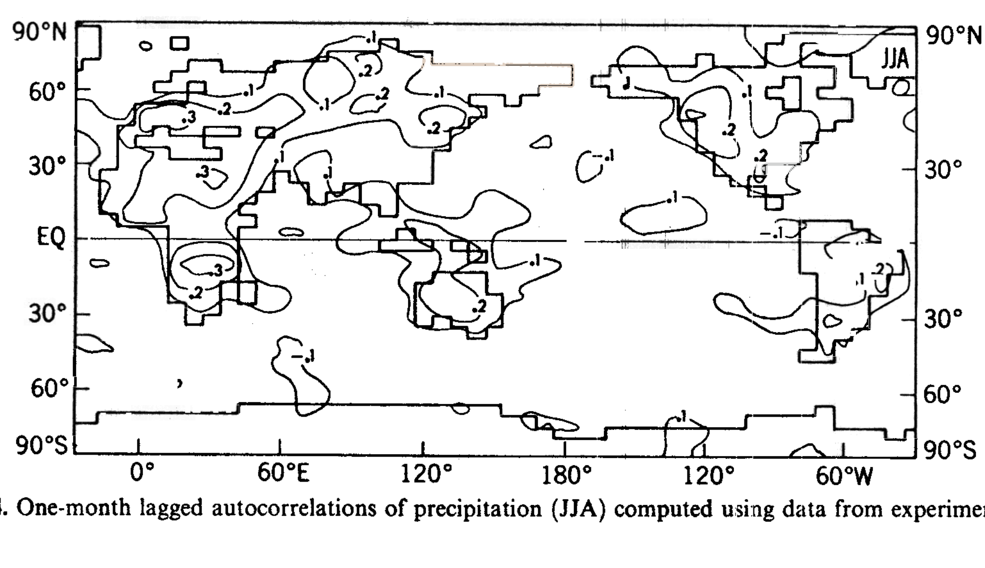

Fig. 14. One-month lagged autocorrelations of precipitation (JJA) computed using data from experiment SMI.[4]

wetness contributes to the very high potential evaporation values computed in SMI for hot, arid regions such as central Asia and northern Africa. The dry soil, combined with large insolation during summer, leads to high ground surface temperatures, further increasing the potential evaporation rate, reducing soil wetness, and increasing temperature.[15]

The differences in near-surface atmospheric variability described in this section occur, however, over large continental regions where the climatologies of the two experiments are quite similar. Thus, while the differences in atmospheric variability described above may be partially attributable to differences in the mean climatologies between the two experiments, they are primarily attributable to interactions between the soil layer and the atmosphere, present only in experiment SMI.

5 SUMMARY AND DISCUSSION

The results of this modeling study have shown in a physically meaningful manner that variations in soil wetness are of potentially profound importance for atmospheric variability. 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. In regions where the ratio of the mean potential evaporation rate ($\bar{E}_p$) to the mean precipitation rate ($\bar{P}$) is greater than one, runoff is quite small, and evaporation is sufficient to balance precipitation. Under these conditions, the temporal variability of soil moisture resembles a first-order Markov process, and the time scale of soil moisture fluctuations is determined by the ratio of the field capacity to the potential evaporation rate. 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.

Where the ratio of the mean potential evaporation rate to the mean precipitation rate is less than one, the soil is frequently saturated and runoff is frequent. Under such conditions, evaporation is at the potential rate, and anomalies of soil wetness depend principally upon the short time scale precipitation anomalies. The time scales of soil wetness in such regions are quite short.

The dependence of the time scales of soil wetness variations on potential evaporation and field capacity is supported by recent observational evidence,[22] where it is stated that, 'Empirical datasets are found to confirm the theoretical conclusions of Delworth and Manabe (1988) that the spectrum of the temporal variations of soil moisture corresponds to a first order Markov process with the decay time scale being equal to the ratio of field capacity to potential evaporation.' This observational evidence is an important indicator that this numerical model contains the essential physical mechanisms controlling soil moisture variability. The dependence of the time scales of soil wetness on potential evaporation suggests that these time scales may be altered as the climate changes due to increasing concentrations of atmospheric carbon dioxide. This issue remains a topic for future research.

The persistence and variability of soil wetness anomalies has a substantial impact on the variability of the lower model troposphere. Anomalies of soil wetness affect the atmosphere through their influence on the surface heat balance. The net radiation at the surface is balanced by outgoing fluxes of sensible and latent heat. The partitioning between the latent and sensible heat flux components depends on the soil wetness. A positive anomaly of soil wetness increases the latent heat flux, thereby decreasing the sensible heat flux (since the sum of the latent and sensible heat fluxes must balance the net radiation at the surface). The net effect of a positive soil wetness anomaly is to moisten and cool the near-surface atmosphere, thereby creating a positive relative humidity anomaly (the opposite effect occurs for negative anomalies of soil wetness).

The seasonal and geographical dependence of the interactions between the soil layer and the atmosphere is strongly influenced 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 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 soil moisture anomalies to have a substantial effect on the variability of the lower atmosphere.

Spectral analyses have demonstrated that model relative humidity near the ground surface has substantial variability on interseasonal to interannual time scales. This occurs despite a pronounced lack of persistence of near-surface relative humidity during winter at middle and high latitudes. The interseasonal to interannual persistence of near-surface relative humidity is principally due to the influence of soil wetness, which has large persistence through the winter, and substantial variability on interseasonal to interannual time scales. Persistent soil wetness anomalies perturb the surface heat fluxes, thereby enhancing the persistence of relative humidity anomalies.

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 summer, 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 which 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[19] 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, thereby precluding any possible interactions between soil wetness and cloudiness. 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 is not in complete agreement with available observational evidence.

An additional model bias is that the variance of surface air temperature over continental regions tends to be overestimated, particularly in the tropics and summer hemisphere extratropics. This bias may be related to the formulation of potential evaporation used in the model. Recent work[15] has shown that the formulation of potential evaporation in this model differs from the version proposed by Budyko.[1] As demonstrated by Milly,[15] Budyko computed the value of T_* (ground surface temperature) from a surface energy balance assuming *saturated* soil conditions. In eqn (2), the model uses the value of T_* derived from a surface energy balance using the computed soil wetness value, which may be less than saturation. Potential evaporation in this model is therefore not independent of soil moisture. The adoption of Budyko's formulation might have reduced the variability of surface air temperature, thereby making it more realistic.

The formulation of land-surface processes in the present model is highly idealized, and this should also be kept in mind. For example, the field capacity is assumed to be the same everywhere. Despite these idealizations, the present model appears to contain the most fundamental processes controlling the heat and moisture budget at the continental surface, and successfully reproduces the observed dependence of the variability of soil wetness on potential evaporation and field capacity.[22] Indeed, the authors 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. Many parameterizations of surface processes recently developed have a large number of parameters specified from observations, which makes the task of assessing the dependence of variability characteristics on physical processes very difficult. The impact of land surface parameterizations on variability—rather than on the mean—is not an issue that is typically discussed with regard to new land surface formulations, but it is an issue of substantial importance from the perspective of climate variability and climate change. In particular, the importance of vegetation on the variability of land-surface processes and the atmosphere needs to be explored.

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