EXPERIENCE WITH SURFACE PROCESSES AT LMD

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1. INTRODUCTION

The exchanges between the earth's surface and the atmosphere play a crucial role in the atmospheric balance. Sensitivity experiments with climate models have shown that significant changes in the atmospheric circulation and in rainfall can result from variations in the land hydrology. Since the work of Walker and Rowntree (1977) many studies [Rowntree and Bolton (1983), Yeh et al (1984) for example)] have shown that soil moisture anomalies have an impact on the evaporation and precipitation rates, because dry or wet soil conditions can persist for months.

An important effort is being made these last years to improve the representation of the soil hydrology in numerical models. The works published by Dickinson (1984) or by Sellers et al (1986) have aimed to define the exchanges between the earth's surface and the atmosphere by biophysically realistic models.

II. PARAMETRIZATION OF THE SOIL MOISTURE CONTENT

The experiments with GCMs which show the sensitivity of climate to the initial soil moisture content were conducted until 1984 with a very simple parametrization of soil hydrology : the bucket method (Manabe, 1969). The soil moisture was defined in one soil layer of 1 meter or so, and the bulk soil moisture content verified the prognostic equation

$$\frac{\partial W}{\partial t} = P - E - R \qquad (1)$$

where P is precipitation, E evaporation and R run-off. E was obtained by the product of the potential evaporation E_o and an evapotranspiration coefficient β which depends on soil moisture.

This dependence was defined by

$$\beta = \begin{cases} \dfrac{W}{0.75\ W_{MAX}} & W < 0.75\ W_{MAX} \\ 1 & W > 0.75\ W_{MAX} \end{cases} \tag{2}$$

W_{MAX} is the soil moisture at field capacity often taken as 150 mm. When W exceeds this maximum value, the excess is taken to be run-off.

Deardorff (1977), considering that soil surface moisture content varies on a time scale much shorter than the bulk moisture content, proposed to define the soil moisture by two variables : the soil surface moisture content W_G (of the upper few millimeters of soil) and the bulk soil moisture content W_B (of a layer of 50 cm). The equations that define these variables are :

$$\frac{\partial}{\partial t} W_G = -\frac{C_1(E-P)}{\rho D_1} - C_2\ \frac{W_G - W_B}{\tau} \tag{3}$$

$$\frac{\partial}{\partial t} W_B = -\frac{E-P}{\rho D_2} \tag{4}$$

with

$$C_2 = 0.9 \tag{5}$$

$$C_1 = \begin{cases} 0.5 & \dfrac{W_G}{W_{MAX}} \geq 0.75 \\ 14 - 22.5\left(\dfrac{W_G}{W_{MAX}} - 0.15\right) & 0.15 < \dfrac{W_G}{W_{MAX}} < 0.75 \\ 14 & \dfrac{W_G}{W_{MAX}} < 0.15 \end{cases} \tag{6}$$

where D_1 is the depth to which the diurnal soil moisture cycle extends, and D_2 is taken as 50 cm, and τ = 1 day.

W_B and W_G are the moisture fractions (volume of water divided by volume of soil).

The temporal variation of W_B is given by the same equation as for the bucket method, but in the temporal variation of W_G there is a first term which wets or dries the surface when evaporation is lower or higher than precipitation and the second term tends to restore W_G towards the bulk value. In this representation C_2 is constant. The three equations which define C_1 correspond to the different stages of the drying of the soil (drying with potential evaporation, drying when evaporation rate is limited by the soil moisture transport to the surface and drying by vapor transfer). The evaporation rate depends on soil surface moisture content and the potential evaporation rate E_o :

$$E = \frac{W_G}{0.75\ W_{MAX}}\ E_o \qquad W_G \leq 0.75\ W_{MAX} \tag{7}$$

$$E = E_o \qquad W_G > 0.75\ W_{MAX} \tag{8}$$

These equations simulate, after a precipitation which wets the soil surface, an evaporation equal to its potential rate.

Hunt (1985) compared the results obtained by a 1-dimensional radiative-convective model using 3 different formulations of the surface hydrology. He ran experiments started from a saturated soil at its field capacity and assumed that no precipitation occurred. He studied the time evolution of soil moisture content, latent and sensible heat fluxes, and temperature. (Figure 1). Since no precipitation was allowed, the soil moisture content and consequently the evaporation rate decreased with time. With the GFDL method [Eq. 1 and 2], the soil was dry after 50 days. With the Deardorff's method [Eq. 3 → 8] the latent heat decreased much more rapidly as the surface layer dried and after, a slower decrease was determined by transfer from the deep reservoir. With this formulation, the soil moisture content decreased much more slowly than in the GFDL method and one can ask the question of the impact of soil moisture anomalies on the simulated climate when one uses this method.

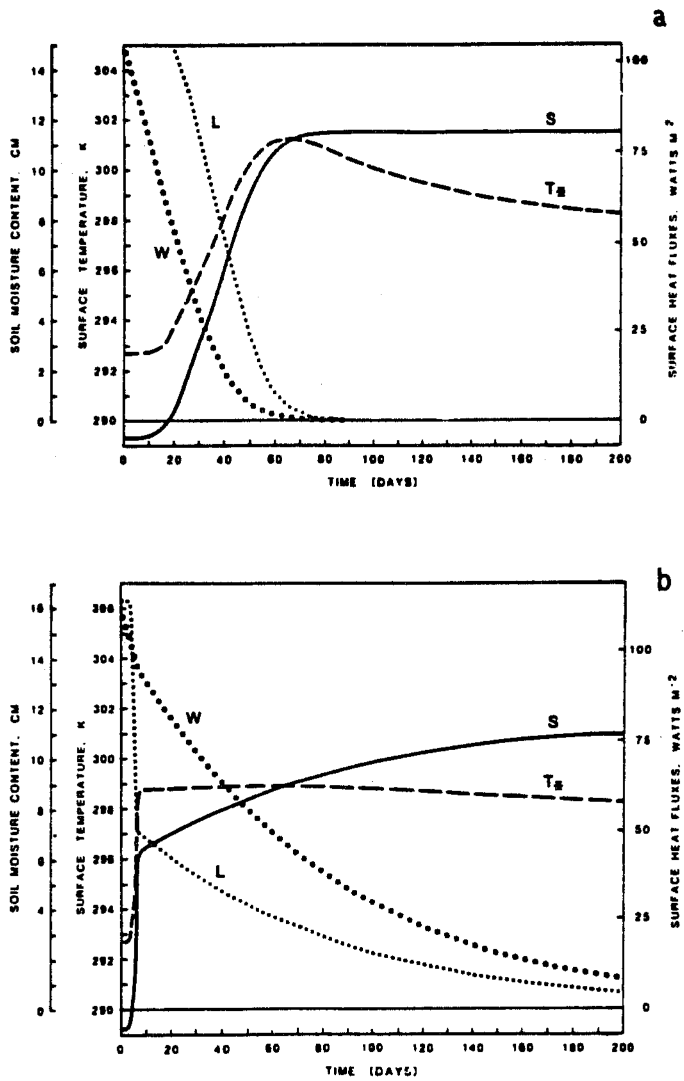


Fig. 1 : Evolution of the surface temperature, T*, soil moisture content W (left hand scales), latent heat flux, L, and sensible heat flux S (right hand scale) a) for the GFDL method and b) for the Deardorff's method. From Hunt (1985).

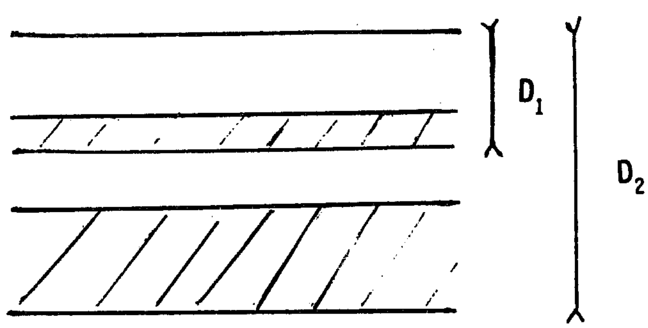


Fig. 2 : The two reservoirs of depth D_1 (variable) and D_2 (constant) defined by Choisnel's method.

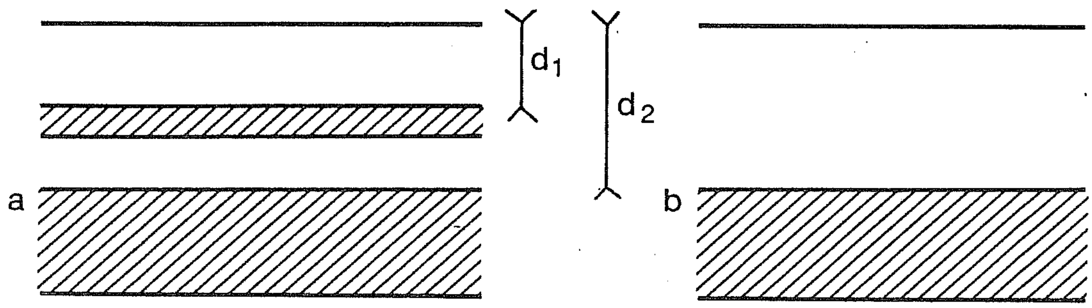


Fig. 3 : The relative humidity U_s depends on the depth of dry soil d_1 in case a) and d_2 in case b).

A different formulation was proposed by Choisnel (1977) to study the variations of the soil moisture content over a region covered by grass land. He defined two layers : a deep layer and a surface layer. The total depth (D_2) of the deep layer was constant with time but the depth of the surface layer was variable (Figure 2). When precipitation or evaporation occurs, the water supply or demand is added or removed to the surface layer, first. If the increase of soil surface content is such that all this dry soil is filled with water, then the depth of the surface layer is increased such that all the surface layer remains saturated. In this case there is no run-off. If the increase of soil moisture content is so large that all the dry soil between the two layers is saturated, then all the soil moisture is considered as in the deep reservoir and surface soil layer is equal to 0. The soil moisture content of the deep layer is then maximum and run-off can occur. If the water demand is higher than the soil surface moisture content, then the excess is removed from the deep layer.

To compute the real evaporation rate, an evapotranspiration coefficient is defined by Choisnel which is not a linear function of soil moisture but rather an exponential function given by

$$U_s = \exp - C \frac{d}{D_2} \tag{9}$$

D_2 is the depth of the deep reservoir. d is the depth of the first layer of dry soil (Fig. 3) : if soil surface moisture present, then d is taken as d_1 ; but if soil moisture content is present only in the deep reservoir, d is taken as d_2, the depth of the level of water in the deep reservoir. C is a constant that was tuned by Choisnel to obtain the best fit between real measurements of the evaporation rate and the values calculated by the model (C = 0.8).

When the deep reservoir or the surface layer is saturated, U_S is maximum. When the soil surface moisture content is 0, the function U_S depends only on the relative humidity of the deep reservoir and except for the formulation with an exponential function instead of a linear function, this dependence is similar to the bucket method. But if the soil surface moisture is present, the dependence is quite different. In other words, fresh water supply in the soil evaporates much more rapidly than soil moisture which was stored from longer time and has gone in the deep reservoir.

3. THE NUMERICAL EXPERIMENTS

We have performed several integrations of our general circulation model (GCM) with July conditions. All start with the same initial state of 11 June 1979 obtained at ECMWF from analysis of FGGE data. The mean July climate simulated by the model is defined by an average of the variables over the days 20-50 even though, for the study of temporal variations, some of the integrations have been carried out for 80 days. This GCM is a model defined on a grid constant in longitude and in sine of latitude. It has 64 points in longitude and 50 points in latitude. The parametrizations include turbulent diffusion in boundary layer, condensation processes and radiative fluxes (Laval and Picon, 1986).

The potential evaporation rate is usually defined in the GCMs by (Manabe, 1969)

$$E_o = \rho \, C_E \, V \, [q_S(T_o) - q] \tag{10}$$

where ρ is the air density, C_E is the exchange coefficient for water vapor, q is the mixing ratio of the air above the surface, $q_S(T_o)$ is the saturation mixing ratio at the surface temperature T_o.

Different formulations were proposed to define the potential or actual evaporation rate over land [Penman, 1948 ; Monteith, 1965 ; Priestley and Taylor, 1972 ; Mc Naughton, 1976, among others]. Perrier (1982) proposed to write :

$$E = \frac{P'}{P'+\gamma} (R_n - F_o) \frac{1 + \frac{\gamma}{P'+\gamma} C_E V r_c}{1 + \frac{\gamma}{P'+\gamma} C_E V r} \tag{11}$$

where R_n is the net radiation reaching the surface, F_o is the soil heat flux, $P' = \frac{\partial q_s(T)}{\partial T}$, $\gamma = \frac{C_p}{L}$ is the ratio of the specific heat of air at constant pressure to the latent heat of vaporization of water, r is the surface resistance and r_c is a climatic resistance defined by

$$r_c = \frac{P'}{P'+\gamma} \rho C_p \frac{T-T_r}{R_n - F_o} \tag{12}$$

T_r is the dew-point temperature of the air above the surface.

To define the evaporation rate over the continents, for the experiments with our GCM that are reported here, we have used essentially the equation defined by (11) where the coefficient

$$\alpha = \frac{1 + \frac{\gamma}{P'+\gamma} C_E V r_c}{1 + \frac{\gamma}{P'+\gamma} C_E V r}$$

was defined by a simple function of the relative humidity at the surface as :

$$\alpha = \begin{cases} 0.8 \dfrac{W}{W_{MAX}} + 0.6 & W > 0.5\, W_{MAX} \\ 2 \dfrac{W}{W_{MAX}} & W \leq 0.5\, W_{MAX} \end{cases} \tag{13}$$

It was shown (Laval et al, 1984) that this parametrization reduces substantially the evaporation rate over continents, compared to the bulk aerodynamic formula (10).

In all the experiments reported here, the soil moisture is time dependent. We have used the three different schemes presented in paragraph 2. The control experiment, called run C, uses the bucket method; the integrations which use the Deardorff's method over bare soil or over all the continental points are called runs D. And the integrations defined by the E. Choisnel's method are called runs E.

The experiments which use Deardorff's method for all the continental points are D_3 and D_4. But for two experiments, D_1 and D_2, we have used altogether the Deardorff's method over bare soil (extended to savannas in our model) and the bucket method over forests. These two types of soil are defined in Fig. 4.

We have defined two sets of initial conditions for soil moisture content. The first distribution is an arbitrary one which comes from a long integration of our model with January conditions. This distribution called A condition (Fig. 5) shows very low values and the continents are generally very dry. The other distribution, called MS condition , is shown in Fig. 6. It is the climatic distribution obtained by Mintz and Serafini (1984) with their simple hydrological model. Even though it may be that this distribution has some bias, it seems that this field reproduces the contrast between different latitudes and climates and is the best that we can use at the present time.

All these experiments are defined in Table 1. The experiment E' is the only one which does not use the evaporation rate as defined by equation (11), but Monteith's equation defined by

$$E_o = \rho \frac{q_S(T_o) - q}{r + \dfrac{1}{C_E V}} \tag{14}$$

and the relation

$$E = \alpha E_o$$

with α given by (13)

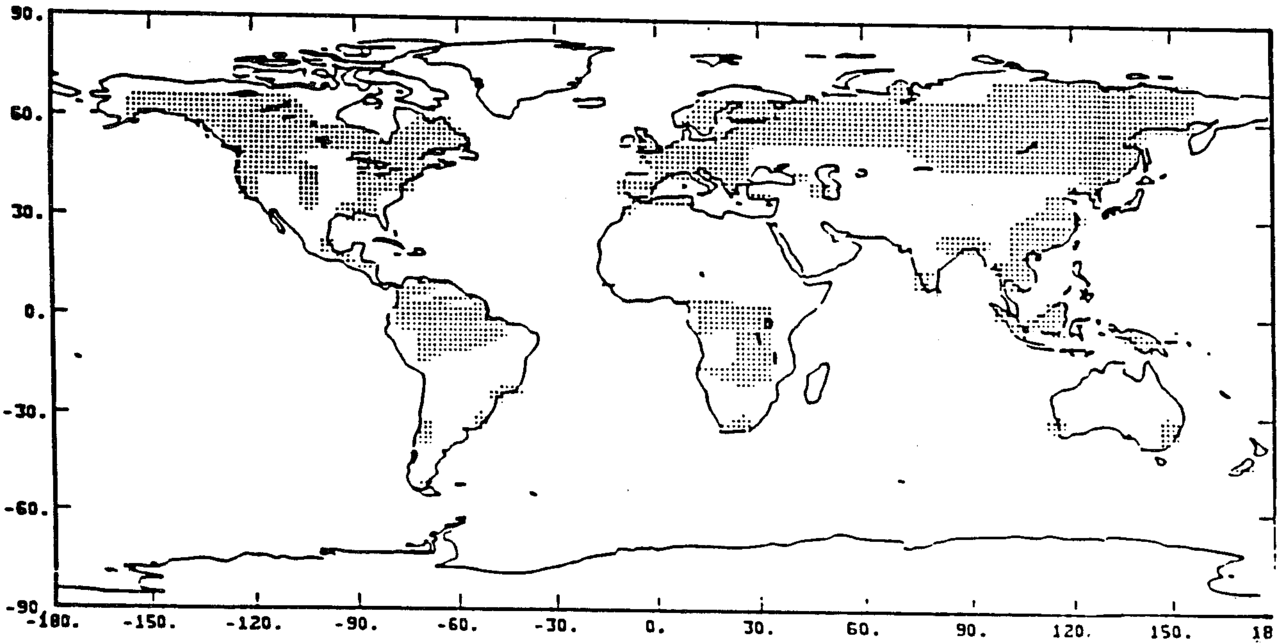


Fig. 4 : Distribution of the two types of soil defined in the two experiments called D_1 and D_2.

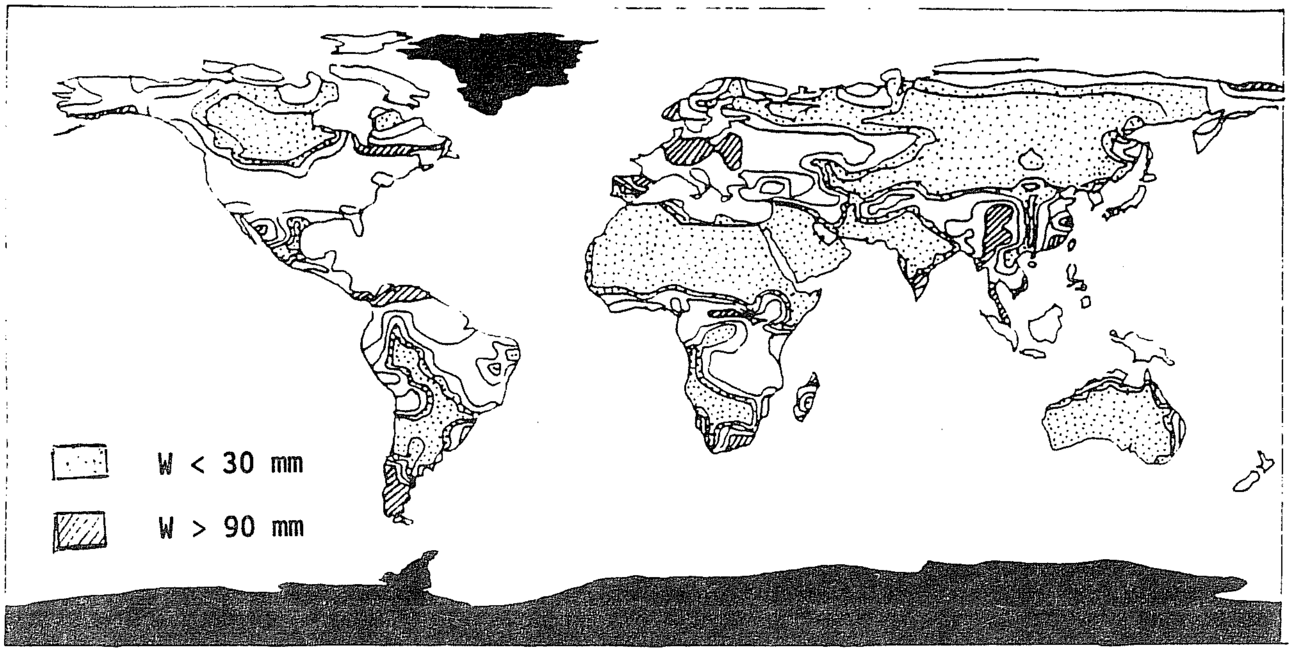


Fig. 5 : The initial distribution of soil moisture content (called A) defined in the C, D_1 and D_3 experiments.

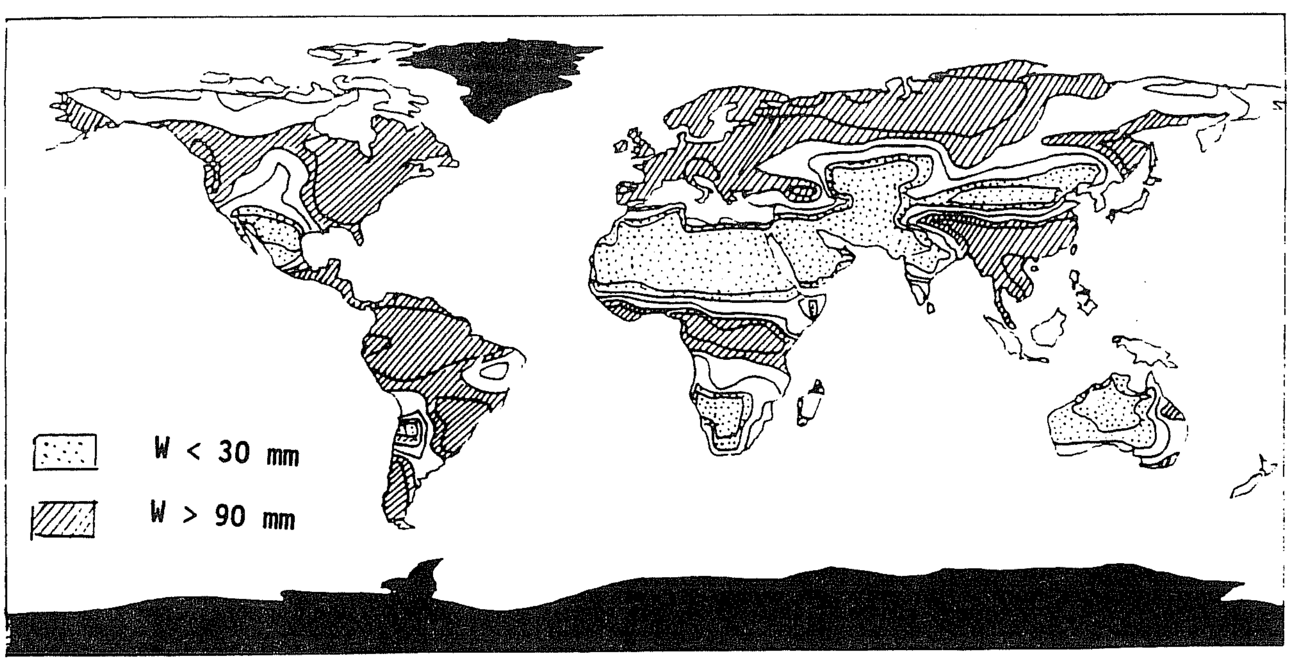


Fig. 6 : The initial distribution of soil moisture content (called MS) defined in the D_2, D_4, E and E′ experiments.

The surface resistance r was prescribed as 60 s/m in this experiment.

	C	D_1	D_2	D_3	D_4	E	E′
Evaporation rate on continents	Eq.11	Eq.11	Eq.11	Eq.11	Eq.11	Eq.11	Eq.14
Initial conditions of soil moisture	A	A	MS	A	MS	MS	MS
Parametrization of soil moisture	Bucket method	Bucket and Deardorff's methods		Deardorff's method		Choinel's method	

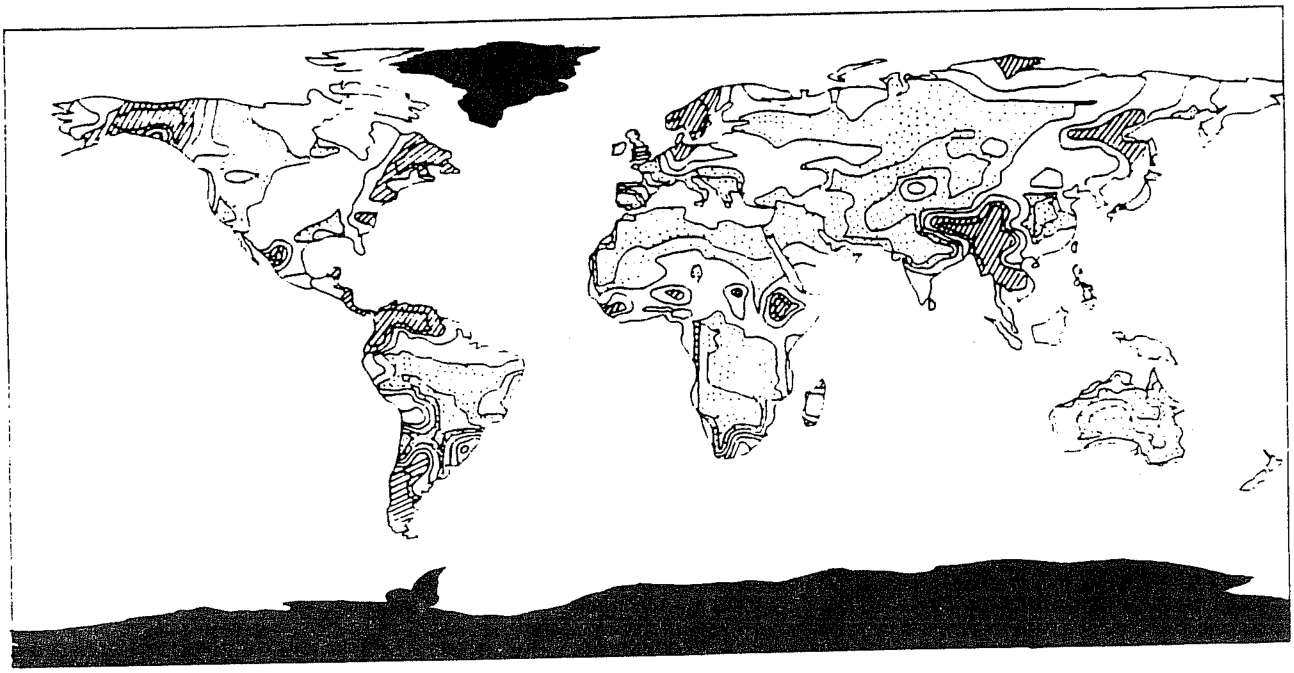
Table 1

The numerical experiments reported here have been achieved with H. Ding during his PhD thesis [H. Ding, 1989].

4. RESULTS

1) Soil moisture

Let us compare first the results obtained by the C and D_1 experiments. These two experiments start with the same initial soil moisture content, given by Fig. 5 and the model is integrated for 80 days. For comparison with other runs integrated only for 50 days, we consider the simulated climate as an average of the state of the atmosphere over the days 20 to 50. The mean soil moisture content obtained by the two experiments is shown in Fig. 7. A striking change is observed over Africa : the dry Sahara is more clearly represented and a dry region at the South of California appear in D_1 run. The humid regions covered by forest do not show important changes since the parametrization is the same for these points for the two runs, but over savannas or grassland the soil moisture content is higher in D_1 than in C. The distribution of soil moisture content obtained for D_1 is more similar to the initial conditions than the one obtained for C. It is obvious that the bucket method dries too much the soil and, as a consequence, the soil moisture content is often less than 30 mm (20%). The regions with W>60 mm are not enough extended in that case : there is a sharp gradient between wet and dry areas. The regions where Deardorff's method was applied in D_1 have, on the contrary, a very small variation a



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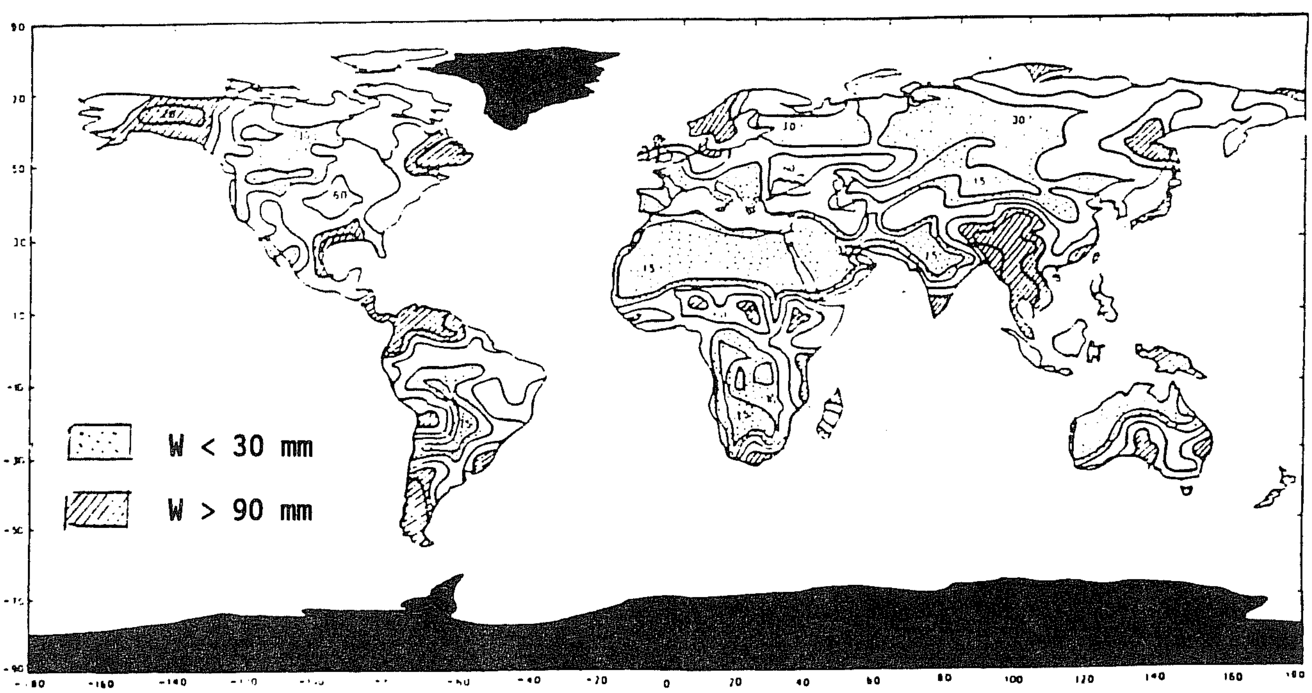


Fig. 7 : Mean soil moisture content simulated by the model for the experiments a)C and b)D_1.

with time of soil moisture content. This explains why the regions covered with savannas have higher soil moisture content in D_1.

Let us consider now the influence of initial conditions of soil moisture content, by integrating the model with an initial distribution of soil moisture content given by Mintz and Serafini (1984). We examine first the variations of soil moisture content between the two initial conditions [Fig. 5 and 6]. The values in MS distribution are generally higher than the other distribution called A. Over middle and high latitudes, in the northern hemisphere, the soil moisture content is around or higher than 90 mm in MS case and often lower than 60 mm in A case, and the difference can exceed 30 mm. Over the low latitude regions, covered by forest (Amazonia, African equatorial forest, south-east of Asia), the soil moisture content is also much lower in A case than MS case. But dry areas, like California, Sahara, deserts of Asia have lower soil moisture content in MS case. In other words, the difference of soil moisture content between an arid zone and a humid region is much more pronounced in MS case than in A case. The distribution of the bulk soil moisture content averaged over days 20-50 obtained by the run D_2 is shown in Fig. 8. It seems more realistic than D_1 since soil moisture content in high latitudes or over equatorial forest is higher. Also, the contrast between dry west and wet east U.S.A. appears more clearly. But, comparing this distribution with the initial distribution (Fig. 6), we can only conclude that soil moisture anomalies persist more than one month. Most of the differences between D_1 and D_2 are present already in the initial distribution and are over regions covered by forest, where the bucket method is used. Then, even with the formulation of evaporation that we have used and which gives lower evaporation rates than the bulk aerodynamic formula, the anomalies tend to perpetuate themselves. Since the most important differences between D_1 and D_2 occur over regions where the bucket method was applied we ran two other experiments where Deardorff's method is applied for all the points over continents. The experiment which starts with initial soil moisture content given by A distribution was called D_3 and the experiment which starts with MS distribution was called D_4.

The simulated soil moisture content, obtained by averaging the last 20-50 days of the integration, are shown in Fig. 9. These distributions are very similar to the initial values (Fig. 5,6). The differences (Fig. 10) between the initial and the simulated distribution appear over the ITCZ which has been

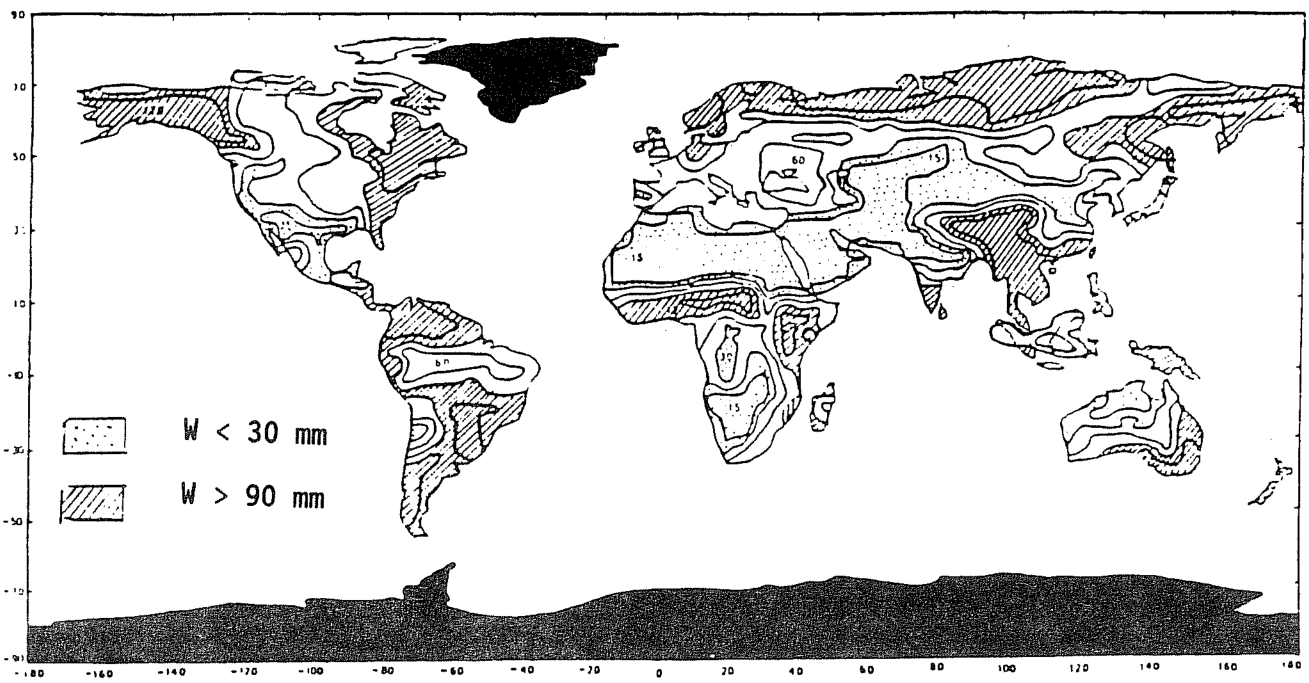


Fig. 8 : Mean soil moisture content simulated by the model for the experiment D_2.

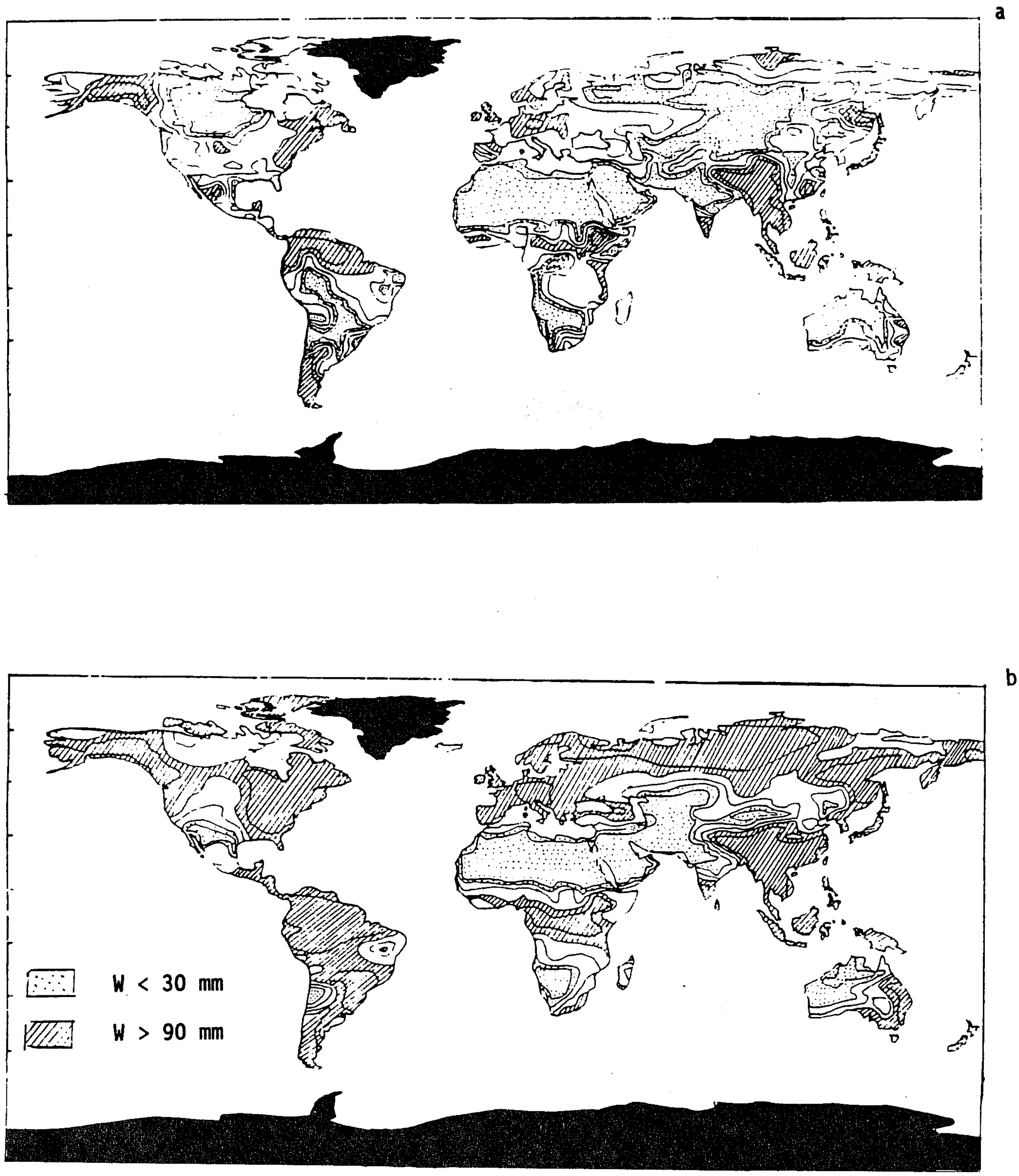


Fig. 9 : Mean soil moisture content simulated by the model for the experiment a) D_3 and b) D_4.

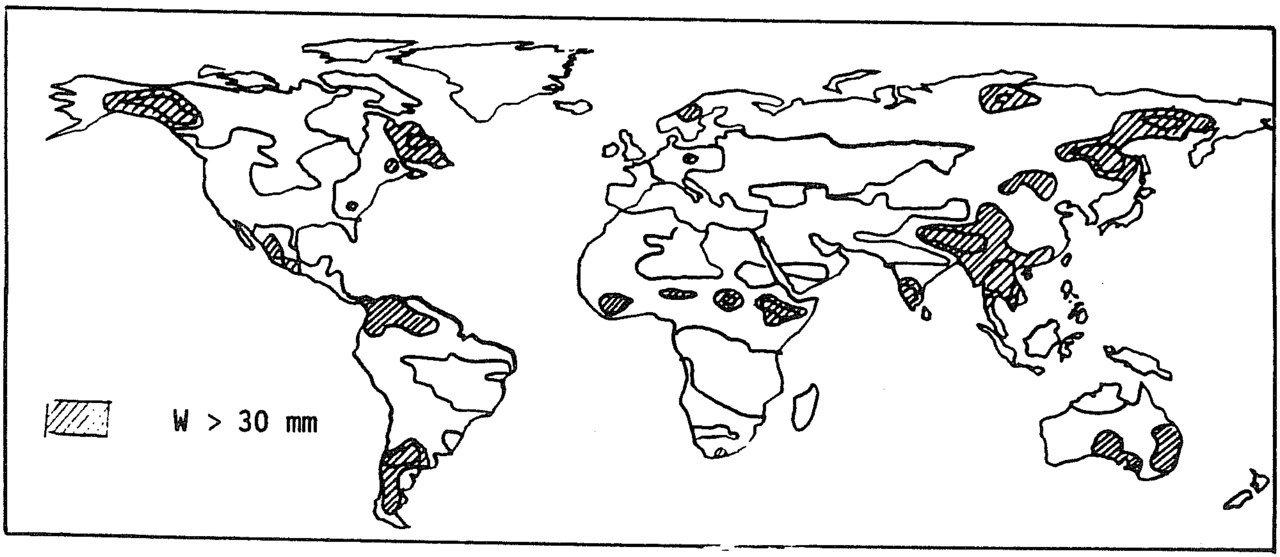


Fig. 10 : Variation of soil moisture content between the average (days 20-50) simulated by the model and the initial state for the D_3 experiment.

wetted. Over Amazonia, African equatorial forest, south-east of Asia, there is a substantial variation of soil moisture content between the averaged values simulated by D_3 and the initial values (the result obtained by D_4 is similar). It is obvious that we have obtained wetter regions in the areas with intense rainfall but we have not obtained substantial decreases of soil moisture content elsewhere. In other words, the transfer from the deep reservoir to the surface is very slow.

Noilhan and Planton (1988) have performed experiments with 1 dimensional model, using the parametrization of Deardorff for the soil moisture content. But they have calibrated the coefficients C_1 and C_2 with an hydrological model which solves the equation for soil moisture potential : they define 26 layers in the soil. They have obtained an increase of C_2 with soil moisture content : C_2, around 1 for low values of humidity, can reach 10. This increase of C_2 will tend to decrease the time scale of the transfer of moisture from the deep reservoir to the surface.

The distribution of the soil moisture content obtained with E run is shown in Fig. 11. This distribution seems the best because the drying and moistening of the different regions are the best related to the ones that Mintz and Serafini have obtained when we compare their July climatology (Fig. 12) to their June climatology. Let us note the different variations that agree with their results : a large part of the USA and of Brasil has been dried. Over Africa, the equatorial belt has been wetted and the dry region over the south has been extended. The decrease of soil moisture over France and Spain is rather well represented.

Some discrepancies still remain : our model simulates a moistening of Alaska and the North-East of Asia but these effects exist in all the experiments presented here. The run which gives the most similar variations to the E run is the D_2 run ; however, the drying of D_2 run is too much pronounced compared to MS results, reaching often 50 to 70 mm.

2) Evaporation

The distributions of the evaporation rates are shown on Figure 13. The effect of the initial soil moisture is obvious when we compare D_1 to D_2 results. The increase of the

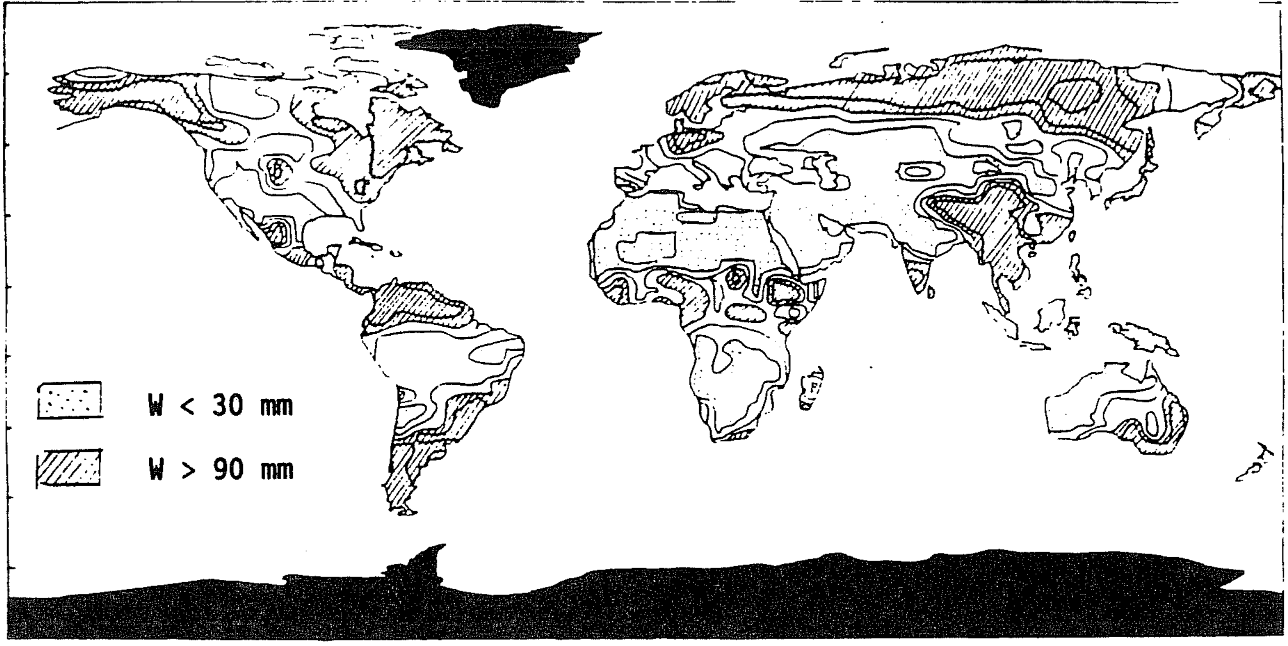


Fig. 11 : Mean soil moisture content simulated by the model for the E experiment.

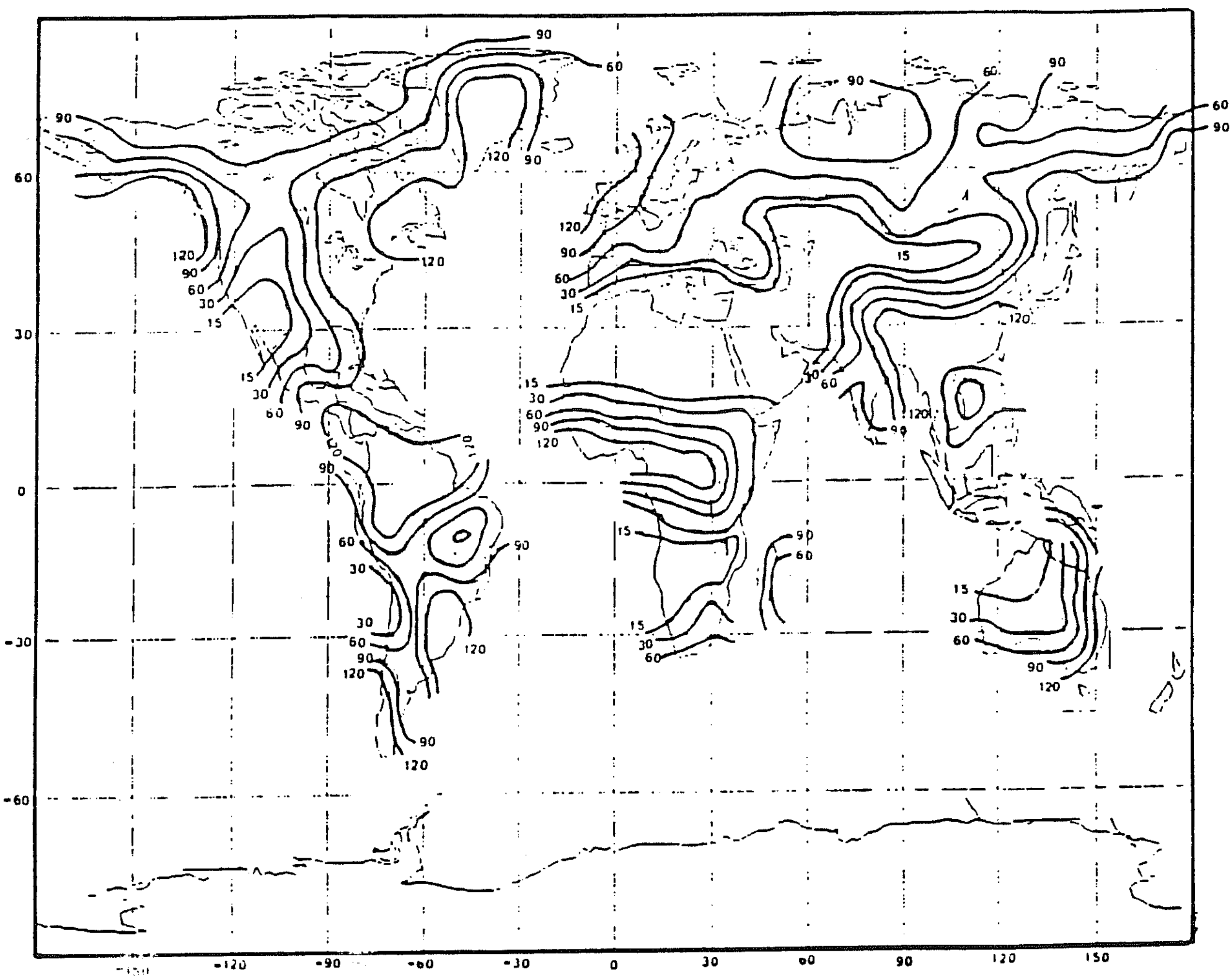


Fig. 12 : Climatology of soil moisture content for July defined by Mintz and Serafini (1984).

evaporation rate over the East of USA and Canada, Amazonia, the high latitudes of Europe and Asia are all related to the variations of initial moisture content. But this influence does not exist at all when we compare D_3 and D_4 integrations which use Deardorff's method for all continental points. On the contrary, the distributions are quite similar. Then the influence of the initial soil moisture content occurs only if the transfer from deep to surface reservoir is sufficiently fast.

The distribution obtained with E run gives high values over Amazonia, equatorial Africa, east of the USA and Canada, and Europe. These values agree rather well with the ones obtained by Mintz and Serafini (1984) with their hydrological model. But the GCM fails to simulate the low rates over the Sahara.

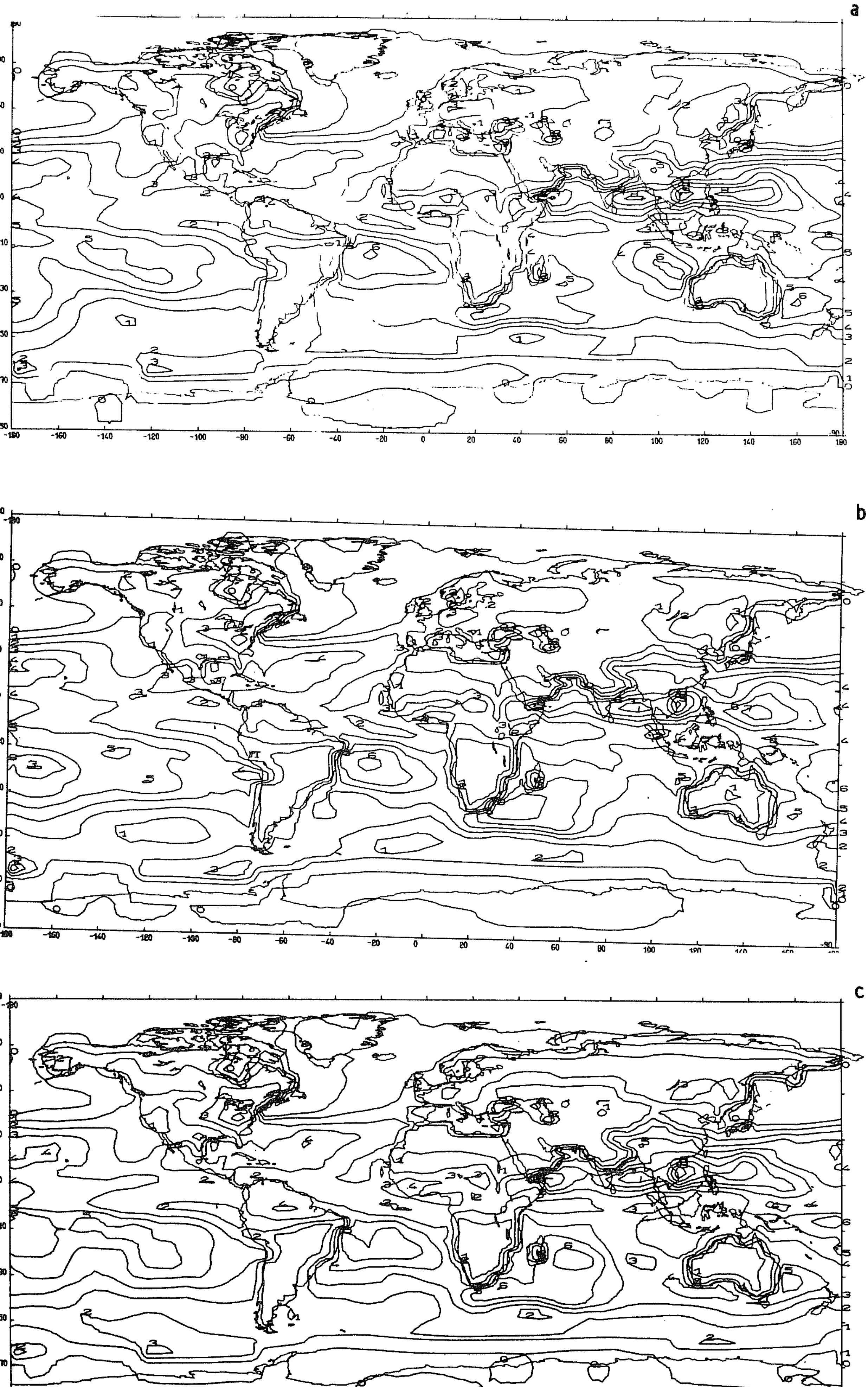
3) Precipitation

Compared to observed values, the distribution obtained with E experiment seems quite realistic (Fig. 14). The regions with high values are rather well represented : east of the USA and Canada, Central America, Patagonia, Europe ; the intertropical convergence zone with a northward shift over eastern longitudes, with the maxima rather well located over equatorial Africa, over India and Burma. Compared to this field, the ones obtained with D_3 run (not shown here) or D_u run quite similar, show too low values (for example over Europe). The distribution obtained with D_2 experiment is rather realistic but some failure are obtained : too low values over the northern middle latitudes, the maxima of the equatorial Africa located too eastward, the maximum over India not enough high compared to the one over Burma ; and high values, that were obtained also with the D_u run, over Bolivia that are quite unrealistic.

4) Evolution with time of hydrological variables

Time series of different variables for several regions are analysed in this paragraph.

Let us consider first the Sahel region. For the run C, the soil moisture content increases slowly from low initial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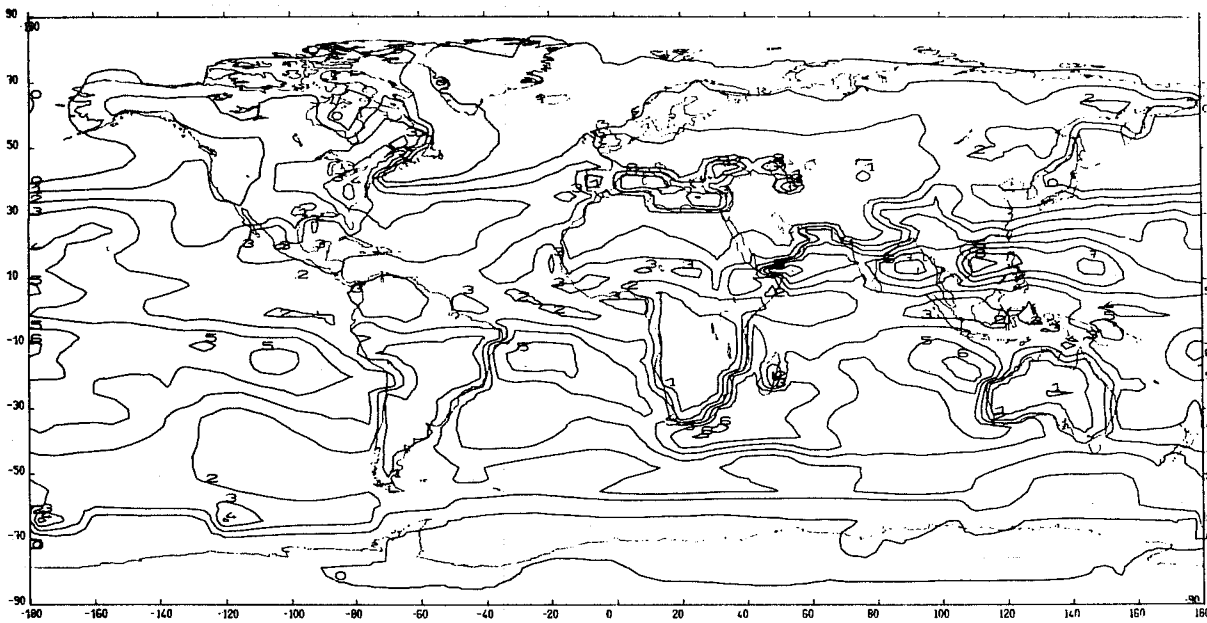

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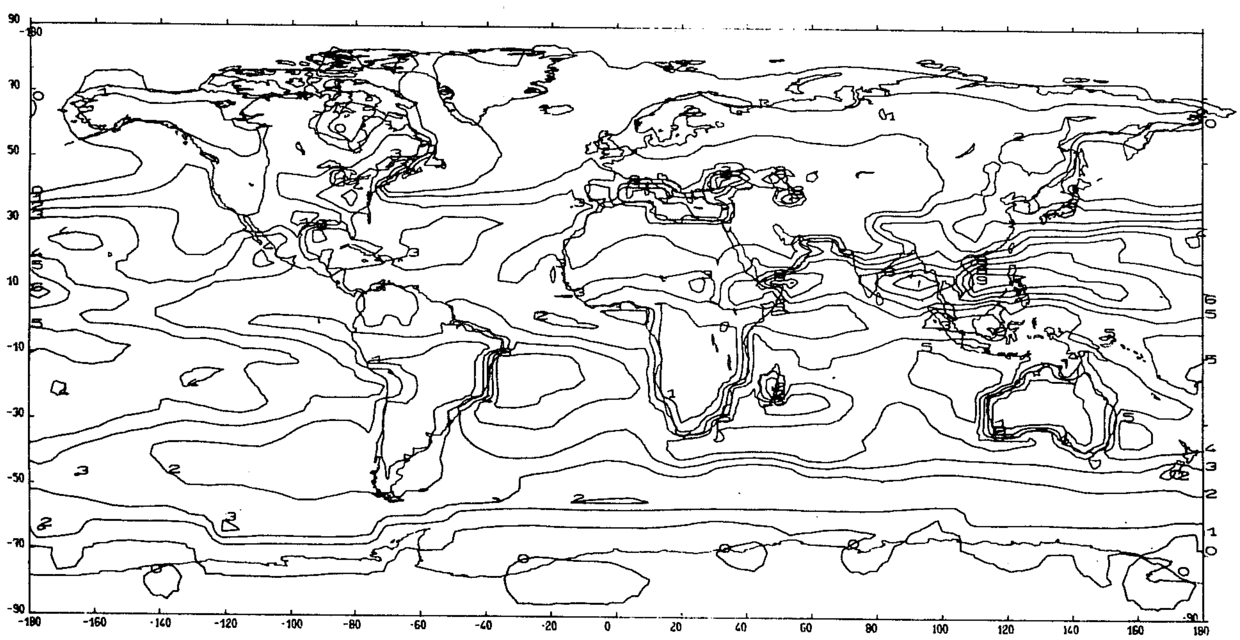

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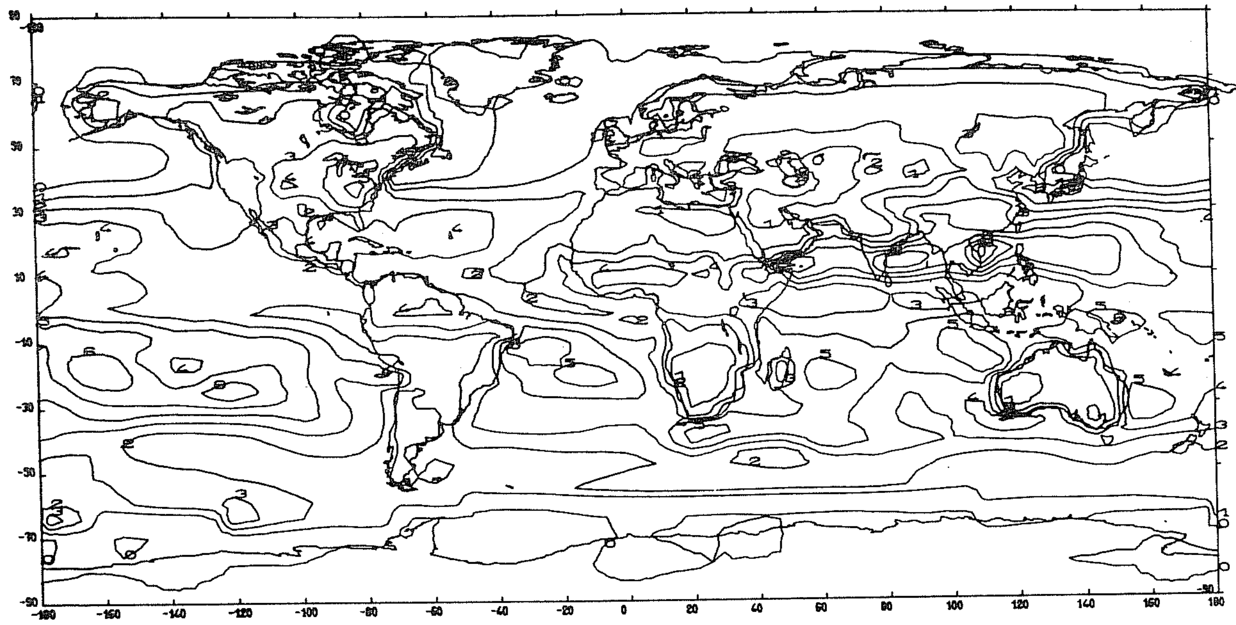

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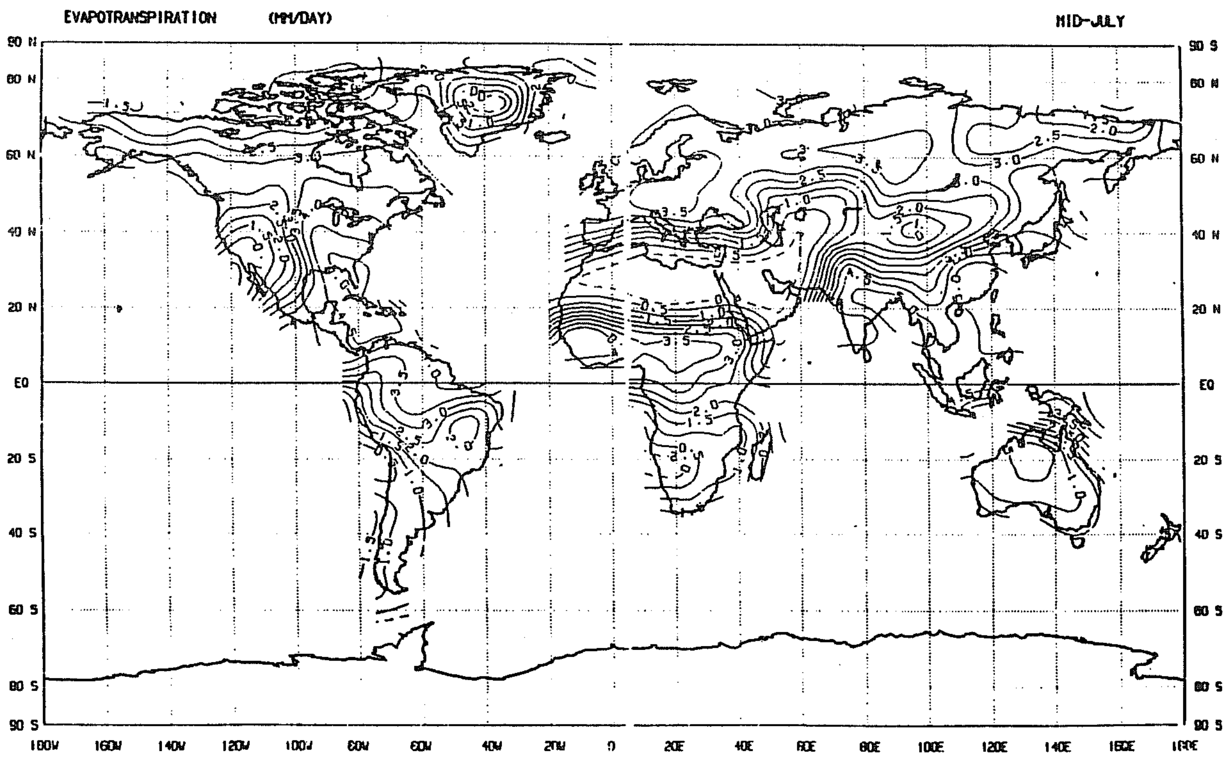
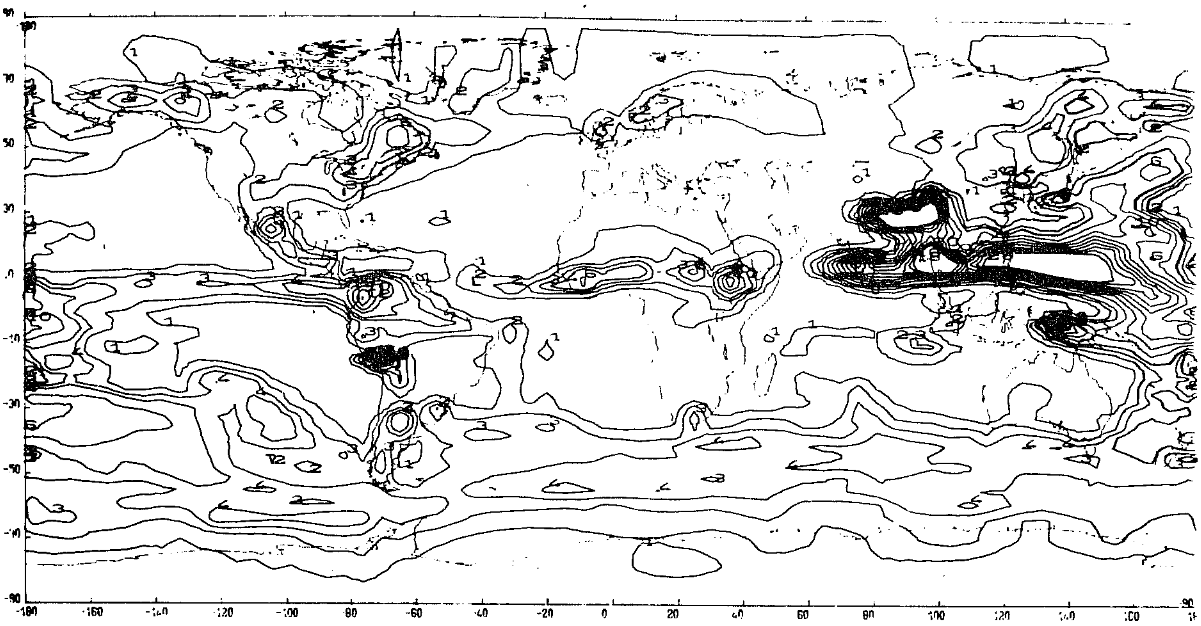


Fig. 13 : The mean evaporation rate simulated by the model for the experiments a)C, b)D_1 ; c)D_2, d)D_3, e)D_4, f)E, g) The mean evaporation rate estimated by Mintz and Serafini (1984) for July.



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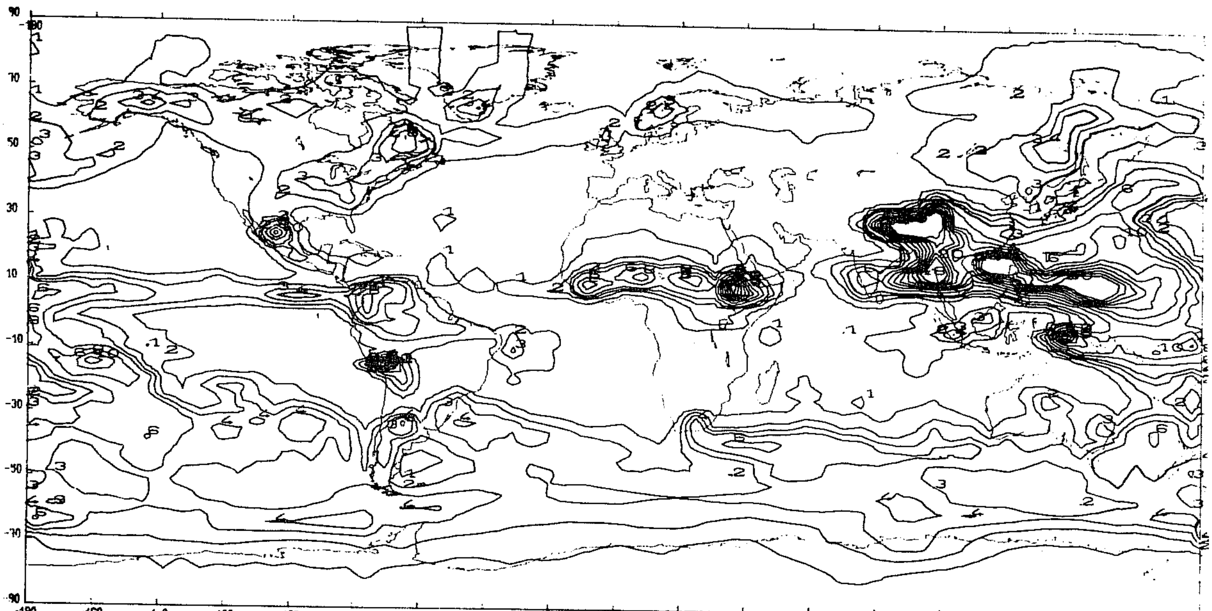

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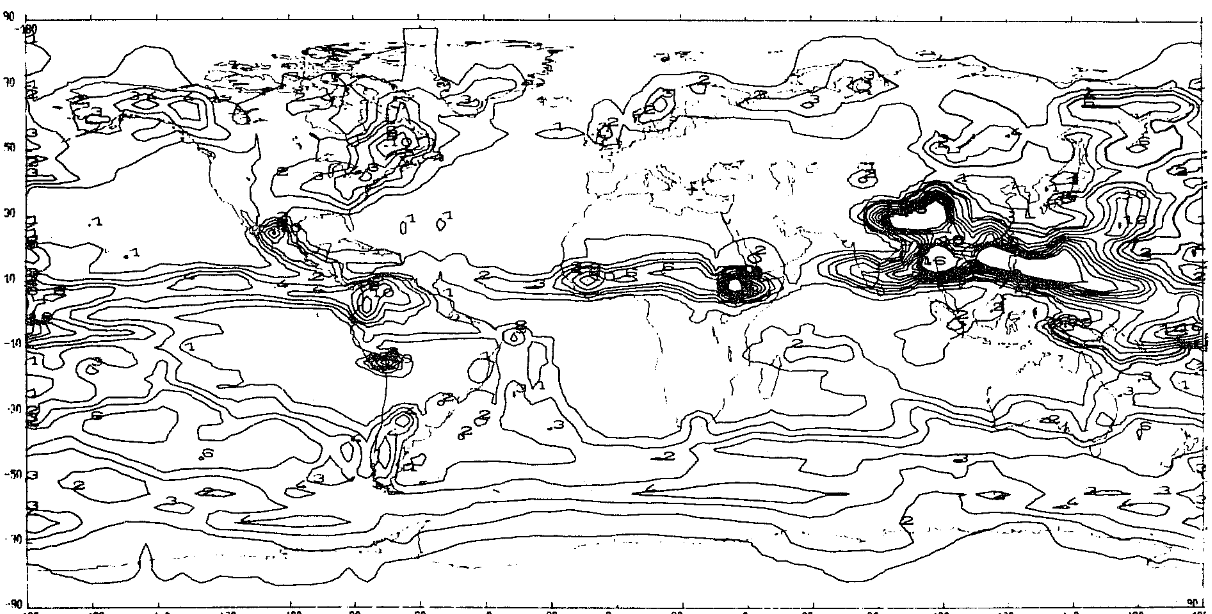

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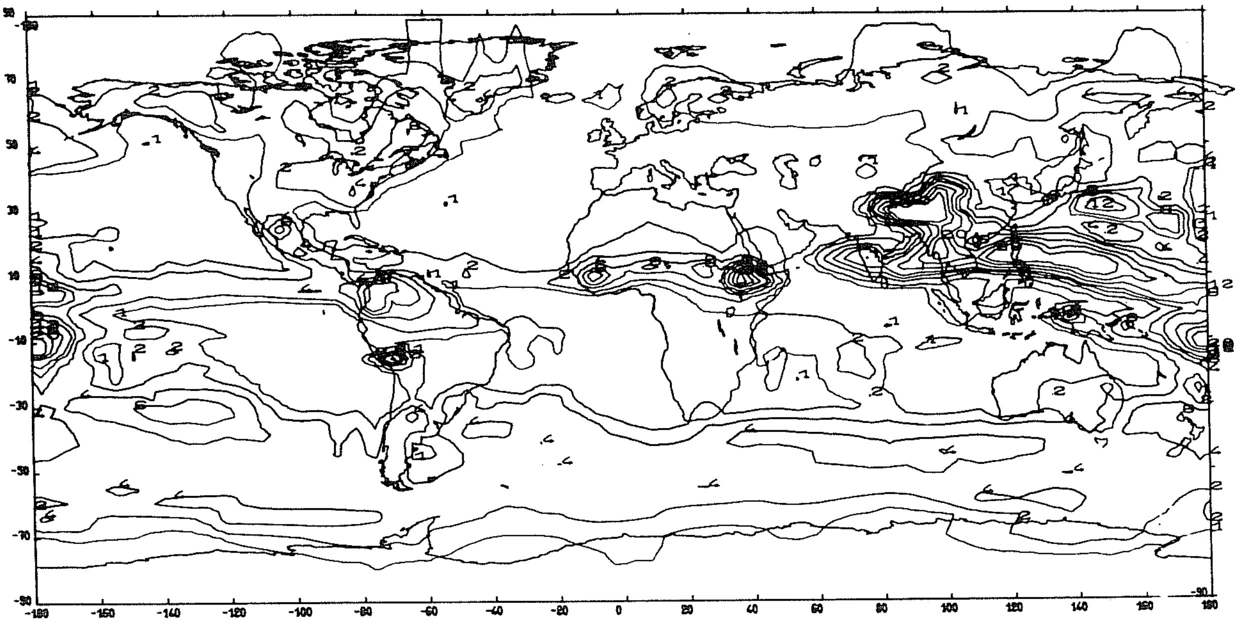
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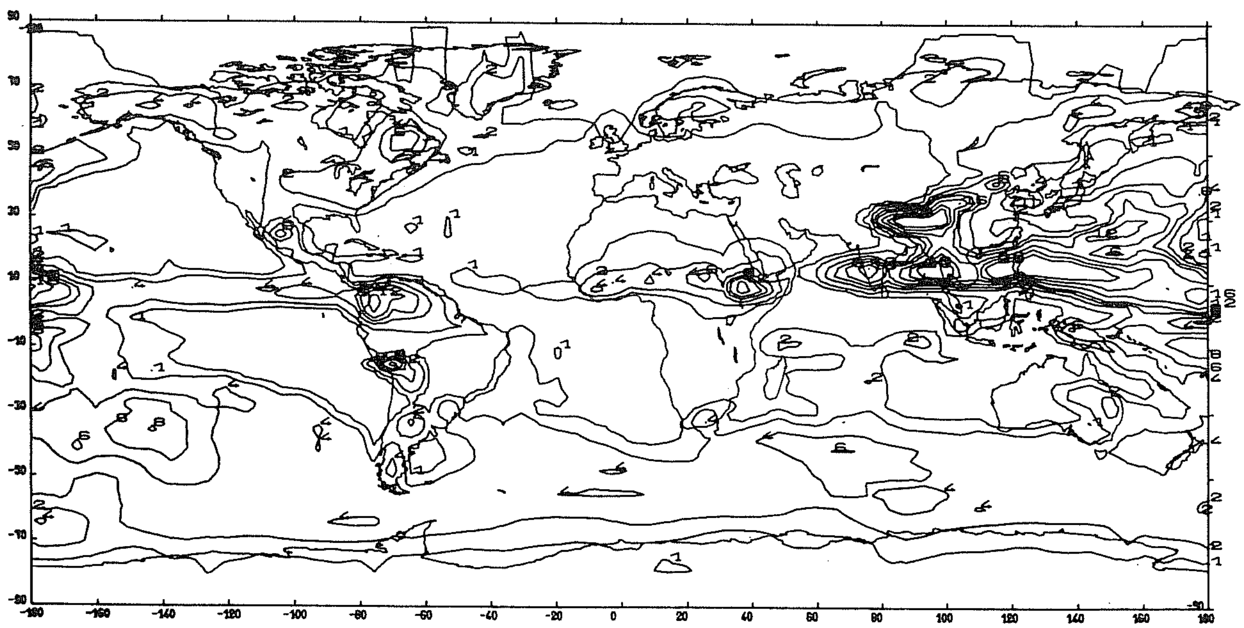
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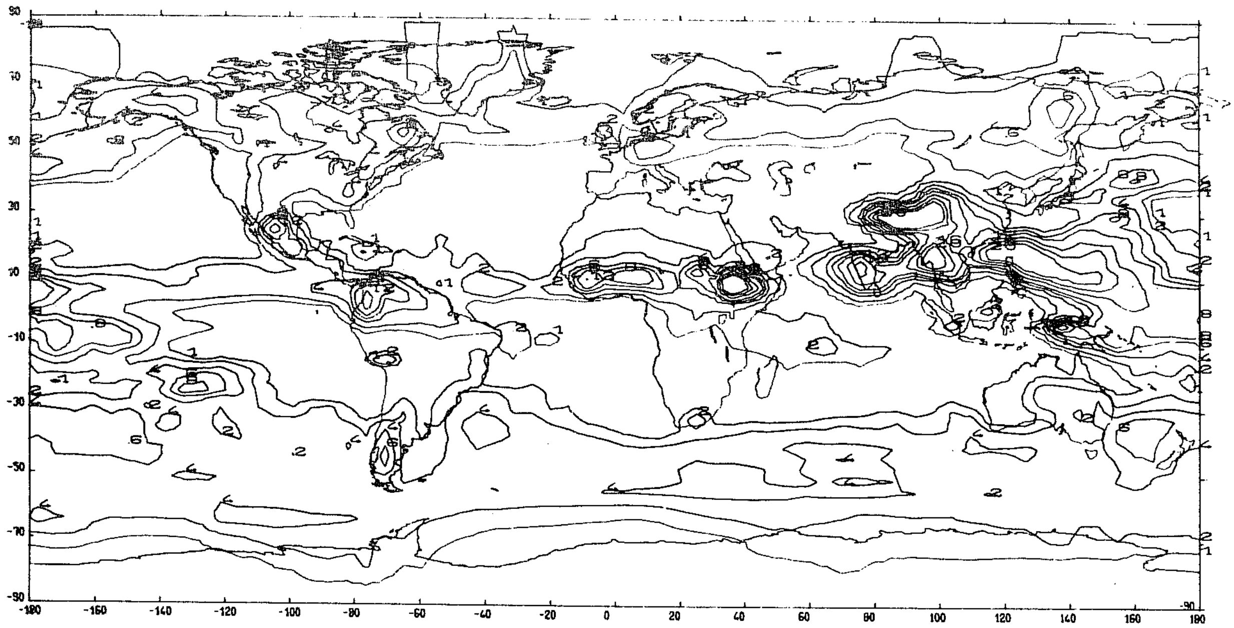

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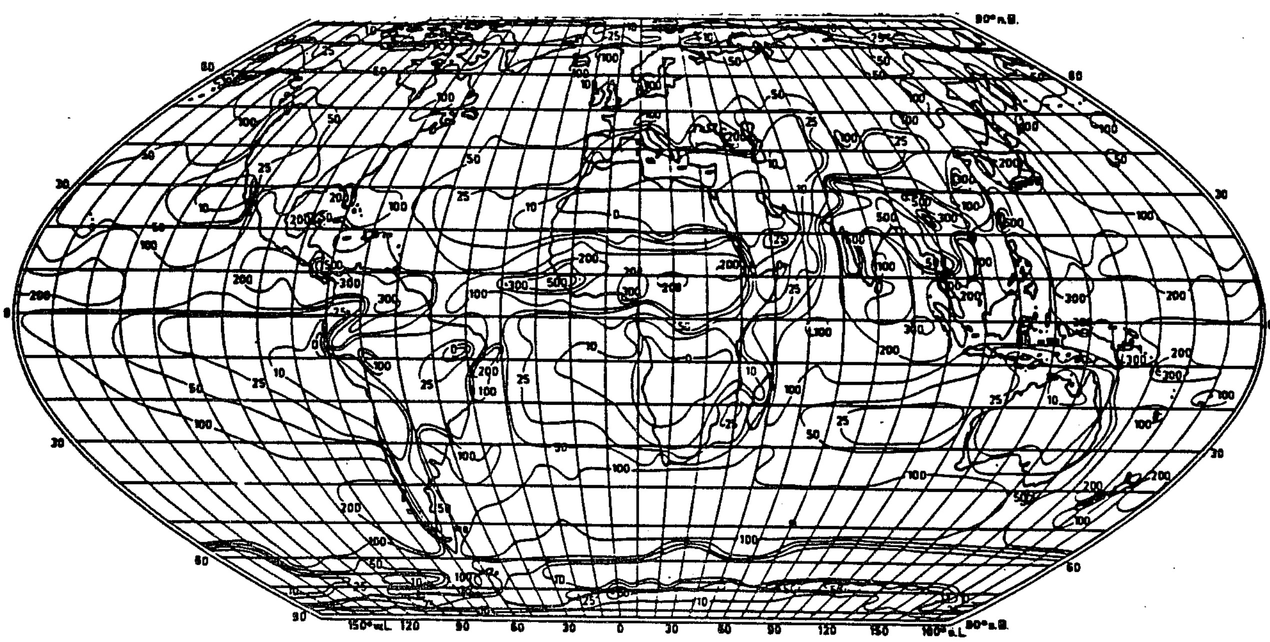

Fig. 14 : The mean precipitation rate simulated by the model for the experiments a)C, b)D_1, c)D_2, d)D_3, e)D_4, f)E, g) observed rate as given by Jaeger (1976).

values to 30 mm and the evolution of the evaporation rate follows this variation (Fig. 15a). But in the case of Deardorff's method (D_1 run), the bulk soil moisture content does not change for 20 days (Fig. 15b). Then an abrupt change occurs towards day 30, together with a very high precipitation rate (Fig. 16). In this run, the evaporation rate follows the variation of surface soil moisture content instead of the bulk soil moisture content. Over this region, where the soil is moistening, we observe : that the two methods (C and D_1) show the increase of soil moisture content but with Deardorff's method, it is obtained over a rather short period ; that the evolution of the evaporation rate shows very rapid fluctuations with Deardorff's method compared to bucket method because this rate is related to surface soil moisture ; and finally that the rates of evaporation and precipitation are much higher in D_1 experiment than the ones obtained in C experiment.

We analyse the results obtained over the Europe with D_3 and D_u experiments (Fig. 17). The initial bulk soil moisture content is quite different for the two runs and this difference is maintained during the integration. The initial surface soil moisture contents differ also from one another because we have prescribed, as an initial value, the same surface relative humidity that the bulk relative humidity for each run. Consequently, we observe a variation of 1 mm/day in the initial values of the evaporation rate (Fig. 18). But after a few days, the time evolution is quite similar for the two runs. The precipitation rates also are almost the same for these two runs over Europe. We have the same conclusions for other temperate regions.

Let us consider now over these two regions, the variations obtained with Choinel's method (run E). Over the Sahel, the evaporation rate shows rapid fluctuations like with the Deardorff's method (Fig. 15b) but over Europe, the soil moisture decreases as in the bucket method but more slowly. Then this parametrization can simulate altogether the two types of variations obtained with the bucket method and the Deardorff's method.

The last remark concerns the E'run compared to E run. We have assumed in E' experiment a surface resistance of 60 s/m which can be a realistic value for forests or savannas in wet cases, but is certainly too small for the surface resistance of bare soil. We have obtained with this integration a distribution of the evaporation rate which shows too large values over arid

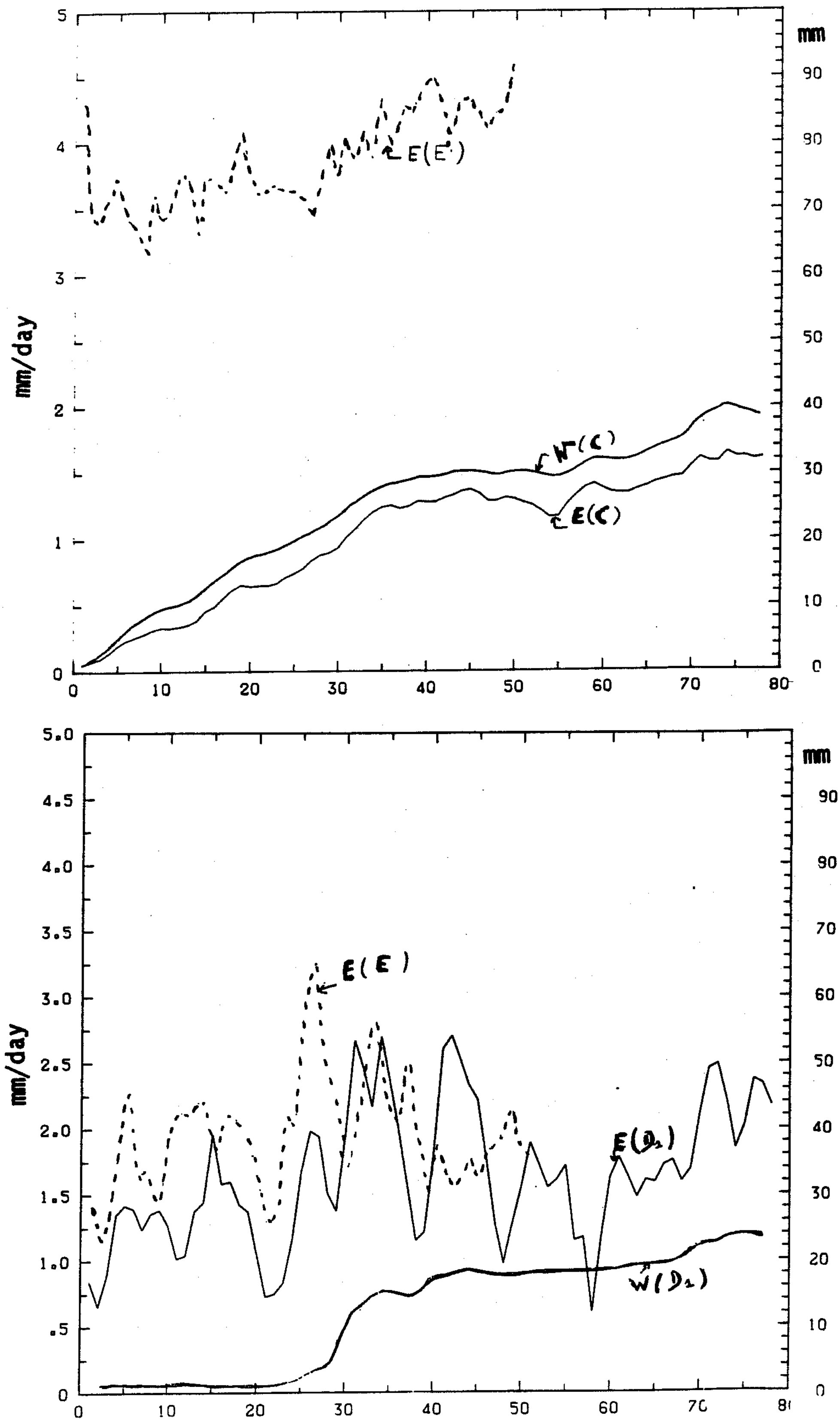

Fig. 15 : Evolution of the bulk soil moisture content and the evaporation rate over the Sahel obtained a) with C and E' and b) with D_1 and E.

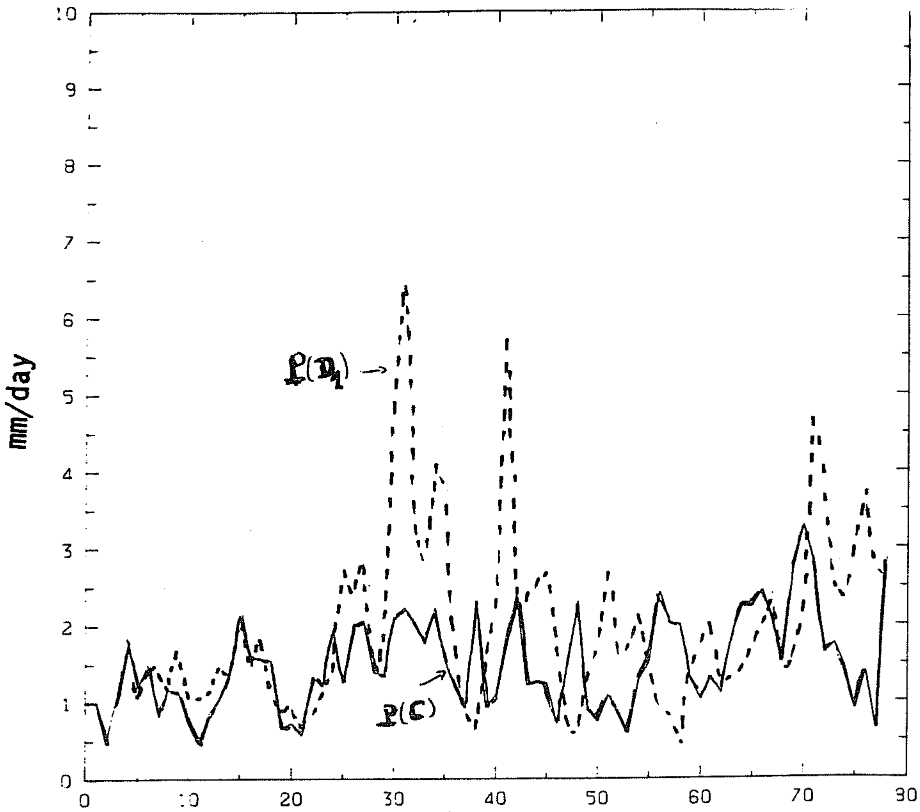

Fig. 16 : Evolution of the precipitation rate over the Sahel obtained with C and D_1.(mm / day)

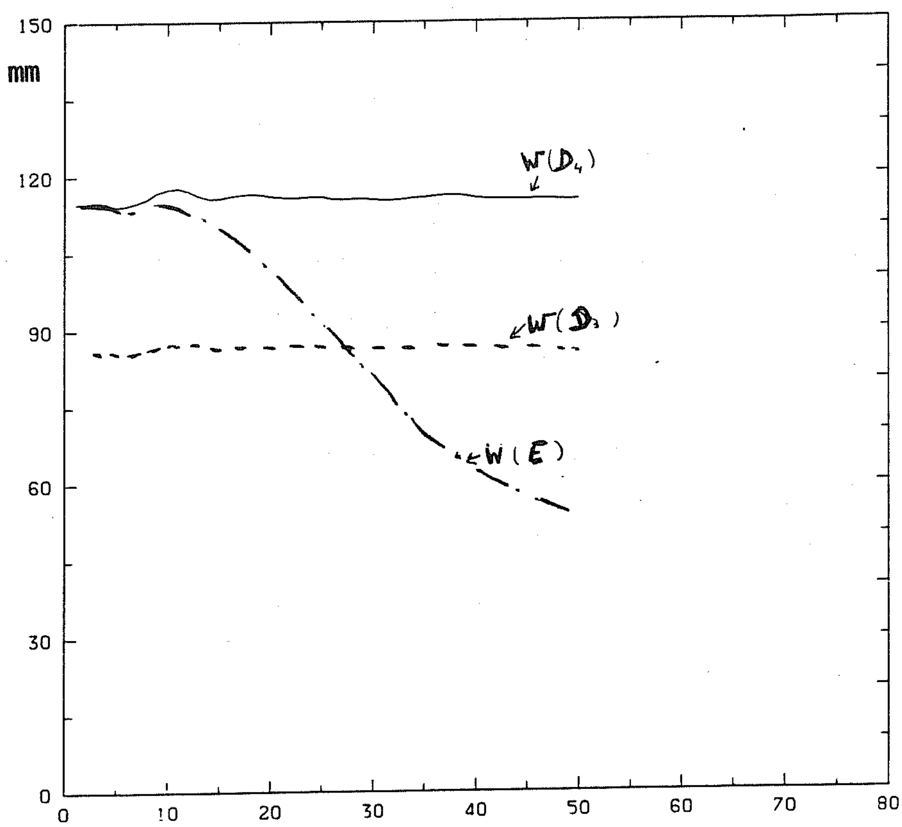

Fig. 17 : Evolution of the bulk soil moisture content over the Europe obtained with the D_3, D_4 and E experiments.

and semi arid zones. For example, the evaporation rate over the Sahel (Figure 15a) reach 3 mm/day. But this failure is true for all the arid zones : we show in Fig. 19 the evaporation rate over the west of USA (30°N → 40°N and 110°W → 120°W) obtained with E and E′ experiments : there is also here a too high evaporation rate in E′ run. Then the use of the concept of surface resistance, in a simple way as we have done in E′ experiment does not allow to obtain low evaporation rates over the arid zone. On the contrary, the formula (11), of Priestley and Taylor type, where the evaporation rate is related to net radiation seems more appropriate.

Over forests, we get completely different results . We show in Fig. 20 the evolution with time of the evaporation rate over two regions covered by rain forests : the Amazonia and the equatorial forest of Africa (Fig. 21). The evolution with time obtained by the two experiments E and E′ are quite similar. These regions, which are very humid, evaporate at the potential rate and, with our model, the choice of the parametrization of soil moisture made in these runs, the two methods to compute evaporation give quite the same results.

V. CONCLUSIONS

The study we have conducted with different formulations of soil moisture with our GCM gives the following results :

. The bucket method dries too much the continents and gives subtantial variations of the bulk soil moisture.

. The Deardorff's method gives rapid fluctuations of surface soil moisture and consequently of the evaporation rate. In the regions which are drying, this method does not allow a sufficient transfer from the deep to the surface reservoir. Over these regions and with this method, the dependence of the evaporation rate on soil moisture anomalies is maintained only for a few days.

. The Choisnel's method gives with our model the most realistic rainfall and probable evaporation rate. The time evolution of this last variable can show rapid or slow variations, depending on the budget of soil moisture.

. The Monteith's equation, used with a fixed surface resistance gives unrealistic high values of the evaporation rate over the semi-arid zones with our model, but the method that we have used

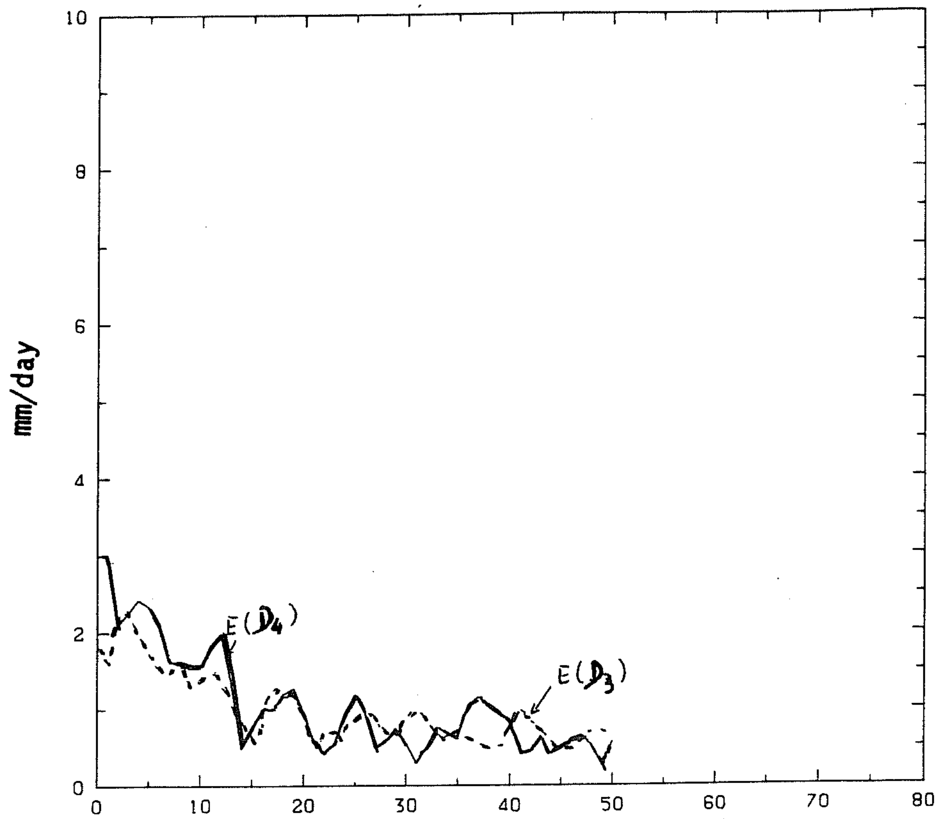

Fig. 18 : Evolution of the evaporation rate obtained over the Europe with the D_3 and D_4 experiments.

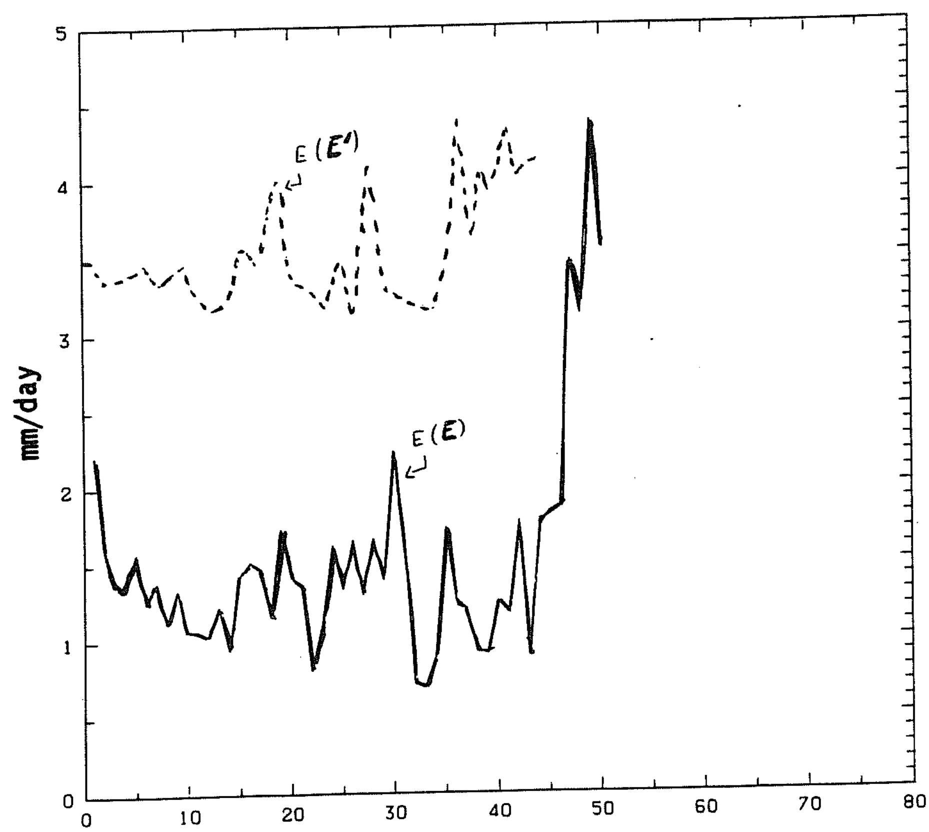

Fig. 19 : Evolution of the evaporation rate over the west of USA obtained with E and E′ experiments.

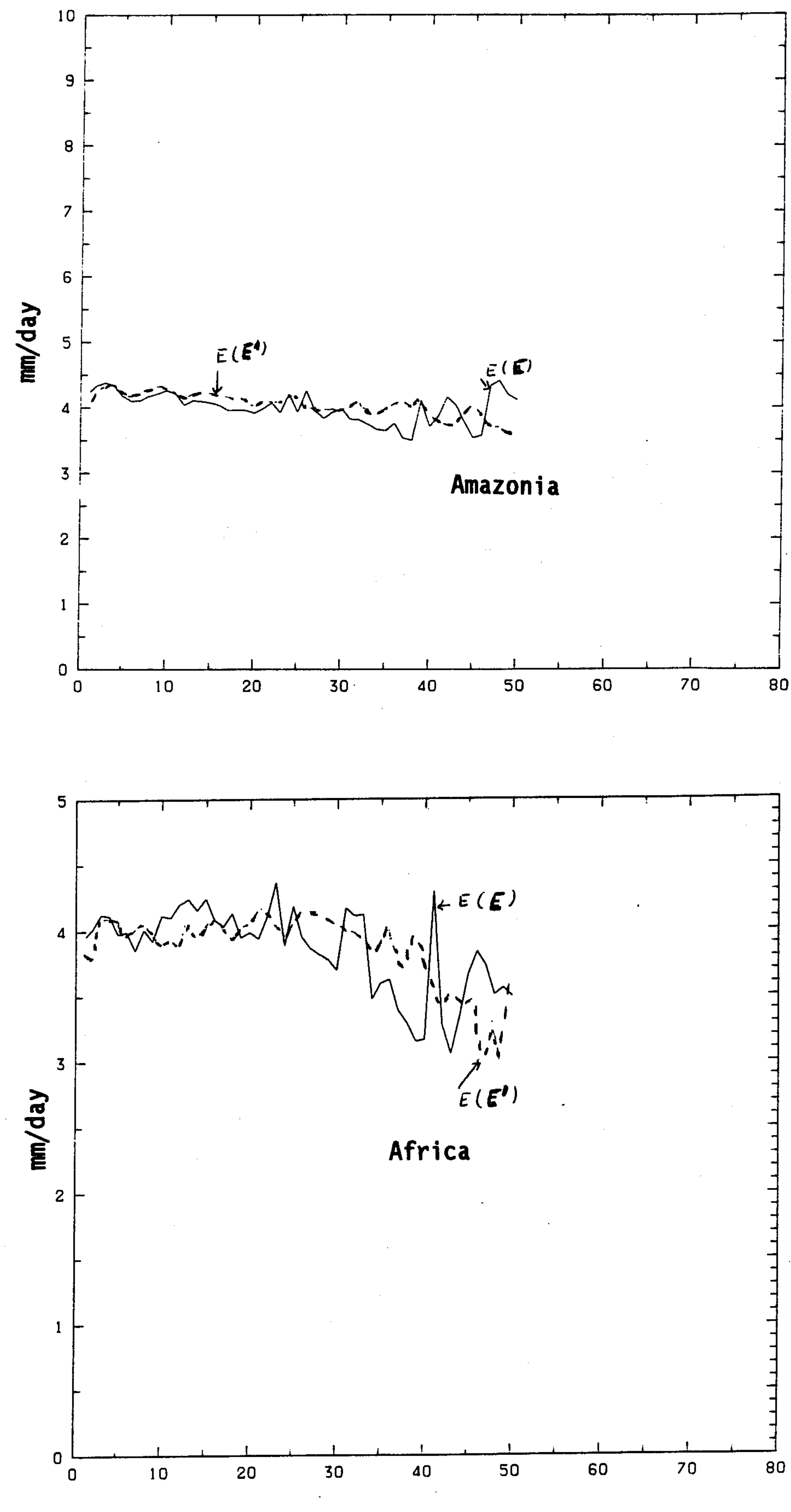

Fig. 20 : Evolution of the evaporation rate over the forests of Amazonia and Africa with E and E′ experiments.

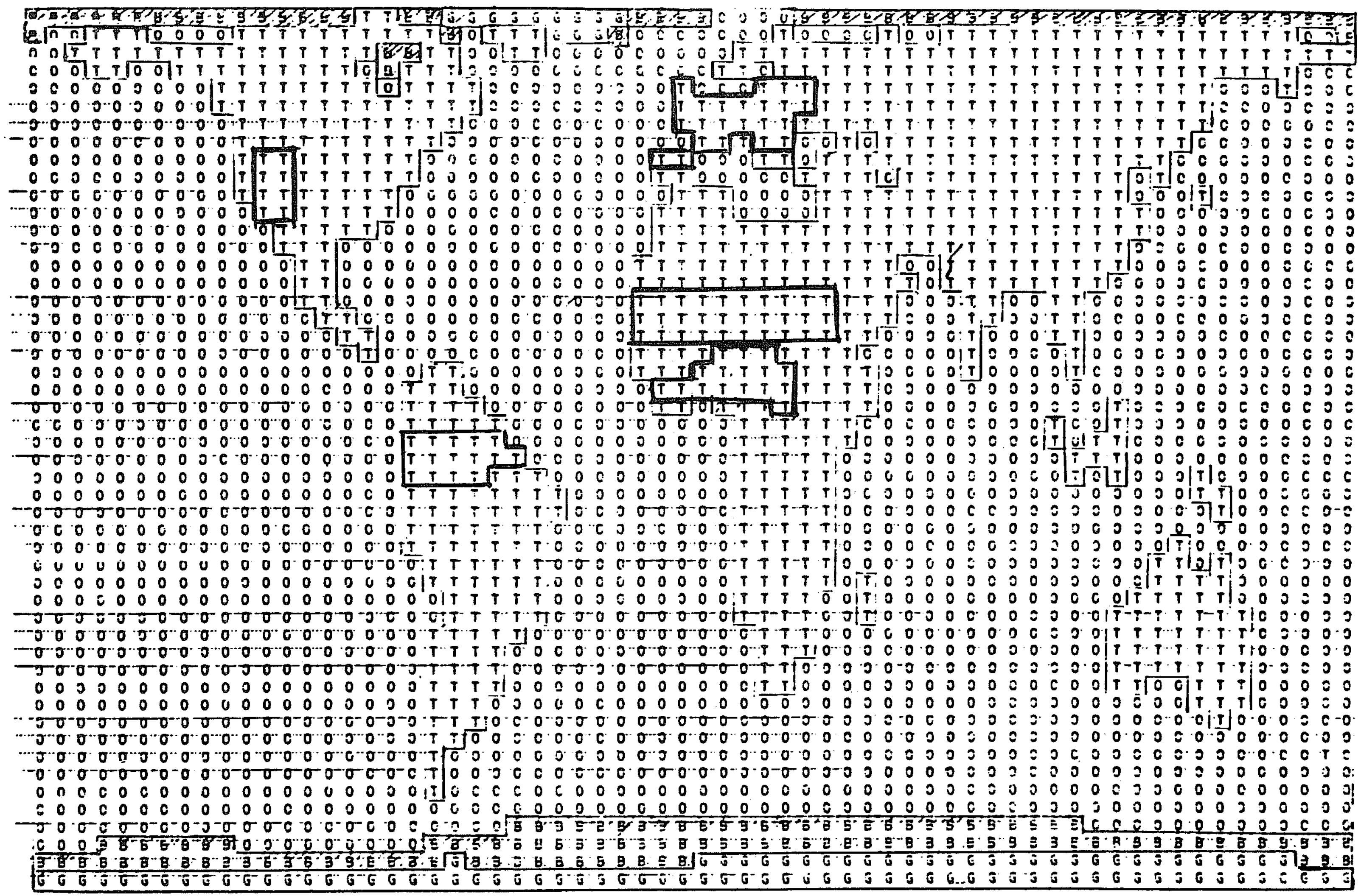

Fig. 21 : Distribution of the continental points of the model.
The areas used for the diurnal variation studies are encircled by solid lines.

for the other integrations seems to simulate the decrease in these zones quite realistically. On the contrary, over regions with high rainfall, the two methods are quite similar.

We should like to emphasize that the results that we have shown for the influence of soil moisture are dependent on the parametrization of the evaporation rate and the results that we have obtained on the influence of the scheme of the evaporation rate are dependent on the soil moisture parametrization. However, the conclusions that we have given are strengthened by the 7 runs that we have reported here and some others that we have not discussed here.

References

Choisnel, E., 1977 : 'Le bilan d'énergie et le bilan hydrique du sol' La Météorologie. Numéro spécial "Evapotranspiration". VIe série, n°11, pp. 103-159.

Deardorff, J.W., 1977 : 'A parameterization of ground surface moisture content for use in atmospheric prediction models.' J. Appl. Meteor., 16, pp 1182-1185.

Dickinson, R., 1984 : 'Modeling evapotranspiration for three-dimensional global climate models.' In Climate Processes and Climate Sensitivity, eds. J.E. Hansen and T. Takahashi, Geophys. Monogr., 29, Amer. Geophys. Union, 58-72.

Ding, H., 1989 : Etude du climat simulé par le MCG du LMD avec différentes représentations de l'humidité du sol. PhD Thesis -Université Paris 6, 200 pp.

Hunt, B.G., 1985 : 'A model study of some aspects of soil hydrology relevant to climatic modelling.' Quart. J. Roy. Meteor. Soc., 111, pp 1071-1085.

Jaeger, L., 1976 : Monatskarten des Niederschlages für die ganze Ende. Berichte Deutscher Wetterdienst, N 139, Offenbach. 38 pp.

Laval, K., C. Ottlé, A. Perrier and Y. Serafini, 1984 : 'Effect of parameterization on climate simulated by a GCM.' In New Perspectives in Climate Modelling, Elsevier, Amsterdam, 223-247.

Laval, K., and L. Picon, 1986 : 'Effect of a change of the surface albedo of the Sahel on climate.' J. Atmos. Sci., 43, 2418-2429.

Manabe, S., 1969 : 'Climate and the ocean circulation. The atmospheric circulation and the hydrology of the Earth's surface.' Mon. Wea. Rev., **97**, 739-774

Mc Naughton, K.G., 1976 : 'Evaporation and advection' evaporation from extensive homogeneous surfaces.' Quart. J. Roy. Meteor. Soc., **102**, pp 181-191.

Mintz, Y. and Y. Serafini, 1984 : Global fields of normal monthly soil moisture, as derived from observed precipitation and an estimated potential evapotranspiration. **Part V**, *Final Scientific Report Under NASA Grant No*. NAS5-26. Dept. Meteorology, Univ. of Maryland, College Park, Md 20742.

Monteith, J.L., 1965 : Evaporation and environment in G E Fogg, (ed.). The state and the movement of water in living organisms, sympos. Soc. Exper. Biol., Vol 19, Academic Press, N.Y., pp. 205-234.

Noilhan, J., and S. Planton, 1988 : 'A simple parameterization of land surface processes for meteorological models' Mon. Wea. Rev., to be published.

Penman, H.L., 1948 : 'Natural evaporation from open water, bare soil and grass'. Proc. Roy. Soc., London, A193, 120-146.

Perrier, A., 1982 : 'Land surface processes : vegetation'. Land surface processes in atmospheric circulation models, ed. P.S. Eagleson, Cambridge University Press, Cambridge, pp 395-448.

Priestley, C.H.B., R.J. Taylor, 1972 : 'On the assessment of surface heat flux and evaporation using large scale parameters.' Mon. Wea. Rev., **100**, 81-92.

Rowntree, P.R. and J.A. Bolton, 1983 : 'Simulation of the atmospheric response to soil moisture anomalies over Europe.' Quart. J. Roy. Meteor. Soc., **109**, pp 501-526.

Sellers, P.J., Y. Mintz, Y.C. Sud, and A. Dalcher, 1986: A simple biosphere model (SiB) for use within general circulation models. J.Atmos.Sci., 43, 505-531.

Walker, J.M. and P.R. Rowntree, 1977 : 'The effect of soil moisture on circulation and rainfall in a tropical model.' Quart. J. Roy. Meteor. Soc., **103**, pp 29-46.

Yeh, T.C., R.T. Wetherald and S. Manabe, 1984 : 'The effect of soil moisture on the short term climate and hydrology change. A numerical experiment.' Mon. Wea.Rev., **112**, 474-490.