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The time scale of land surface hydrology in response to initial soil moisture anomalies: a case study

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ABSTRACT

The sensitivity of a general circulation model to changes in the initial soil moisture distribution is illustrated through the detailed analysis of a case study. Two summer climate simulations are compared where initial soil moisture was obtained from a climatological distribution or from a previous integration of the model. In a third experiment, the climatological distribution of soil moisture is used as a fixed boundary condition throughout the integration. At low latitudes, the initial difference in the zonal mean soil moisture causes a significant difference in the soil temperature or land-surface evapotranspiration, but persists only up to 15 days. At high latitudes, the initial difference is more persistent, lasting through the end of the integration (50 days). In particular, large soil moisture differences persist over Siberia or North America during the whole integration and cause persistent temperature differences of a few degrees in the same regions. At high or middle latitudes, the soil-moisture values at the end of spring may therefore significantly influence the climate of the subsequent summer. Potential evapotranspiration determines to a large extent the characteristic time scale of the ground hydrology.

1. Introduction

The sensitivity of the atmospheric general circulation and hydrological cycle to soil moisture conditions has been demonstrated by a large number of numerical studies. These studies compared the climatic state resulting from the prescription, on a global or continental scale, of sustained dry or wet conditions (Manabe, 1975; Walker and Rowntree, 1977; Kurbatkin et al., 1979; Shukla and Mintz, 1982; Rowntree and Bolton, 1983). Although the experiments above were very different in nature, they all illustrate the importance of land surface processes in the maintenance of the present climate. The surface temperature can vary by as much as 20°C, from dry to wet surface conditions.

The climatological effect of sustained soil moisture anomalies appears thus to be rather well established. It is still an open question, however, whether the atmospheric effects of such anomalies permit improvements in long-range forecasting (LRF). Such improvements will depend on the time scale of the persistence of anomalous land surface conditions in the presence of interactive soil hydrology. Walker and Rowntree (1977) have shown extended persistence of soil-moisture anomalies over the Sahel, which were confirmed by subsequent experiments (Cunnington and Rowntree, 1985). In the experiments of Rowntree and associates, the climate system seems to be intransitive: the equilibrium reached starting from initially dry conditions is different from the one which is reached starting from initially wet conditions.

The role of latitudinal variations in the time scale of land surface hydrology has been emphasized by Yeh et al. (1984): they found that soil moisture anomalies persist longer at high latitudes, due to the diminution of solar heat flux. In the study by Yeh et al. (1984), although relatively large initial soil moisture differences were prescribed, the model behaved in a transitive manner and the same equilibrium was

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reached in all the experiments. This was also the case for the study of Rowntree and Bolton (1983) concerning the climatic impact of land surface hydrology over Europe. Rind (1982) made a similar study over the United States with a weaker modification of the initial soil moisture; he showed the impact of a change in soil moisture content at the end of spring on the climate of the subsequent summer.

Serafini and Sud (1987) derived a typical time scale for drought inception in the absence of rain from a simple and local water storage model. If there is no precipitation, the evapotranspiration rate gives the rate at which the soil is drying. By assuming that evapotranspiration is the product of a moisture availability factor and potential evapotranspiration, they get a drying time scale in the absence of rain which is inversely proportional to the potential evapotranspiration and which, for present soil moisture conditions, varies from a few days at low latitudes in semi-arid regions to several months at high latitudes. One should note that their moisture availability function was taken from Mintz and Serafini (1981) and slightly differs from the one used in our GCM: this is detailed in Subsection 2.2. The study by Serafini and Sud (1987) does not include all the feedback effects which are taken into account in a GCM (recirculation of evaporated water in the form of rain) and is limited to a rainless situation, but it gives an analytical formula that can provide a simple explanation for our GCM results. The same issue, but for variability at longer times scales, has also been addressed in a different form by Delworth and Manabe (1988). They do not restrict themselves to rainless situations and model the soil moisture variability which occurs in long integrations of their GCM as the response of a diffusive system to a white noise input (rain and snow melt). Their approach does not refer to any particular initial conditions and the results are more universal than those derived from usual sensitivity experiments. The time scale for the diffusive process is also inversely proportional to the potential evapotranspiration and increases with latitude, which means that the time scale for soil moisture variability is also typically larger at high latitudes.

The purpose of the present article is to illustrate the possible impact of soil moisture initial conditions on LRF through the example of a case study where the differentiation of the model response with latitude appears very clearly. The model was integrated from the European Center for Medium Range Weather Forecast (ECMWF) analysis for 11 June 1979. The experiments were carried out for 50 days, with boundary conditions fixed in time and corresponding to a mean July climatology (albedo, insolation, sea surface temperature). We have chosen two different fields to initialize the soil moisture content in our model: one is the result of a 50-day integration made with a former version of the Laboratoire de Météorologie Dynamique (LMD) General Circulation Model (GCM), the other is the climatology derived by Mintz and Serafini (1981) for the month of July. The initial differences are generally larger than the model variability as will be commented further in Subsection 2.2, but the geographic location of the major arid region of the world is generally unchanged: the impact on the general circulation will be much smaller than that obtained for example in the studies by Walker and Rowntree (1977), and the results will mainly depend on the changes in the local ground water budget and associated evapotranspiration rate. We therefore expect smaller time scales for the hydrology and we do not expect any intransitive behavior to appear, as in Walker and Rowntree (1977).

The model and the experimental design are described in Section 2. Results for evaporation, soil moisture, precipitation, and ground temperature are analyzed in Section 3. Concluding remarks follow in Section 4.

2. Model and experimental design

2.1. Model description

Various descriptions of the LMD GCM are available elsewhere (Sadourny, 1983; Sadourny and Laval, 1984). We review here only the main features relevant for the present study. For horizontal coordinates, the model uses longitude (x) and sine of latitude ($y = \sin \phi$). The grid is a uniform-area, staggered, Arakawa's C-type grid, with regular spacing in x and y. The current resolution is 64 points in longitude and 50 points from pole to pole. The mesh size at the equator is 5.6° in the East–West direction against 2.5° in the North–South direction. The vertical coordinate is the usual sigma coordinate (pressure normalized to its value at the surface). The model has eleven layers unevenly spaced in σ. To compute turbulent fluxes in the boundary layer, the model uses diffusion coefficients that depend on the wind shear and on the thermal stability. The physical parameterizations take into account a description of the hydrological cycle: precipitation occurs through supersaturation, moist adiabatic adjustment and a Kuo-type scheme, which are used sequentially. In the following experiments the cloud fractional area and cloud optical properties used to compute the solar and infrared fluxes are not computed but prescribed, following London's (1957) climatology. This means that although we have a fully consistent representation of the hydrological cycle, its radiative forcing is mainly climatic and does not interact with it (or only weakly through the variations of the atmospheric water vapor content). This approach has been chosen because the cloud-radiation feedbacks are difficult to parameterize adequately within the GCMs, and because this is unlikely to affect our main conclusions. What we neglect is mainly the fact that under rainy conditions the surface radiative flux, and therefore the amount of energy available for evapotranspiration from the ground, is modified (generally diminished) due to the cloudiness impact on both the telluric and solar radiation: we take this effect into account only through mean climatic values of cloudiness.

Ground temperature and soil moisture are prognostic variables of the model. There is a single reservoir for soil water W, and a separate reservoir for snow, which is allowed to accumulate, melt, or sublimate. But neither soil moisture nor snow is taken into account in the computation of albedo. Run-off occurs instantaneously whenever the soil moisture is larger than $W_* = 150$ mm, where W_* is the maximum amount of water which may be held by the ground at root level. Evaporation from the ground is parameterized as the product of potential evaporation, which is the evaporation from a wet soil for given meteorological conditions, and a moisture availability factor, β. The potential evapotranspiration E_p is:

$$E_p = \rho V_s C_D (q_s(T_s) - q_a), \tag{1}$$

where ρ is the density of air, V_s the surface wind speed, C_D a drag coefficient; $q_s(T_s)$ is the water vapor mixing ratio at saturation for the temperature of the ground T_s, and q_a the water vapor mixing ratio of the air near the surface.

The moisture availability function is written:

$$\beta = \min(2 \cdot W/W_*, 1). \tag{2}$$

The formulation of β takes into account the fact that evapotranspiration remains near its potential value for a large range of soil moisture values and then decreases quickly when the soil dries out. The representation of soil moisture is more adapted to vegetated areas than bare ground: the implications for the validity of our results are discussed in Section 4.

2.2. *Design of the experiment*

Three 50-day experiments have been run with the LMD GCM starting from the initial conditions of 11 June 1979 (ECMWF analysis), and July boundary conditions. They are all perfectly similar except for the initialization or specification of soil moisture. Their main characteristics are given below and summed up in Table 1.

(i) In a first control experiment (hereafter denoted by SMICLI), the initial soil moisture field is the climatological value for the month of July (Mintz and Serafini, 1981).

(ii) In a second experiment (hereafter denoted by SMIMOD), another initial soil moisture field is chosen, somewhat arbitrarily; it is the result of a 50-day integration with a previous version of the model.

(iii) In a third experiment (hereafter denoted by SMFCLI), the soil moisture climatological value for the month of July is used as a fixed boundary condition of the model.

Table 1. *A list of the three separate experiments conducted in the present study*

Experiment	Soil moisture	Initial field of soil moisture
SMICLI	prognostic	climatology
SMIMOD	prognostic	previous model experiment
SMFCLI	prescribed	climatology

The three experiments differ in the treatment of soil moisture, W. The same parameterization of the evapotranspiration has been used in all the experiments. The climatological field of soil moisture for the month of July is from Mintz and Serafini (1981).

The geographical distribution of soil moisture for the month of July is shown in Fig. 1. The global distribution of the initial soil moisture difference between experiments SMICLI and SMIMOD will be shown in Section 3 (see Fig. 6a). We note that in both distributions the arid regions of the world have about the same amount of soil moisture (with the exception of the South-West of United States), and that the major differences appear over humid areas, for which the initial values used in SMIMOD are consistently smaller than those used in SMICLI.

Mintz and Serafini (1981) (hereafter referred to as MS) obtained the normal global field of soil moisture for the month of July on a 4° × 5° grid by forcing a simple water-budget model with observed mean precipitation (Jaeger, 1976) throughout an indefinitely repeated seasonal cycle. Although a direct validation of these results is not possible, the fact that the authors obtained a good quantitative agreement between the computed annual run-off, over each of the world's largest river basins, and the measured annual river flow constitutes an indirect validation of their work. Another climatology was derived in the last years by Willmott et al. (1985): the main differences with the study by MS come from the different input data, but the method is very similar, except for the treatment of snow. There are various reasons why the distribution by MS may differ from the model climatology. First, of course, the precipitation and potential evapotranspiration simulated by the model differ from the monthly climatologies of precipitation and potential evapotranspiration used as input by MS. But the parameterizations for the water budget equation are also different. In particular, the functions of soil moisture which are used to specify the moisture availability factor β in the present study (see Subsection 2.1) and MS are different. Both functions are displayed in Fig. 2.

Coming to the interpretation of the results, we may consider two parts in the model variability. One is due to its transient behavior. The other correspond to the variability of the mean climatic state once the model has reached equilibrium. This latter contribution has been studied using the LMD GCM from a series of 80-day integrations with perpetual July boundary conditions and insolation. All of them start from the same initial state of 11 June 1979, but with a small random perturbation on the mid-troposphere zonal wind. We have not included these results in

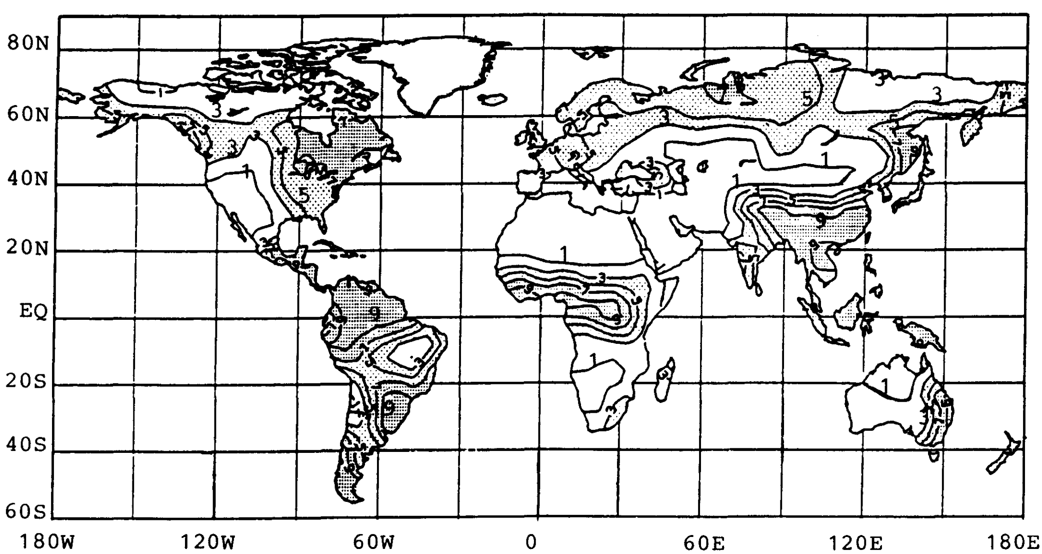

Fig. 1. Climatological distribution of available soil moisture for the month of July, expressed as the soil moisture ratio (in tenths). The maximum available soil moisture is 150 mm. From Mintz and Serafini (1981).

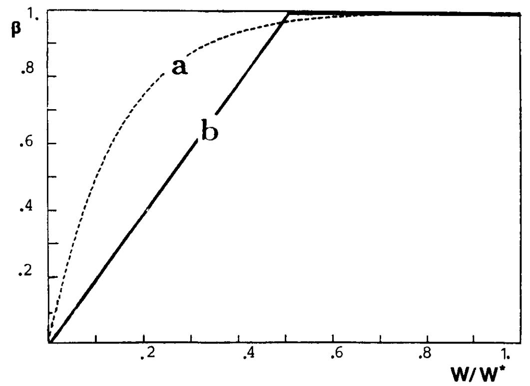

Fig. 2. Soil moisture availability function (β) used: (a) in the derivation of the soil moisture climatology produced by Mintz and Serafini (1981) and in Serafini and Sud (1987); (b) in the present study.

detail in the present paper, which focusses on a small number of unambiguous features concerning the transient behavior of the model. We can note however that the soil moisture standard deviation for 10 independent monthly means is almost everywhere less than 10 mm and can reach higher values only in some very specific locations in the Eastern Siberia, the Eastern North America and the equatorial Africa. The ground temperature variability is less than 1.2°C. The initial differences whose maintenance we study are much larger.

But the transient behavior itself is the consequence of a variety of causes: there is first an initial spin-up period which results from the inadequacies between the (ECMWF) initial state and the (LMD) GCM parameterizations of the hydrological cycle and lasts about three days. There is then a climate drift of the model which is the result of the model convergence toward a mean climatological state that has systematic errors. The experiment SMFCLI therefore provides a necessary reference to assess the role of the initial soil moisture in spite of the complexity of the transient behavior, as will be shown in Section 3.

3. Results

In the following subsections, we review the most important differences that occur in the evolution of the hydrological cycle, as stimulated in our three experiments. The results are often presented as averages over three periods: days 0–2, 20–30 and 40–50. The first interval represents the immediate atmospheric response to the given initial data. After about 20 days, purely atmospheric transients have settled down, and the response of the combined soil/hydrology/atmosphere system dominates the results.

3.1. Hydrology

To gain some insight into the nature of the time evolution of the three experiments, it is first interesting to look at the time evolution of the global averages of land evaporation and precipitation. In Tables 2 and 3, we show respectively the mean evaporation and precipitation rate over

Table 2. *Mean global evaporation over the continents (mm/day) for 6 successive time intervals (days)*

Experiment	(0–2)	(3–10)	(11–20)	(21–30)	(31–40)	(41–50)
SMICLI	3.21	3.11	3.04	2.68	2.84	2.75
SMIMOD	2.67	2.84	2.74	2.64	2.64	2.38
SMFCLI	3.34	3.41	3.44	3.44	3.44	3.46

Table 3. *Mean global precipitation over the continents (mm/day) for 6 successive time intervals (days)*

Experiment	(0–2)	(3–10)	(11–20)	(21–30)	(31–40)	(41–50)
SMICLI	4.45	4.34	3.77	3.60	3.35	3.78
SMIMOD	4.08	4.09	3.56	3.55	3.48	3.14
SMFCLI	4.50	4.42	4.00	3.97	3.89	3.87

the continents for the three experiments and for 6 successive periods of time. At the beginning of the integration, SMICLI and SMFCLI results are significantly similar. In the case of evaporation these results then diverge progressively from each other with SMICLI approaching SMIMOD. In the case of precipitation there is no clear convergence of the experiments SMICLI and SMIMOD, this being in large part due to the variability of this field. However, it is interesting to note that there is a general decrease of continental precipitation, which is apparent even in the SMFCLI case. This is compensated by an increase in the oceanic precipitation (not shown here): the model does not sustain the initial land–sea contrast in the distribution of atmospheric water vapor. The existence of such a drift is inherent to our choice of starting from a realistic initial state to integrate the model in a forecast mode. It stresses the importance of experiment SMFCLI, which is useful to distinguish the variations related or not to changes in the moisture content, transient behaviors which are or are not related to changes in the soil moisture values from those which are not. In principle, it is not obvious that the experiment SMFCLI is sufficient to make that distinction, but it seems to be partly the case in the present study.

We have displayed in Fig. 3 the time evolution of the zonal mean evaporation rate computed over land for two latitudinal bands where the

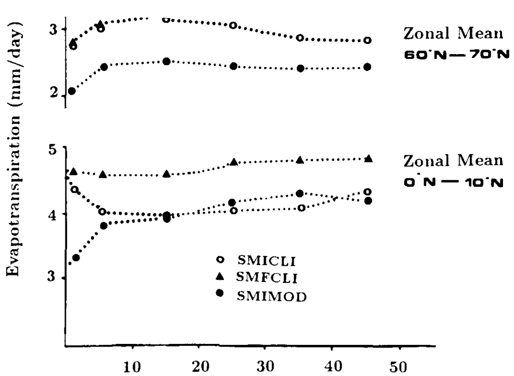

Fig. 3. Time evolution of the evaporation rate computed over land. Values are zonal means throughout the three experiments for two latitudinal bands: 0° N–10° N and 60° N–70° N. Experimental points are averages for days 0–2, 3–10, 11–20, 21–30, 31–40 and 41–50.

initial difference is very large: 0° N–10° N and 60° N–70° N. The two experiments with the same initial state, SMICLI and SMFCLI, predict similar initial values of the evapotranspiration over the two areas. The SMIMOD values are lower. Throughout the model evolution, the two experiments with predicted soil moisture values tend to converge to each other, indicating that the initial difference is forgotten. At the same time, the experiments SMICLI and SMFCLI, which have the same initial state, but either predict the soil moisture (SMICLI) or prescribe it (SMFCLI) tend to diverge from each other. This occurs very quickly at low latitudes (about 5 days), and much slower at high latitudes, where the process appears to continue beyond 50 days. The comparison of the various experiments give some indications about what is due to some climatic trend independent of soil moisture and what is due to the changing soil conditions. For example at higher latitudes the three curves are first parallel, indicating that the evolution up to day 15 or 20 is mainly independent of the soil conditions.

We now turn to the zonal distribution of these initial differences. Fig. 4 shows the difference between the mean zonal evaporation over the continents from experiments SMICLI and SMIMOD. The distributions are time means for the three successive periods: days 0–2, 20–30 and 40–50. These distributions are closely related to the corresponding zonal differences of soil moisture shown in Fig. 5. Initial soil moisture differences in the tropics induce initial differences in the evaporation which are not very persistent. We can see in Fig. 5 that the strong initial differences in soil moisture disappears very quickly. This is also true for the evaporation (Fig. 4) although somewhat less apparent because of the larger variability of this field. In contrast, at higher latitudes (we consider here only the Northern Hemisphere, because the results in the Southern Hemisphere at these latitudes involve only a small number of continental points) the soil moisture initial differences are very persistent, and consequently induce a smaller but very persistent evaporation change. This may also be seen in Fig. 6a, b which are global spatial distributions of the soil moisture difference between SMICLI and SMIMOD, averaged for days 0–2 and 40–50 respectively. Soil moisture

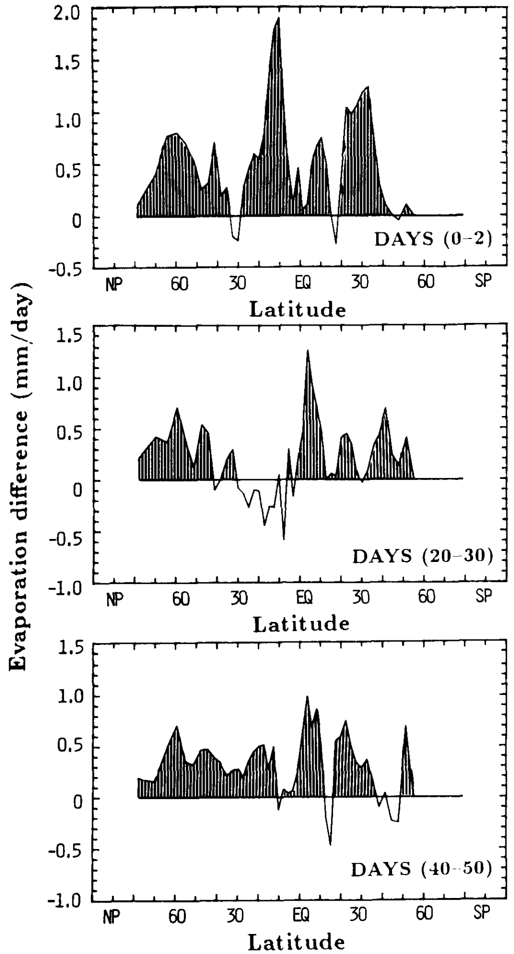

Fig. 4. Zonal mean distributions of the difference of evaporation between experiments SMICLI and SMIMOD. Mean values are computed only over land and for three periods: days 0–2, 20–30 and 40–50. Shaded areas represent positive values.

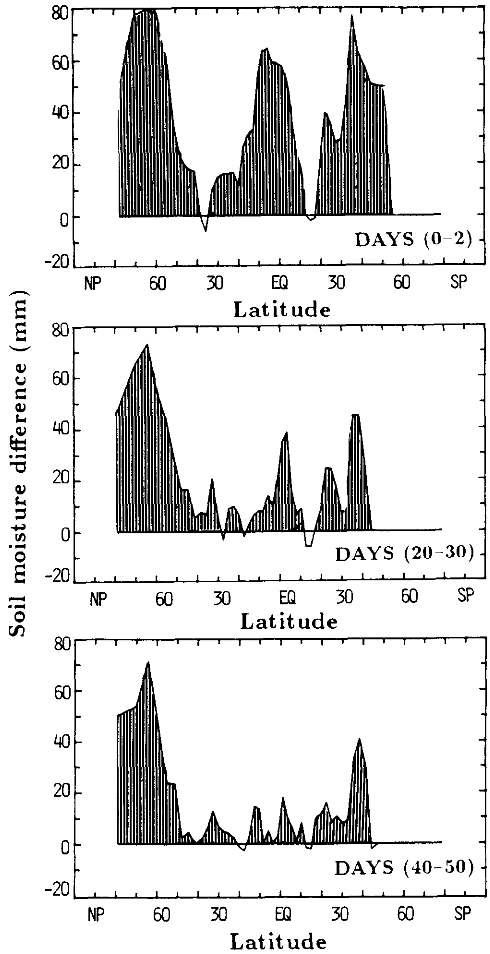

Fig. 5. Zonal mean distributions of the difference of soil moisture between experiments SMICLI and SMIMOD. Mean values are computed for three periods: days 0–2, 20–30 and 40–50. Shaded areas represent positive values.

differences north of 50° N have remained almost unchanged. This contrast between high and low latitudes is larger than in the study by Serafini and Sud (1987) for the month of July. This is due to the fact that in the present study the time scale of the hydrology also depends on the renewal of the soil water reservoir by rainfall. Precipitation is large in the zone from 0° N to 10° N and can restore quickly higher values of soil moisture in SMIMOD, at the same time as high evaporation rate can decrease quickly the soil moisture values of SMICLI, which are larger than those of the model equilibrium. The same equilibrium between precipitation, evaporation and runoff is reached more slowly at high latitudes. Although potential evapotranspiration is certainly the main parameter which determines the time scale of the hydrology, it is certainly also affected by precipitation. Delworth and Manabe (1988) in particular show the existence of two hydrological regimes, depending on the ratio of the mean evapotranspiration over the mean precipitation. If it becomes

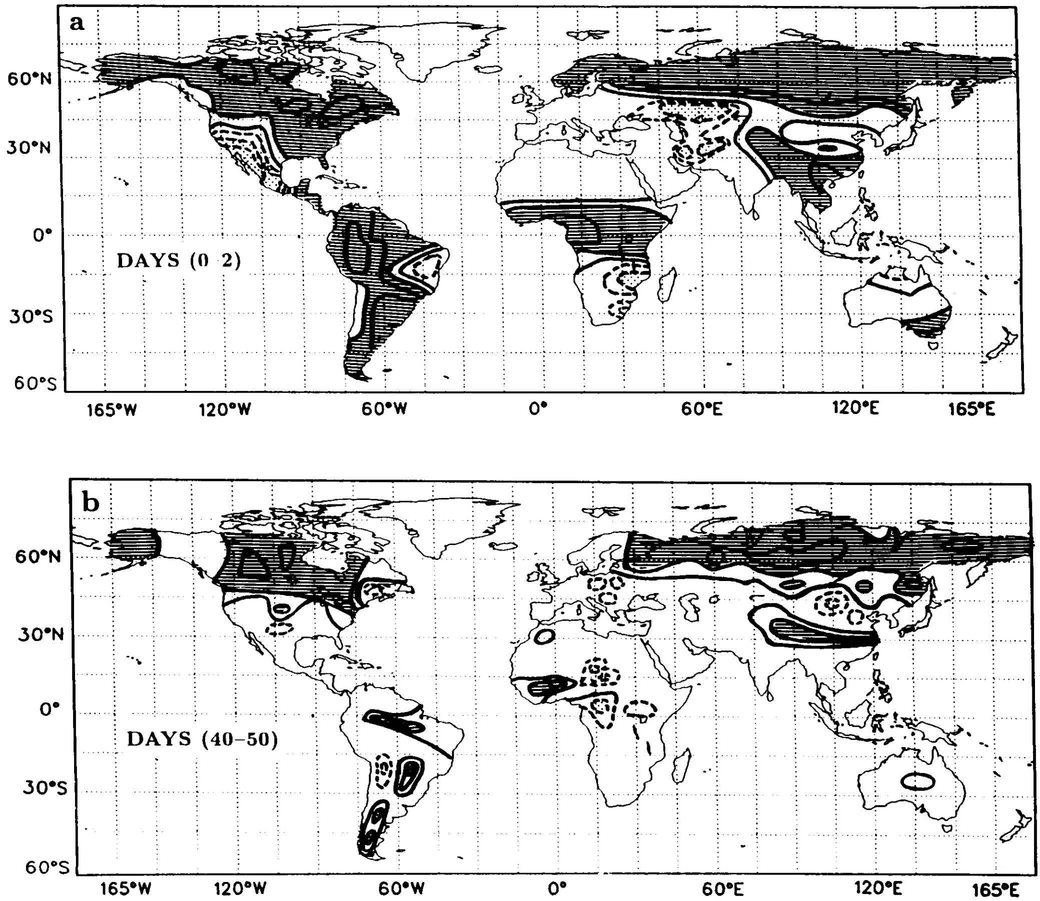

Fig. 6. Global distribution of the soil moisture difference between experiments SMICLI and SMIMOD, averaged for: (a) days 0 to 2, (b) days 40 to 50. Dark hatched outlines regions of soil moisture difference larger than 30 mm and light stipple outlines regions of soil moisture difference lower than −30 mm. Full lines are 10, 30 and 50 mm and dashed lines are −10, −30 and −50 mm, respectively.

smaller than one, frequent run-off occurs and the time scale of the hydrology is reduced. This is more likely to occur at high latitudes. In our case, however, the equilibrium of the model is drier than the climatology, and run-off is not frequent in our experiments. There is consequently no significant reduction of the time scale of the hydrology by this process.

3.2. Ground temperature

There are two main processes linking soil moisture anomalies to the general circulation of the atmosphere: first, the evaporation feeds the atmosphere with water vapor which may condense again as precipitation with a release of latent heat; second, the evaporation produces a cooling of the ground, and therefore of the surface air. This cooling is associated with a decrease of the sensible and net infrared heat flux at the surface, which, for given solar heating can compensate the increase of latent heat flux. The latter effect may be strengthened further because the enhanced evaporation will tend to decrease the net surface solar flux, due to increased cloudiness; this effect is not taken into account in the present experiment for the sake of simplicity (its actual representation in global models is a major problem).

We have noted that changes in precipitation rate are difficult to study within the framework of the present relatively short experiments. On the contrary, changes in ground temperature appear quite clearly. This is shown in Fig. 7, where we have displayed zonal mean differences in ground temperature between SMICLI and SMIMOD, for the same periods already selected in Fig. 4. Similar features to those in Fig. 4 appear, only with the reverse sign. The temperature anomaly is larger initially in the tropical belt, being as large as 5 °C, but it disappears quite rapidly. In contrast, at high latitudes in the Northern Hemisphere, initial differences are smaller, being less than 3 °C, but tend to persist for the whole 50 days. Results at high latitudes in the Southern Hemisphere are not significant as already explained.

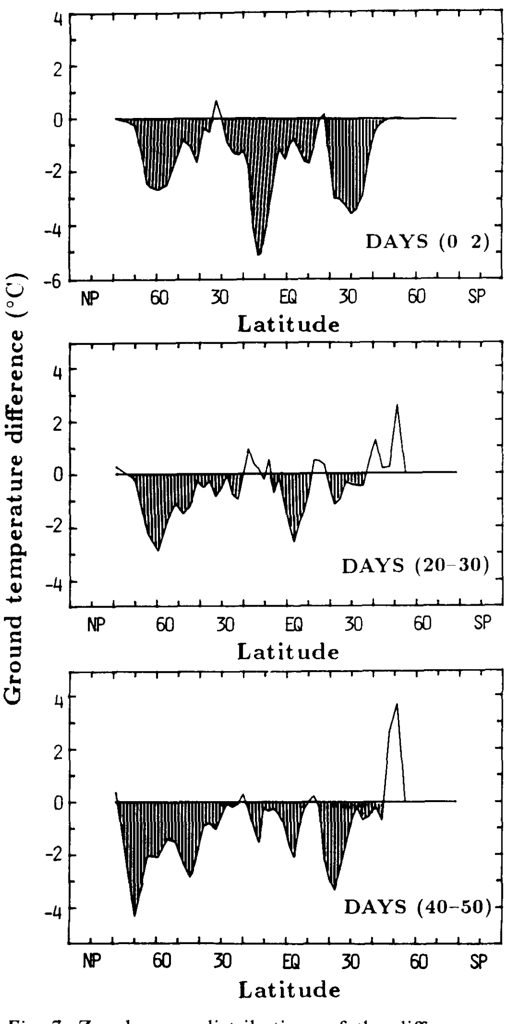

Fig. 7. Zonal mean distributions of the difference of ground temperature between experiments SMICLI and SMIMOD. Mean values are computed for three periods: days 0–2, 20–30 and 40–50. Shaded areas represent positive values.

4. Conclusions

The present results show the persistence of initial soil moisture differences and associated soil temperature differences in a series of numerical experiments using the LMD GCM. They give a rough estimate of the time scale associated with initial errors in the treatment of ground hydrology. We find a strong dependence of the obtained time scale on latitude. This was found using a simpler water budget model by Serafini and Sud (1987). In their study the time scale of the ground hydrology is defined as the time necessary to reach the ground wilting point in the absence of rain. The same dependency was also derived by Delworth and Manabe (1988) from long GCM experiments and more sophisticated methods, which take into account the role of the precipitation forcing. Our experiments therefore constitute both a confirmation and an illustration of these previous studies. At low latitudes the response time is small even in regions of large precipitation where Yeh et al. (1984) had noted an increase in the time scale of the surface hydrology due to the recycling of the evaporated water by rain. It is important to note that we have considered here initial soil moisture perturbations different from those in previous studies because they mostly respect the distribution of dry and moist areas. Yeh et al. (1984), for example, considered the effect of irrigating latitudinal bands extending over 30 degrees of latitude. Our perturbations have a lower impact on the dynamics of the model and we obtain smaller time scales. Longer lasting modifications of the tropical hydrology would require soil moisture anomalies large enough or located so as to modify the large scale tropical atmospheric circulation.

At high latitudes, the time scale obtained is large enough to influence the seasonal variations of the climatic system. It should be noted that in the present study, we have used a representation of the ground hydrology where the soil moisture is stored in a unique reservoir, contrary to the studies of Rind (1982) where two layers were taken into account, each having a field capacity which depends on the vegetation type. Hunt (1985) has shown that a two-layer description of the soil with a thin upper layer increased the time scale of the hydrology in wet regimes and decreased it in dry regimes. Our single-layer model is a fair description of vegetated areas only. In arid areas, there is evaporation mainly from a thin superficial layer. This tends to decouple the evapotranspiration from the total soil moisture content, and the present analysis is not relevant. It is also important to note that the variations of the plant stomatal resistance when the water stress increases are modelled in a crude way through our soil moisture availability factor.

Our results suggest that in such regions as Canada or Siberia the soil moisture content at the end of the spring may affect the climate of the whole subsequent summer, as hypothesized by Namias (1982) and illustrated by Rind (1982). In operational weather forecasting centers, in the absence of a soil moisture analysis, results of a previous numerical prediction are often used to initialize the following one: caution must be exercised to prevent errors in the initial state of the model from persisting over large periods and deteriorating the forecast. This also indicates that the development of observing systems able to monitor soil conditions on a monthly basis should bring a significant improvement in long-range forecasts.

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