SENSITIVITY OF GENERAL CIRCULATION MODELS TO
LAND SURFACE PROCESSES

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1. LAND SURFACE PROCESSES

Land surface processes affect the atmosphere by modifying the transfer of heat, moisture and momentum between surface and atmosphere. The surface balance equations representing the heat and moisture transfers may be written:

$$G_0 + H + LE = (1 - \alpha)R_S + \epsilon_{*\downarrow} R_{L\downarrow} - R_{L\uparrow} \equiv R_N \quad (1)$$

$$M_0 + Y_0 + E = P \quad (2)$$

Here G_0, M_0 are the fluxes of heat and water into the soil;

H, E are the fluxes of heat and moisture into the air;

Y_0 is the run off of water along the surface;

R_S, R_N are the downward solar flux and downward net radiation;

$R_{L\downarrow}$, $R_{L\uparrow}$ are the downward and upward long wave fluxes with the upward flux $R_{L\uparrow} = \epsilon_{*\uparrow} \sigma T_*^4$;

T_* is surface temperature;

$\epsilon_{*\downarrow}$, $\epsilon_{*\uparrow}$ are surface emissivities for downward and upward longwave radiation;

α is the surface reflectivity for the downward solar radiation;

L is the latent heat of evaporation or sublimation, depending whether the surface is unfrozen or frozen.

Note that the turbulent fluxes of heat and moisture (H, E) and momentum ($\underline{\tau}$) are affected by the surface roughness usually represented by the surface roughness length Z_0.

A few processes are neglected in the above including the freezing and thawing of water, ice and snow at the surface and the effects of precipitation on the heat budget. The consideration of a layer of snow complicates the treatment of land surface processes, mainly because the subsurface processes such as percolation of water and transfer of heat are greatly changed. As we are here concerned with the interface between the atmosphere and the underlying medium we shall neglect snow except for its direct effect on the fluxes through the radiative parameters and moisture availability.

Note that the emissivities and solar albedo can vary both through changes in the nature of the surface as occurs with snow and with drying or moistening, and also through variations in the spectral distribution of the relevant radiation caused by varying solar zenith angle, clouds and atmospheric aerosol, surface temperature and atmospheric temperature and moisture content.

The land surface's influence on the atmosphere as expressed in the above equations will vary (a) because of changes in the parameters $\alpha, \epsilon_{*\downarrow}, \epsilon_{*\uparrow}, z_0$; (b) (i) because of changes in the surface temperature and surface moisture availability which modify H, E and $R_{L\uparrow}$ due to atmospheric processes and to the sub-surface thermal and hydrological processesl or alternatively (ii) because of variations in the sub-surface fluxes G_0, M_0 and surface run-off Y_0 again due to the sub-surface thermal and hydrological processes.

Note that the surface temperature and moisture availability included under (b)(i) are the surface manifestations of the sub-surface processes and are not independent of them; however in some circumstances they are prescribed surface parameters - e.g. surface temperature for underlying media such as oceans which have large heat capacities and surface moisture availability in some GCMs including the GLAS model described by Randall at this Workshop. As discussed by Randall, this surface moisture availability (β) can be used in different ways. It was originally used by Manabe (1969) as the ratio of the actual evapotranspiration E to the potential evapotranspiration E_p ($E = \beta E_p$), with E_p calculated for a moist surface at the modelled surface temperature T_*. However when β becomes small, the consequent reduction in E allows T_* to become much larger than it would be for a moist surface. Mintz (personal communication) has suggested that the consequent unrealistic values of E_p should be avoided by using a separate T_* calculated as for a saturated surface. This procedure seems more consistent with observations such as those analysed by Priestley and Taylor (1972). Randall has reported that the change has substantial effects on the GLAS model simulation over Africa though, as β was prescribed in these experiments, the importance of the change requires confirmation with interactive β. In the sensitivity experiments discussed here β has been used as originally proposed by Manabe.

2. RELATIVE IMPORTANCE OF LAND SURFACE PARAMETERS AND PROCESSES

In this section the likely importance to the heat and moisture budgets of variations in surface parameters and processes will be considered as an introduction to the sensitivity experiments discussed in later sections.

2.1 Surface parameters

2.1.1 Surface solar albedo (α)

The surface albedo is of major importance in determining the absorption of solar energy. Large variations are possible due to vegetation. Generally albedo decreases for a given vegetation type as the height of the vegetation increases, because of internal reflections, though for short sparse vegetation this may not be true if the albedo of the soil is low relative to that of the vegetation. Albedo also decreases generally for soil and vegetation as the surface wetness increases. The range of albedo for snowfree conditions is from about 0.1 for tropical forests to about 0.4 for some dry sandy surfaces. With incident mean daily solar fluxes typical of the tropics (300-400 Wm^{-2} in cloudless conditions), a variation of about 100 Wm^{-2} is possible. However this would be an upper limit to spatial variations and is only conceivable at one place with an extreme climatic change or extensive human intervention - deforestation or irrigation on highly reflective soil. Actual changes on a large scale (1000 km upwards) seem unlikely to exceed about half this magnitude or 50 Wm^{-2}. These could be due to variations in surface wetness (Idso et al 1975, Norton et al 1979) or replacement of forests by grassland or dry soil.

Much larger variations in albedo can occur due to snowcover which may have an albedo of over 0.9 for a complete new cover compared with less than 0.2 for wet soil or tundra. Mean daily solar fluxes will not usually be so large as those for the tropics cited above except for high mountains such as the Himalayas and Andes in spring and summer. In such conditions the effect on the mean daily surface heat budget could be as much as 200 Wm^{-2}; generally 100 Wm^{-2} will be an appropriate figure. Because snowcover seldom remains on trees for long, the effect of snow on the heat budget is less for forested regions. It follows that the effects on albedo of deforestation and afforestation in middle and high latitudes may be a maximum in spring before the snow melts.

2.1.2 Surface longwave emissivities ($\epsilon_{*\downarrow}, \epsilon_{*\uparrow}$)

For most natural surfaces ϵ_* is between 0.9 and 1.0. If $\epsilon_{*\downarrow}$ and $\epsilon_{*\uparrow}$ are equal, the effect of a given departure $\delta\epsilon_*$ from unity is a corresponding reduction in the net longwave flux R_{LN} of $\delta\epsilon_* R_{LN}$ which with a net flux of generally below 100 Wm^{-2} would increase the net downward surface radiation R_N by less than 10 Wm^{-2}. The real situation is more complicated because the departure from unity is not uniformly distributed through the longwave spectrum. However as $\delta\epsilon_*$ tends to be a minimum in the region around 11μm (Kondratyev et al 1982) where downward longwave flux is usually most deficient relative to upward longwave flux, $\epsilon_{*\downarrow} < \epsilon_{*\uparrow}$, the downward flux is reduced more than would otherwise be expected and the increase in R_N will be less.

2.1.3 Surface roughness length (Z_0)

An increase in surface roughness increases the surface transfer coefficients for heat, moisture and momentum. It is however difficult to predict the effects on the fluxes because the corresponding gradients decrease in consequence of the increased transfers. On average over land the sum of the latent and sensible turbulent heat fluxes will change little, being constrained to equal (R_N-G_0) (Equ. (1)). The initial effect of such a change in a model - an effect which persists in sign though declining in magnitude - is to increase the turbulent heat fluxes so reducing the surface temperature and slightly decreasing the outgoing longwave radiation. The decrease in surface temperature tends to reduce the sensible heat flux because this is driven in moist conditions by a small excess of surface temperature over air temperature which is eliminated much more easily by the surface cooling than is the corresponding moisture difference. Consequently the effects of an increase in Z_0 with moist soil and surface radiative heating are likely to be a small decrease in sensible heat flux and a larger increase in latent heat flux, an increased surface frictional stress and a reduction in near-surface gradients of temperature, moisture and wind-speed. The magnitudes of the turbulent heat flux changes, though difficult to quantify, are not likely to be much in excess of 10 Wm^{-2} on average.

2.2 Subsurface processes

2.2.1 Thermal

Analysis of monthly mean soil temperatures (e.g. from Jen Hu Chang 1958) shows that in middle and high latitudes there is a maximum downward heat flux (G_0) into the soil in early summer (May - June in the northern hemisphere) and a maximum upward heat flux in early winter. Quantitative assessment of these fluxes requires a knowledge of soil heat capacities which is not generally provided but reasonable estimates indicate that these maximum monthly mean fluxes are of order 10 Wm^{-2}. Larger values may occur in regions of central Asia with extreme seasonal temperature ranges where snow cover is relatively light and so provides little additional insulation.

In regions of marginal snowcover, the upward soil heat flux during autumn and winter may remove the snowcover and so maintain low albedos. The opposite may occur in spring when albedo is more important. Thus soil heat flux could on occasion have much larger effects than 10 Wm^{-2} but the net effect is so difficult to assess that even the sign is uncertain.

2.2.2 Hydrological

The ground hydrological processes - percolation, diffusion and subsurface and surface runoff - determine, in conjunction with the characteristics of the soil and vegetation, the surface moisture availability for given atmospheric forcing. The surface moisture availability controls the partitioning between sensible and latent heat fluxes of the 100 Wm^{-2} or more, commonly available from net radiation less soil heat flux throughout the summer hemisphere and the tropics. In moist conditions 80% or more of this quantity is transferred in the latent form so that the change from moist to arid conditions can enhance the direct heating of the atmosphere by of order 100 Wm^{-2}.

This gain may of course be compensated by latent heating if precipitation is affected by the lack of evaporation. There is ample evidence that this occurs from numerical experiments (Manabe 1975, Walker and Rowntree 1977, Charney et al 1977, Mintz 1981, Rowntree and Bolton 1982).

Some recent model results relevant to timescales from a day or two up to several months are presented later in this paper. Observational confirmation is more difficult to obtain.

However, a calculation of the effects on relative humidity of repartitioning like that specified above is instructive. If an evaporation rate of 3mm/day typical of middle latitudes in summer is reduced to 1 mm/day by limited surface moisture availability as was observed in eastern England in summer 1976 (Richards 1979) the effect on the tropospheric mean relative humidity of the repartitioning (including 58 Wm^{-2} more sensible heat flux) is about 12%/day. Clearly such a large difference could affect the precipitation from an initially moist air mass within a day or less. It is interesting that the increase in saturation vapour pressure due to the warming contributes nearly half the decrease in relative humidity. Mintz (1981) has discussed the summer moisture budget over the eastern United States to demonstrate the important role of local evaporation. He suggests that because surface evaporation moistens the boundary layer it can stimulate convective instability which leads to the condensation of moisture advected from the ocean as well as that evaporated from the land.

The present treatment of ground hydrological processes in models is quite crude as is evident from WMO(1982) (particularly the contributions by Carson, Eagleson and Dooge). The problem of representing, in terms of one or two variables, the moisture availability of a topographically complex 200 km square of the earth's surface with subgrid-scale variations in slope, aspect, elevation, soil, vegetation, and stream drainage and also precipitation is clearly enormous. A particularly crude feature of present parametrizations is the treatment of runoff, which is often assumed to occur only to limit the soil moisture below some maximum value. Also many schemes represent the soil moisture content by a single variable, so omitting variations in vertical structure important for evaporation.

3. INTRODUCTION TO SENSITIVITY EXPERIMENTS

From the discussion in Section 2 it appears that the surface parameters and processes to which the atmosphere is most likely to be sensitive are the surface albedo and the surface moisture availability. These sensitivities imply sensitivities to the specification or simulation of snowcover, soil type and vegetation and ground hydrological and thermal processes. However, the nature of present parametrizations and the limited range of sensitivity experiments available requires that we restrict the discussion to the two surface parameters seen directly by the atmosphere. For each of these we shall consider the response to global-scale variations and the response to regional anomalies. These results have been obtained from experiments by Carson and Sangster (1981), Cunnington (1980) and Rowntree and Bolton (1982) using low and medium resolution versions of the Meteorological Office 5-layer model and the Meteorological Office 11-layer model. The medium resolution 5-layer model differed from that described by Corby et al (1977) in including snow and soil moisture variables as described by Slingo (1982) and Mitchell (1983). The low resolution 5-layer model differs from the medium resolution model in that the horizontal grid-length is increased from 333 km to 500 km. The 11-layer model has been described by Saker (1975); it has a horizontal grid length of 222 km and the vertical resolution is a maximum in the boundary layer, the lowest layer being about 200 m thick.

4. THE EFFECTS OF GLOBAL SCALE SOIL MOISTURE ANOMALIES

A series of experiments has been run by Carson and Sangster (1981) to investigate the sensitivity of a model to the initial specification of soil moisture and to the treatment of surface albedos. Because of the expected long durations of the experiments an inexpensive low resolution model, the 500 km 5-layer model, was employed. The seasonal cycle was not included so keeping the analysis of the results relatively simple as was appropriate with the idealised global scale anomalies whose effects were studied. The radiation scheme was the simple one described by Corby et al (1977) based on the calculations of Rodgers (1967) with radiation dependent only on latitude, height and, for longwave radiation, temperature, the effects of zonally-averaged climatological humidities and clouds being implicit in the scheme.

The two experiments discussed in this section were run with prescribed snowfree albedos of 0.2, one from dry (zero soil moisture) conditions over all land, the other from wet conditions (15 cm soil moisture). The magnitude of the effect is evident on the Day 21-50 mean rainfall maps (Fig. 1). At this stage the results are quite similar to those obtained by Mintz and Shukla and Suarez and Arakawa with the GLAS and UCLA models respectively in experiments reported by Mintz (1981) in which the continents were prescribed dry or wet throughout. Generally,

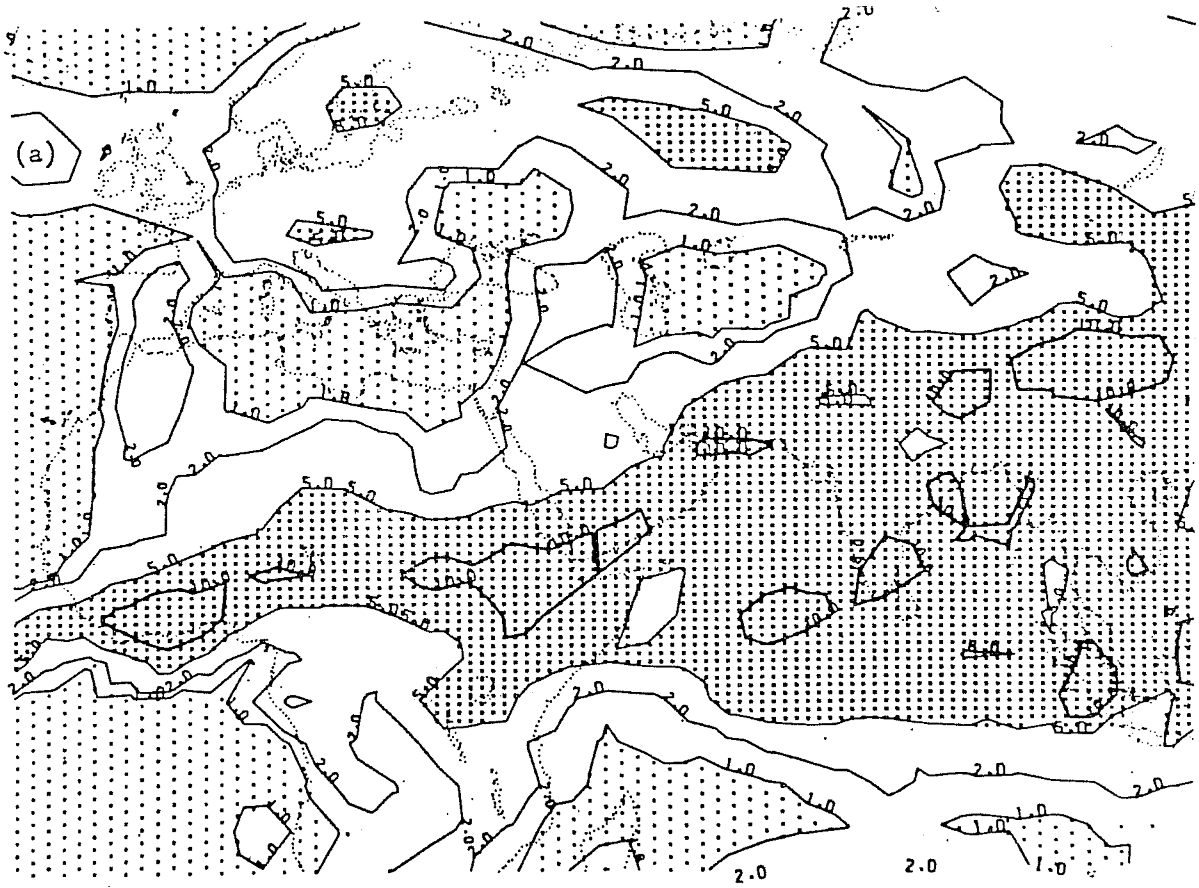


FIGURE 1 Precipitation for Days 21-50 of experiments with all land initially (a) wet (b) dry. Contours for 1, 2, 5, 10, 20 mm/day

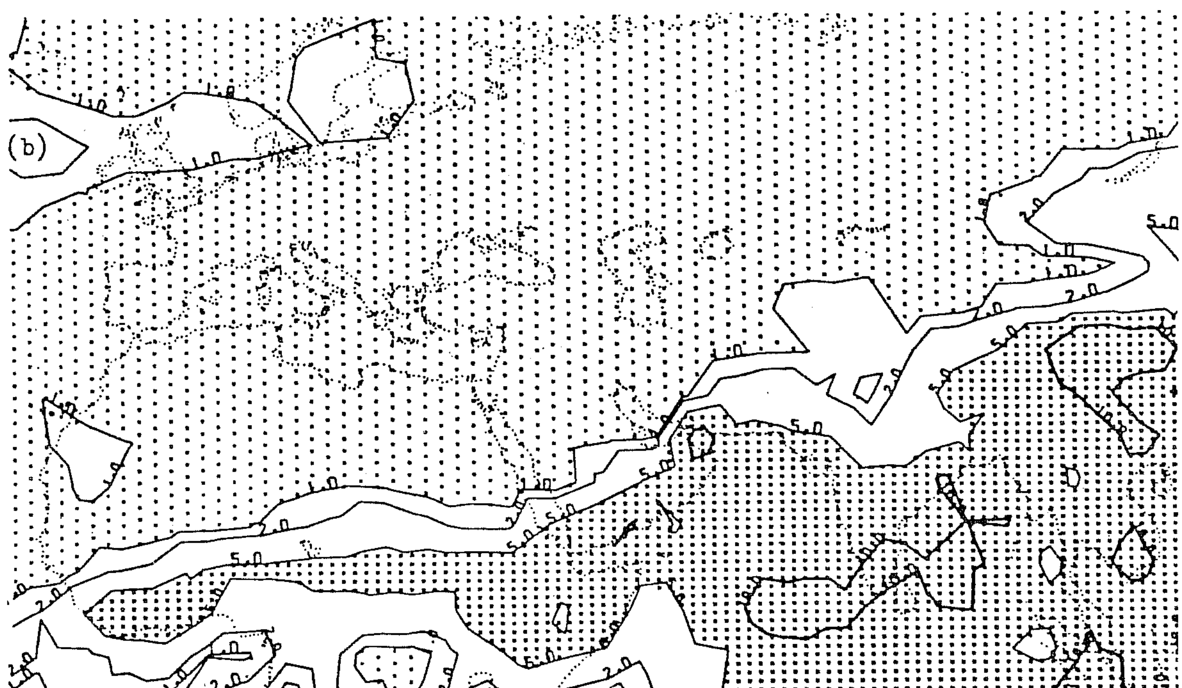


FIGURE 2(a) As Figure 1(a) for Days 171-260

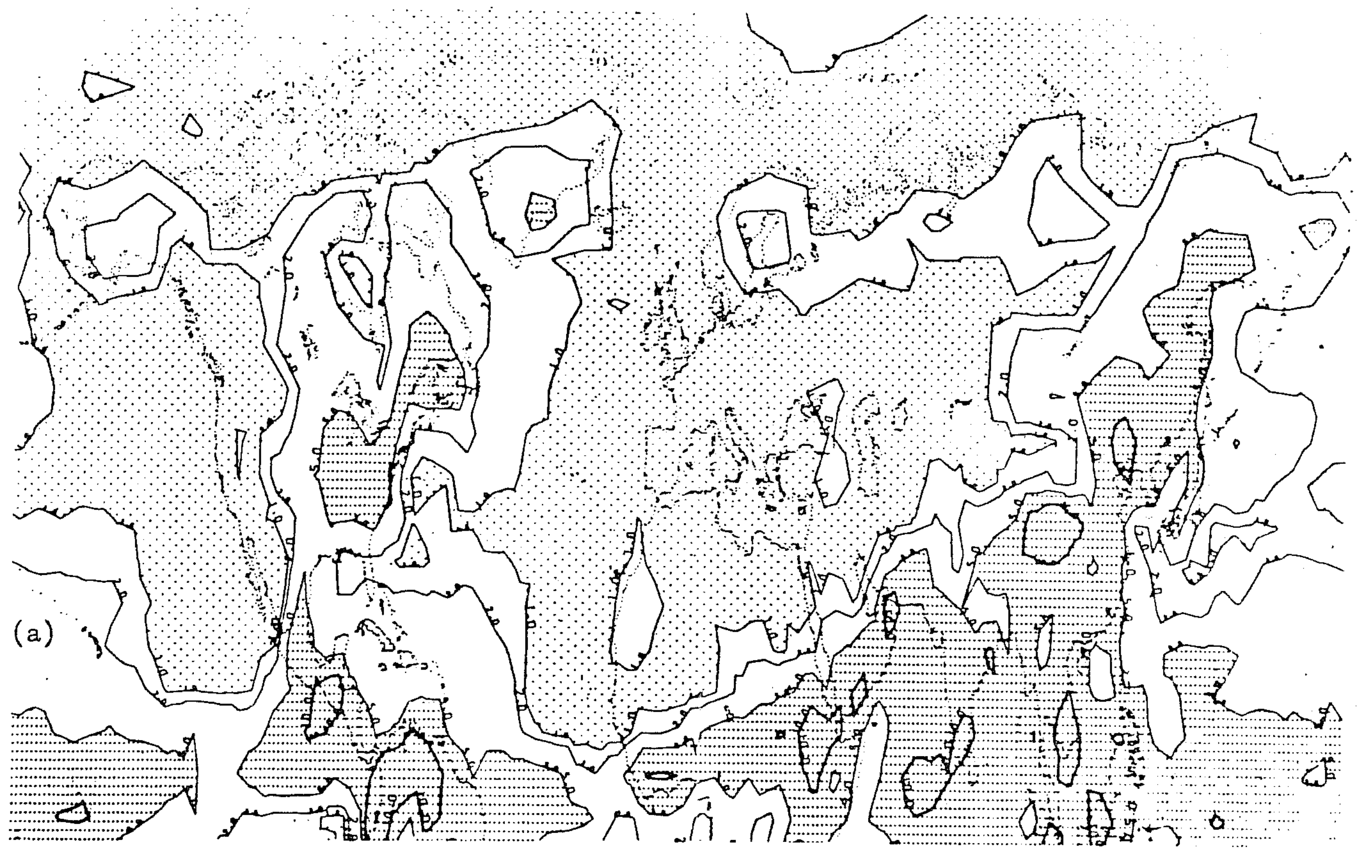


FIGURE 2(b) As Figure 1(b) for Days 171-260

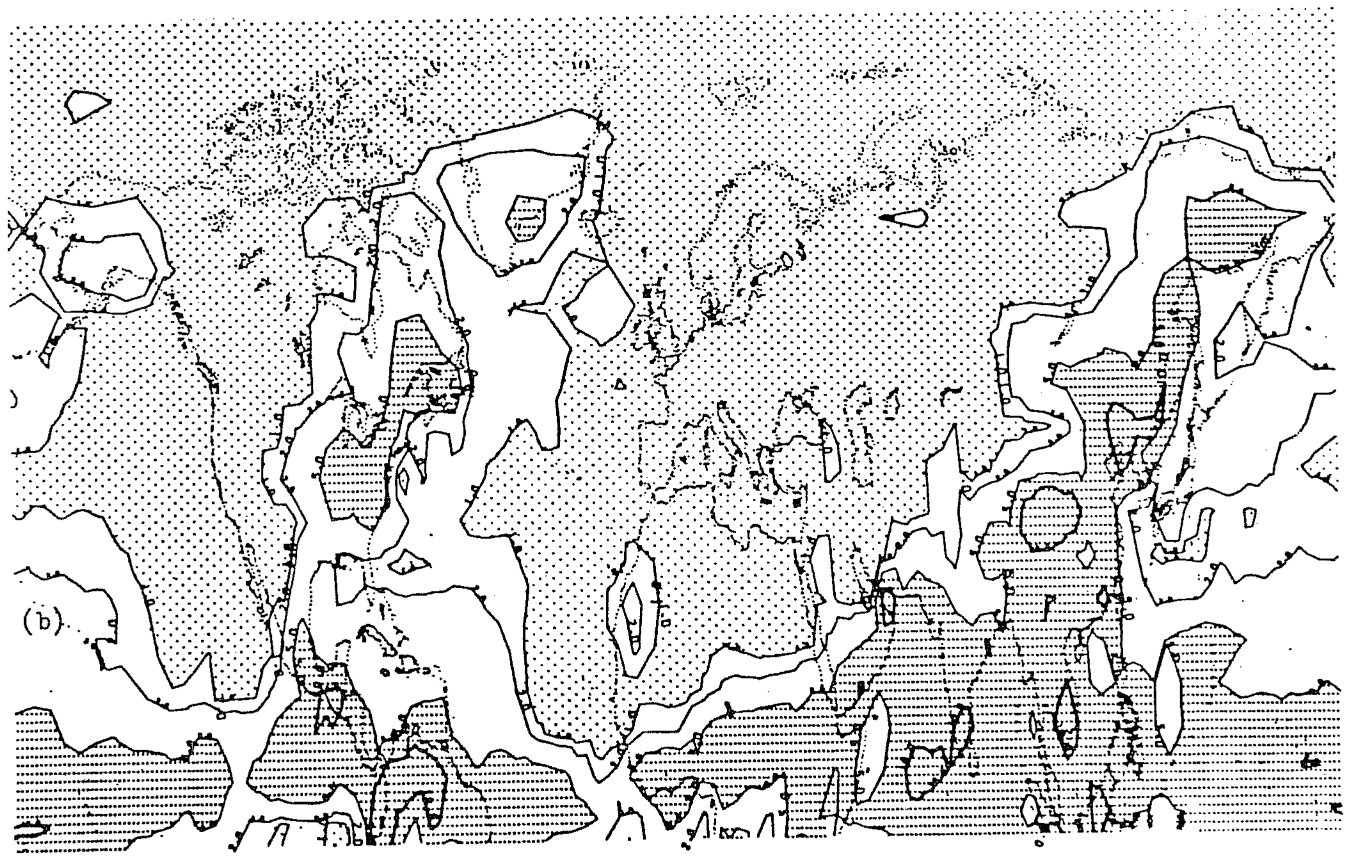
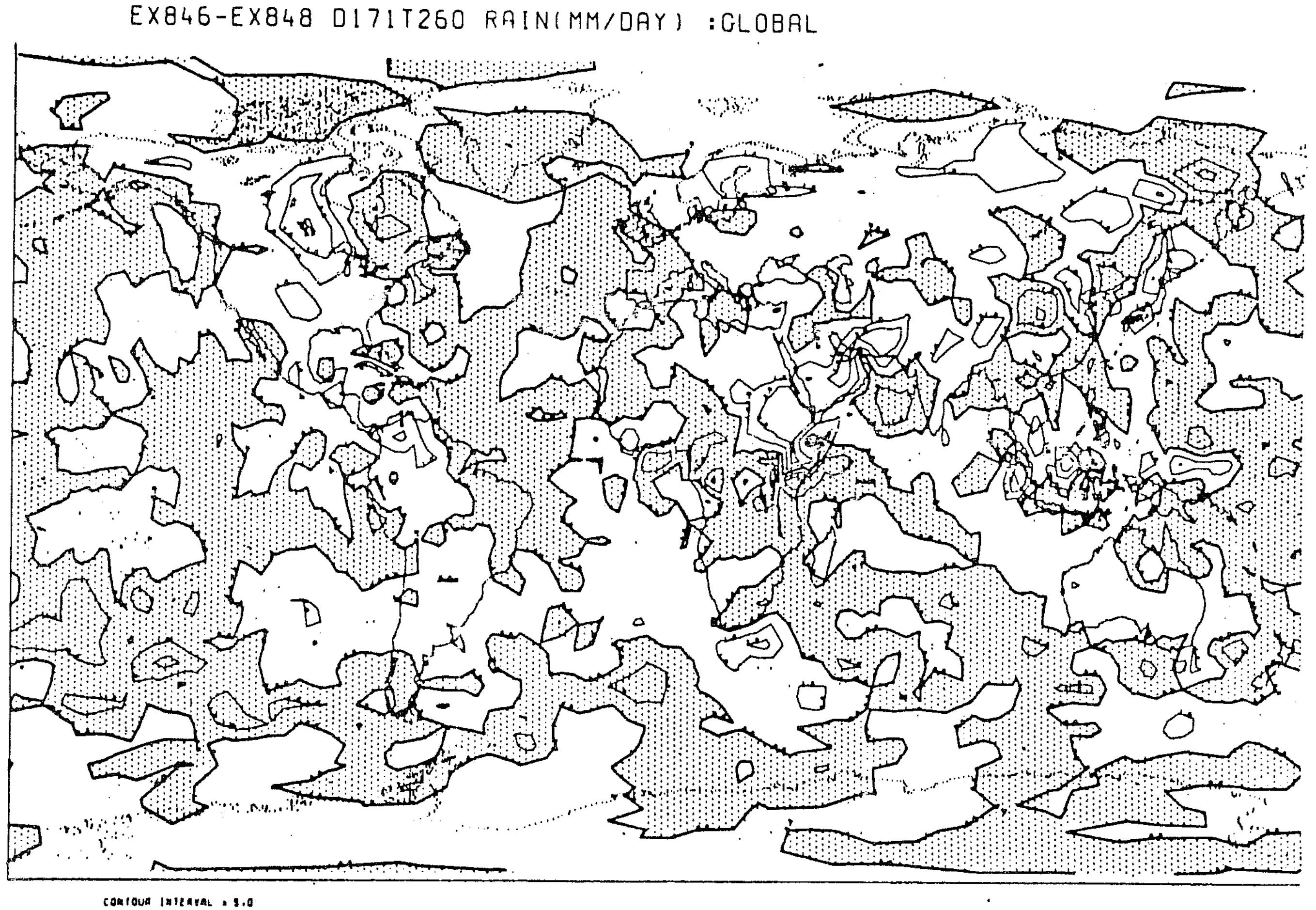


FIGURE 3 Differences in precipitation at Days 171-260 of experiment with land initially wet from that with land initially dry.

Contours for 0, ± 1, 2, 5 mm/day with areas of decrease stippled.



according to these experiments, with no evaporation over land rainfall over land in July would be confined to areas of the tropics with a nearby oceanic water vapour source such as northern South America and the southern coastal regions of West Africa and Asia. Fig. 1 shows that with wet ground the land areas with rainfall less than 1 mm/day are quite small and the 5 mm/day regions where rainfall can maintain the soil moisture against potential evaporation much more extensive. However the latter are limited to tropical Africa and the southern and eastern maritime region of Asia so it is not too surprising that eventually most of the rest of the continents become dry as shown by the Day 171-260 maps (Fig. 2). Even by this time convergence is not complete. Obviously the northern continents near 60°N are still wetter; differences also persist over northeast Africa and southwest Asia as shown by the difference patterns (Fig. 3). Most of the Sahel region had converged by about Day 100 but in the east differences still exceed 2 mm/day. These experiments demonstrate that even without the positive feedback effects of humidity changes on radiation - as well as the negative cloud feedbacks - soil moisture differences can show considerable persistence. Inclusion of seasonal variation would obviously modify the effect but not necessarily reduce it, at least in the tropics and subtropics where evaporation remains important for rainfall throughout the year.

5. EFFECTS OF CHANGES IN ALBEDO ON RAINFALL

The same model was used to investigate the sensitivity of the models to surface albedo. In order to avoid initially the complication of interactions between albedo and soil moisture two experiments were run with the soil prescribed to remain wet and albedos of 0.1 and 0.3 respectively for all land. The effect of this decrease in albedo over land is to heat the continents relative to the oceans, so enhancing the monsoon effects of land heating. Sea level pressure was reduced over the land by up to 12 mb in the low albedo case and a considerable enhancement of rainfall over land was obtained (Fig. 4) affecting almost the whole of tropical Africa and Asia.

A more realistic impression of the impact of albedo was obtained from comparison of the experiment with modelled soil moisture from initially dry conditions, already discussed, which used an albedo of 0.2 with a similar experiment using an albedo of 0.3. The rainfall differences (Fig. 5) are of generally similar character to those with wet land but there is an important difference over west Africa where more rain is obtained with the higher albedo over a large area west of Lake Chad. Examination of vertical motions (Fig. 6) shows that ascent is much stronger over east Africa with the lower albedo but little changed over west Africa. Changes in sea level pressure show maximum rises with the higher albedo over the central Sahara and maps of the near-surface flow show that the relatively

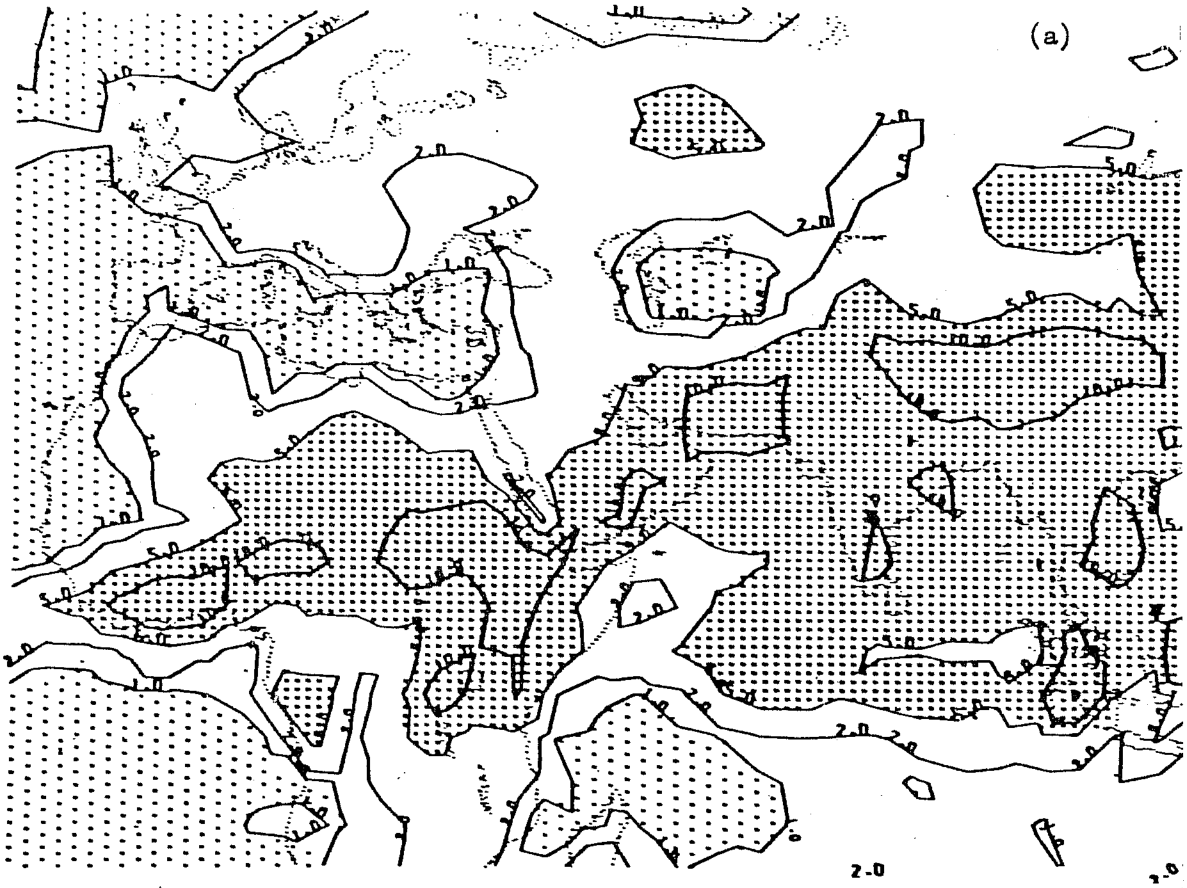
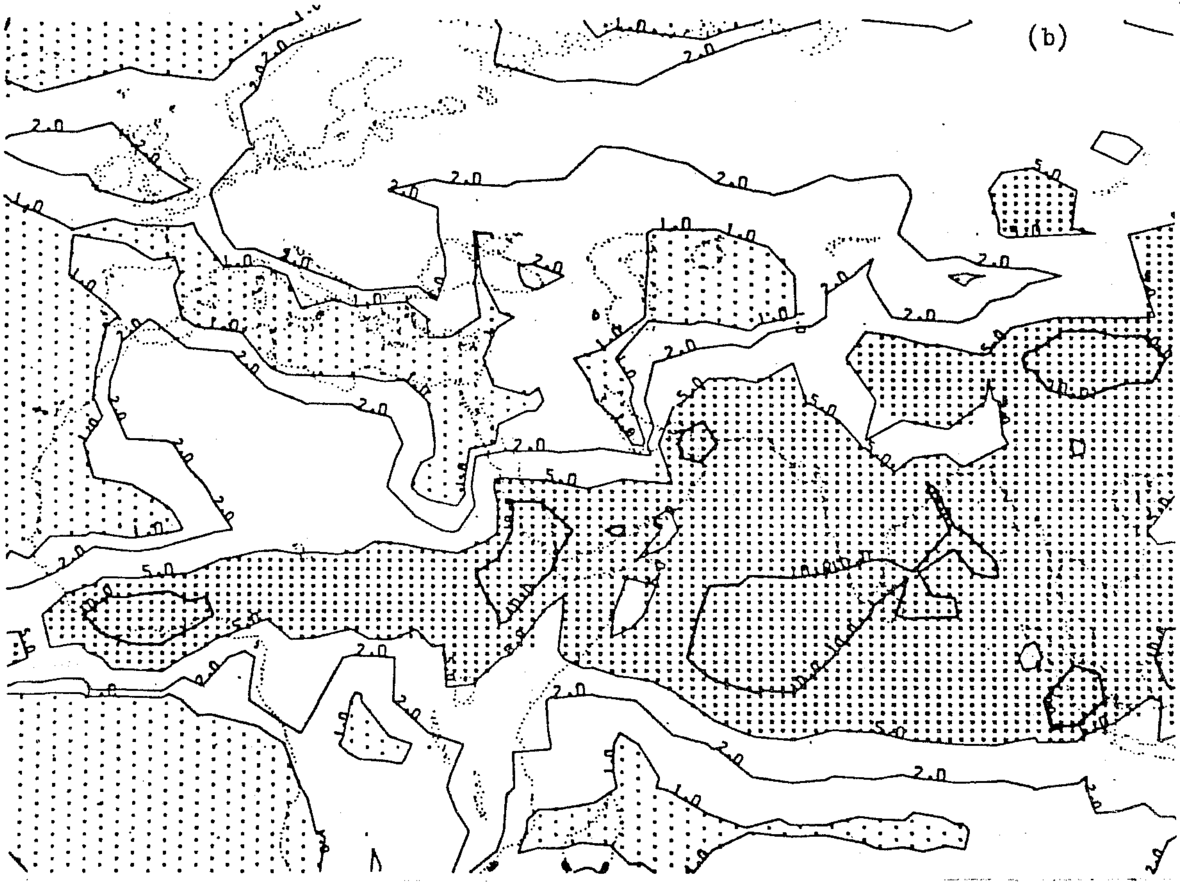


FIGURE 4 Precipitation for Days 21-110 of experiments with wet land and albedo
(a) 0.1 (b) 0.3 Contours for 1, 2, 5, 10, 20 mm/day



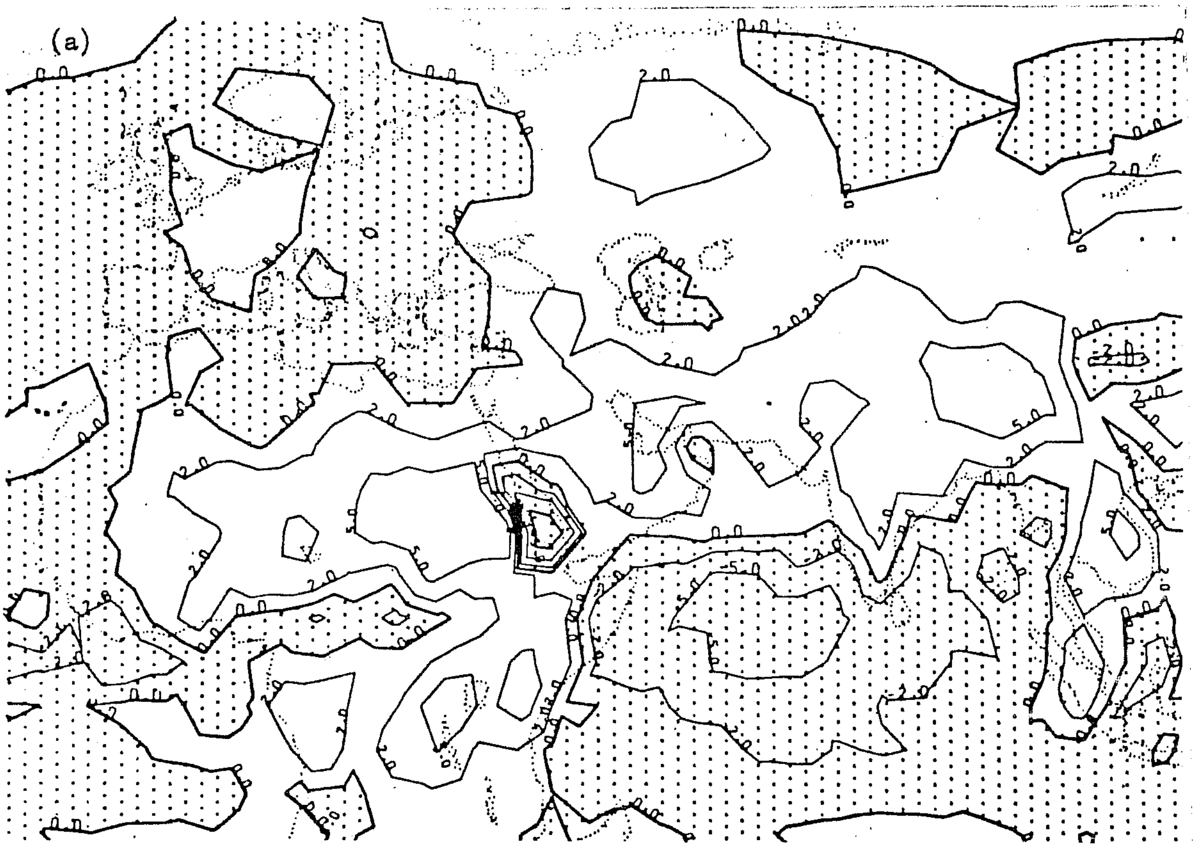
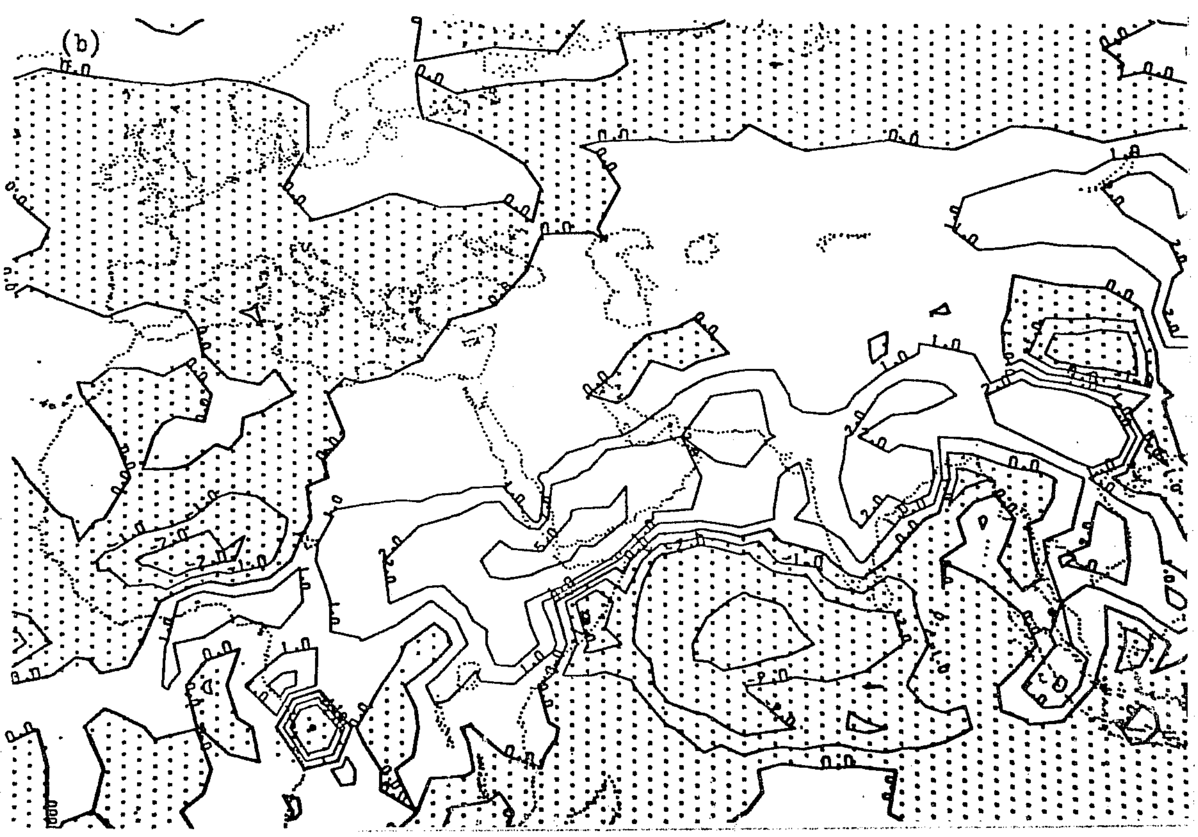


FIGURE 5 (a) Excess of precipitation with albedo of 0.1 relative to that with albedo of 0.3 at Days 21-110 (soil moisture fixed at 15 cm)
(b) Excess of precipitation with albedo of 0.2 relative to that with albedo of 0.3 (interactive initially zero soil moisture).

Contours for 0, ± 1, 2, 5 mm/day.



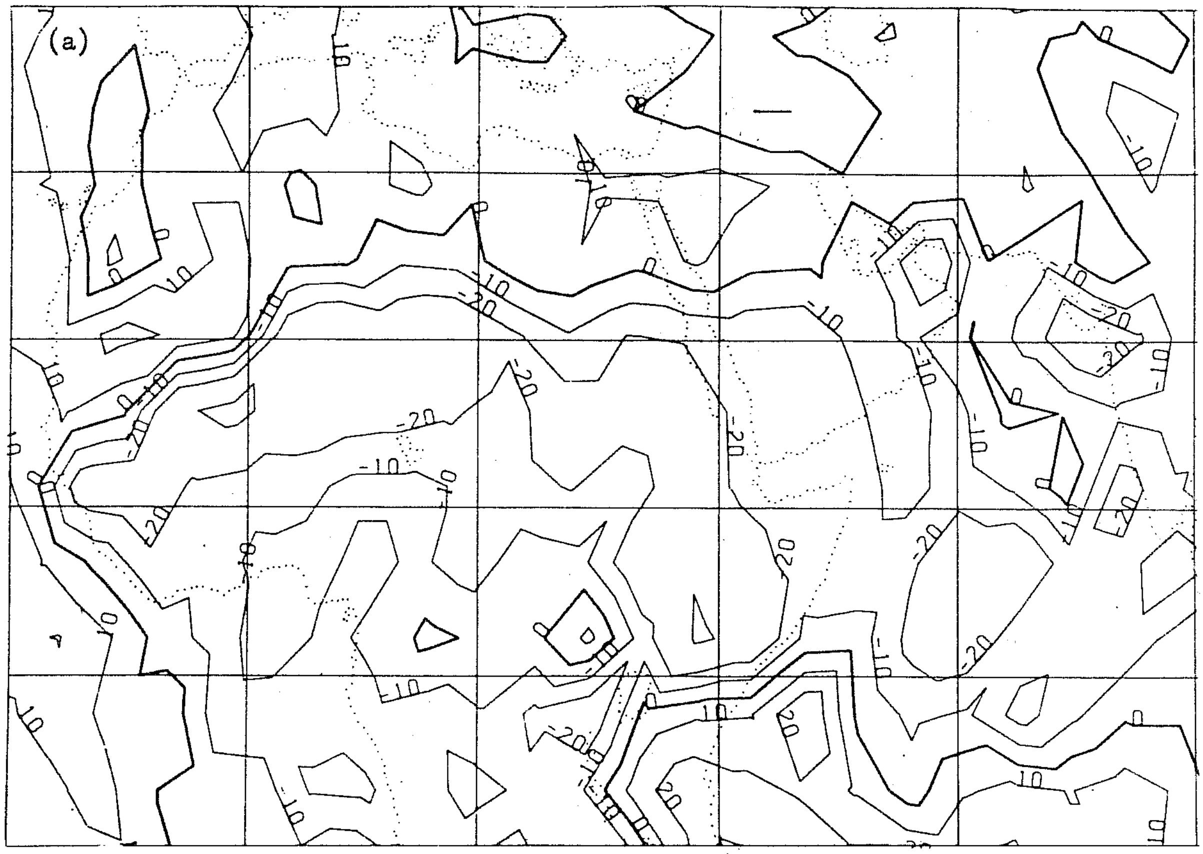
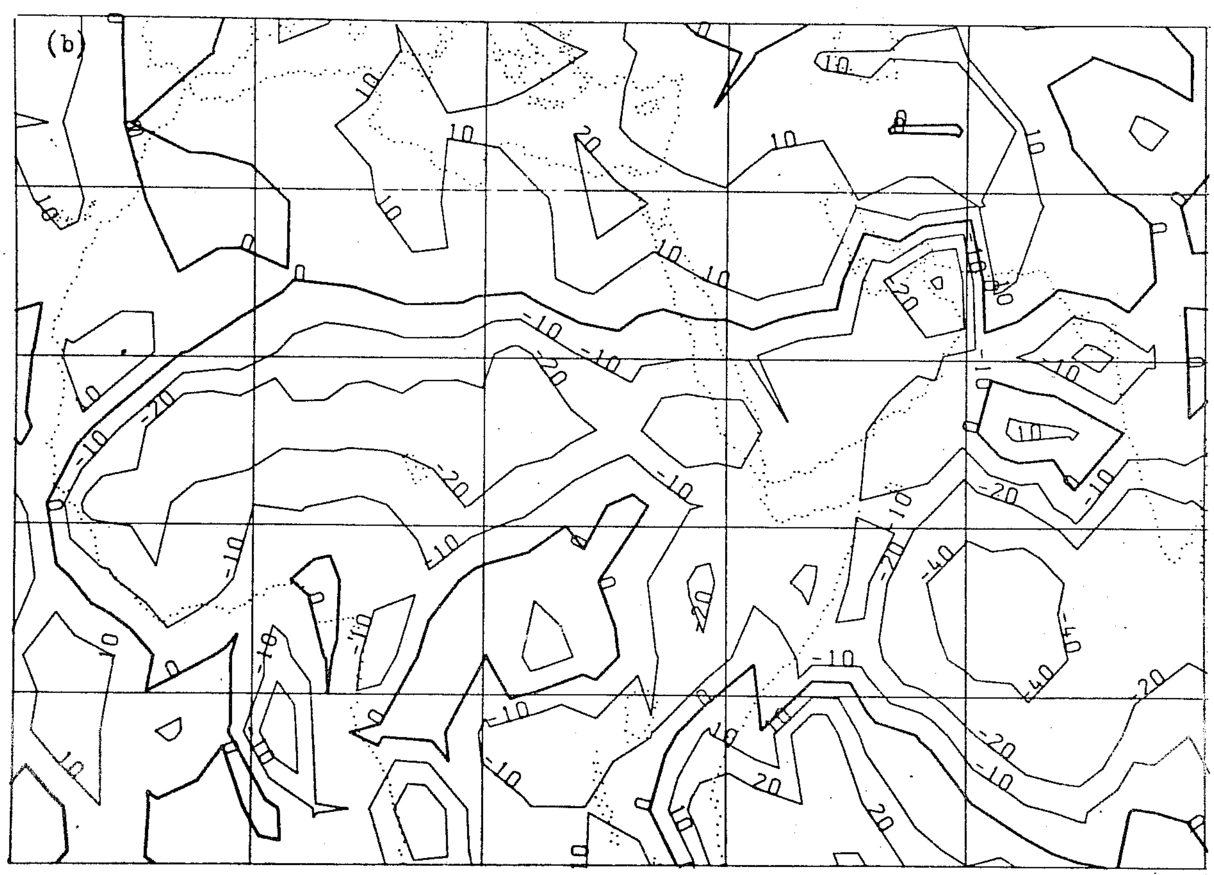


FIGURE 6 Vertical velocity (0.1 mb/hr) at $\sigma = 0.5$ for experiments with
(a) albedo of 0.2
(b) albedo of 0.3



dry northerly winds are much weaker over West Africa. These experiments demonstrate the possible complexity of the response to large-scale changes in albedo.

6. THE EFFECTS OF DEPENDENCE OF ALBEDO ON SOIL MOISTURE

Analysis of albedo observations (e.g. Idso et al 1975, Norton et al 1979) indicates that surface albedo depends strongly on the wetness of soil, the wetness of vegetation and the 'colour' of vegetation, with variations by up to a factor of two from, say, 0.15 for wet to 0.3 for dry soil. Albedo in dry conditions can also depend on the extent to which the ground has been subject to over-grazing (Otterman 1981). Experiments were made incorporating a dependence of albedo on modelled soil moisture: $\alpha = 0.3 - 0.15\left(\frac{SM}{10\ cm}\right) \geqslant 0.15$.

One might expect that with such a formulation the lowering of the albedo in wet regions and its raising in dry regions would tend to enhance the contrast between dry and wet regions, relative to the experiments with a fixed albedo of 0.2.

The rainfall maps (Fig. 7) for Days 21-50 of the two experiments starting from dry initial soil conditions confirm this expectation. The contrasts between southern India and northwest India/Pakistan and between tropical Africa and the Sahel are indeed sharpened. Note also the large change over the coastal belt of Arabia, which is much better simulated with interactive albedo. Another interesting feature is the slight increase in rainfall over southwestern North Africa reminiscent of the contrary differences found in the previous experiments with albedo increased from 0.2 to 0.3. This difference develops rapidly and significant (> 1 mm/day) rainfall spreads northward to about 25°N at Days 51-80 - compared to near 15°N with the albedo of 0.2. This may be explained by the initial increase of rain being due to the albedo increase of most of the rest of Africa and Asia favouring increased rainfall over western North Africa as discussed above and then the increased soil moisture and decreased albedo favouring a local enhancement of convergence. The difference patterns at Days 111-200 (Fig. 8) show that the enhanced contrast over India/Pakistan and east Africa/Arabia has persisted and is also present in experiments with wet initial conditions. The increased rainfall over West Africa is a feature of both the initially dry and initially wet experiments but the location of the changes is somewhat different. Indeed in the tropics and subtropics the results of Carson and Sangster suggest that the initially dry and wet cases are converging more slowly than with fixed albedos. These experiments demonstrate the possible importance for rainfall of the dependence of albedo on soil moisture in the tropics. Such impacts may be enhanced by the conversion of the forests, for which seasonal albedo variations are relatively small, to arable land for which the variations can be large, due to alternations between the vegetated and ploughed condition and between wet and dry soil.

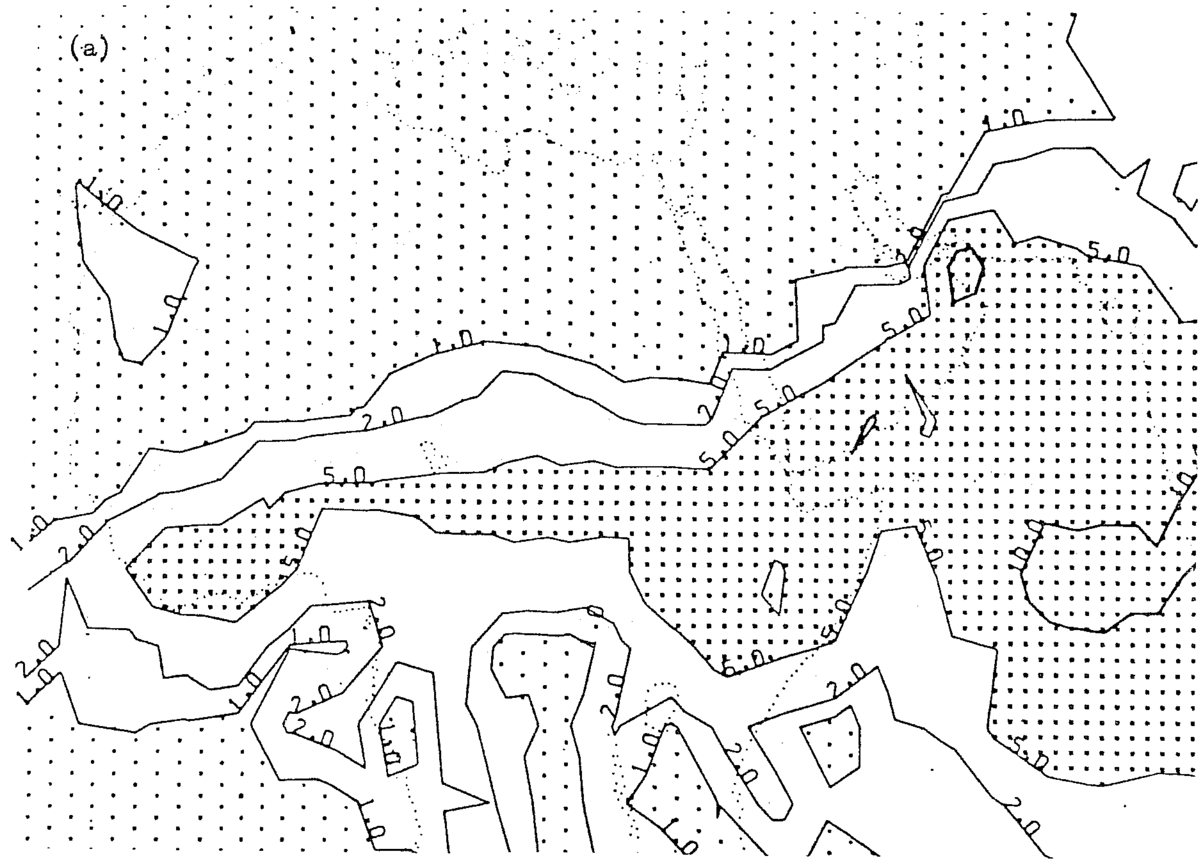
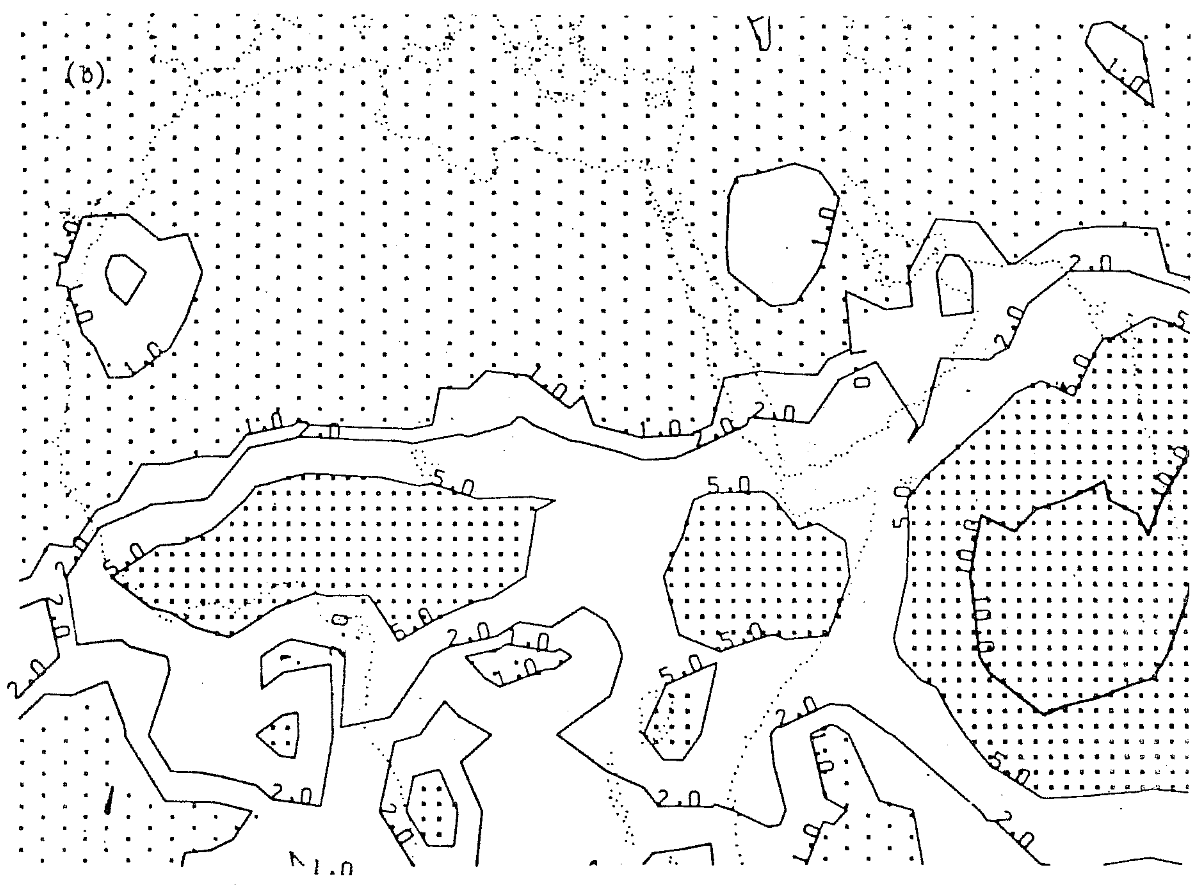


FIGURE 7 . Precipitation for Days 21-50 of experiments with initially dry soil with albedo (a) of 0.2 (b) dependent on soil moisture.



(a)

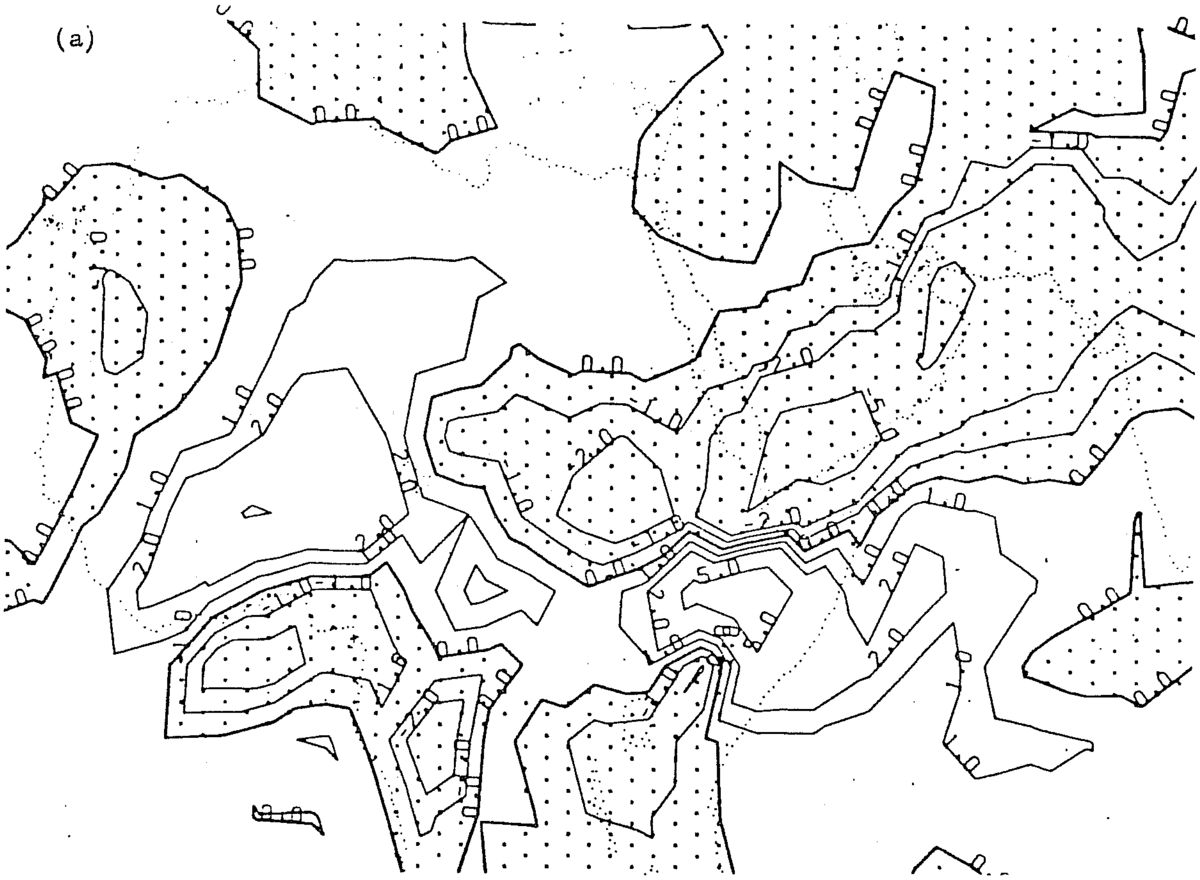
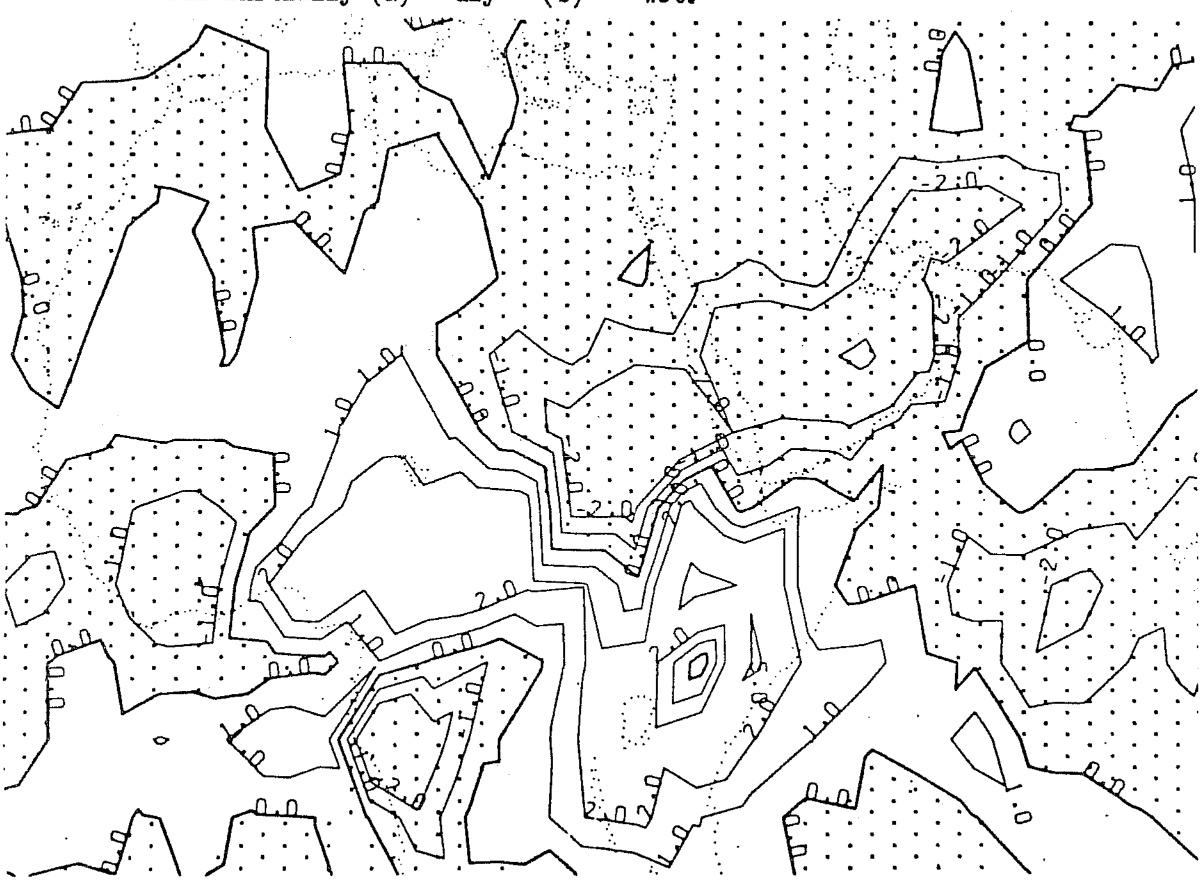


FIGURE 8 Differences in precipitation at Days 111-200 from an experiment with an albedo of 0.2 of experiments with albedo dependent on soil moisture with soil initially (a) dry (b) wet.



7. EXPERIMENTS WITH SAHARAN MOISTURE ANOMALIES

A series of experiments has been run by Cunnington (1980) to explore the sensitivity of the simulation of the Saharan region in the 11-layer model to variations in the initial conditions. The best simulations were obtained when the initial conditions were specified as realistically dry soil and atmosphere over the area of Africa from 10^{o}N - 32^{o}N from the Atlantic to the Red Sea (15^{o}W - 35^{o}E). The rainfall shown in Fig. 9(b) averaged for Days 21 to 50 was mostly confined south of about 15^{o}N over Africa, in much better agreement with observations(Fig. 9(c)) than the results obtained in an earlier experiment (Fig. 9(a)) in which the initial atmospheric humidities, derived from an integration with a moist surface over North Africa, were unrealistically wet. The effects of changing the soil moisture over North Africa were very striking in the rainfall field as shown in Fig. 10. Rainfall with the wet anomaly exceeded 2 mm/day over much of North Africa with some areas of over 5 mm/day north to 25^{o}N and in 30^{o} - 35^{o}N.

The changes were not confined to the moisture and rainfall fields. The flow over West Africa at 700 mb was a realistic easterly in the dry case (Fig. 11(a)) but with the wet anomaly (Fig. 11(b)) was quite different with weak westerlies around a Saharan low - instead of the anticyclonic flow of the dry case. There are similar differences at 850 mb though the flow is westerly in both cases. Newell (1982) has reported a difference in the same sense in 850 mb zonal winds at 15^{o}N from 15^{o}W to 15^{o}E between a relatively wet period (1958-1962) and a dry period (1970-1973) in the Sahel with easterlies weaker by 1.5 to 3.5 ms^{-1} in the wetter summers.

Although the wet soil anomaly experiment was only taken to Day 16, the experiment in which the atmosphere was initially wet and which was taken to 50 days can be compared with the dry case. The time series in Fig. 12 show that with the larger initial humidities, the atmosphere does dry initially but after about 10 days the water content starts to increase and after 50 days the difference between the two experiments is about as large as at the start. The soil moisture in the wet case has however increased steadily to reach 2.5 cm by Day 50; the total moisture in the dry case has increased by 0.4 g/cm^2, in the wet case by 2.9 g/cm^2, indicating a convergence of water vapour in the wet case. This is associated with, and probably due to changes in the heat balance which are shown for Days 7 - 16 and 21 - 50 in Table 1.

Note especially the larger sum of sensible and latent heats which, since ground heat capacity is small, must balance the net radiation. This rather large difference in the net radiation must be due to the longwave radiation, as the drying of the atmosphere will slightly increase the net downward solar radiation. There

RAINFALL 21-50 day mean

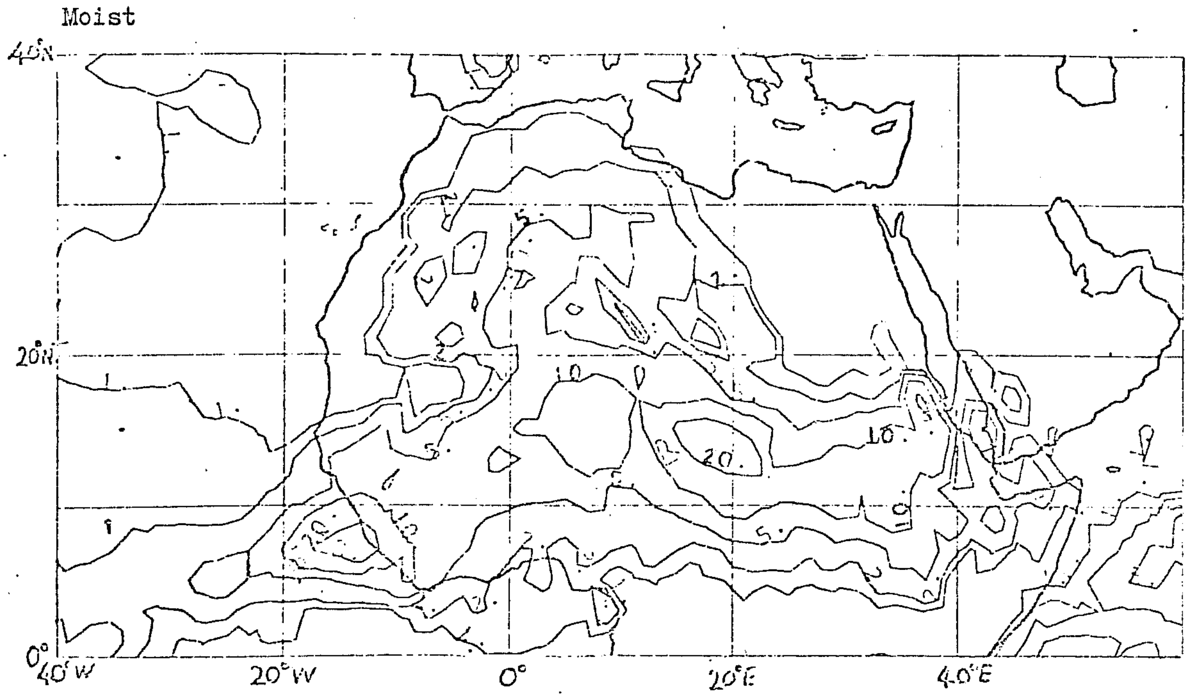
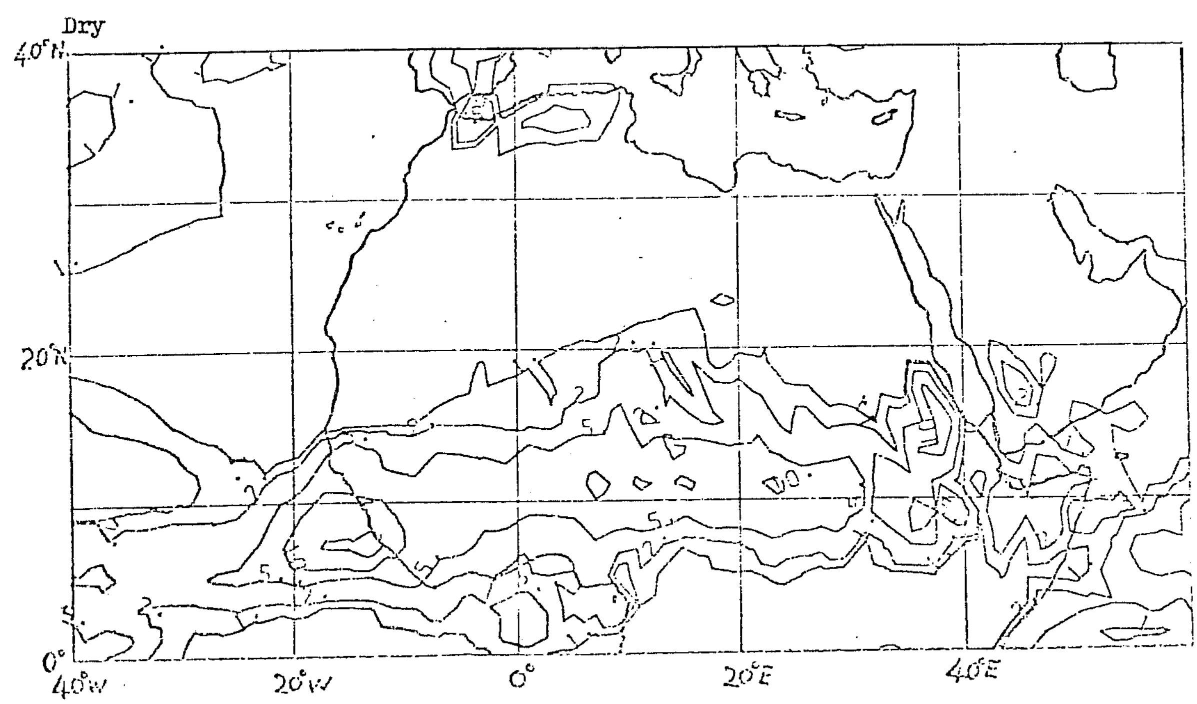


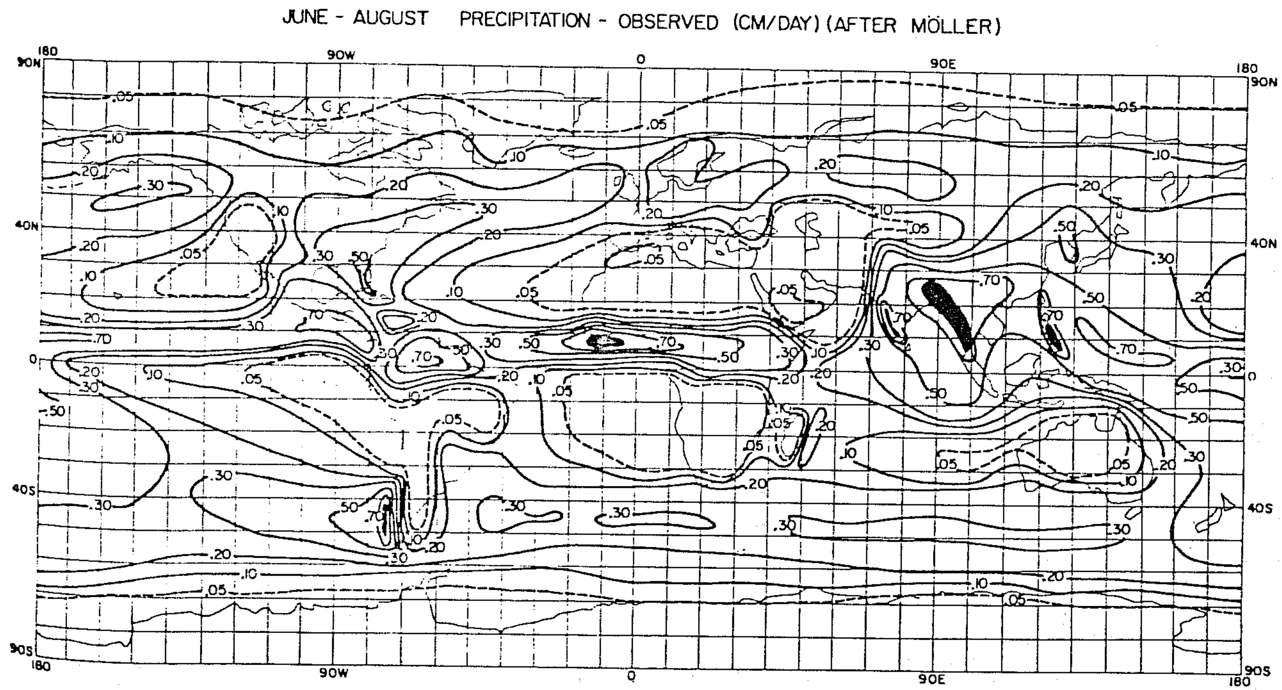
FIGURE 9 (a) Precipitation averaged for Days 21-50 of 11-layer model experiment with moist initial atmosphere over North Africa

(b) As (a) but with dry initial atmosphere over North Africa



Contours at 1,2,5,10,20 mm/day

FIGURE 9 (c) Observed June-August global precipitation (from Washington et al (1977))



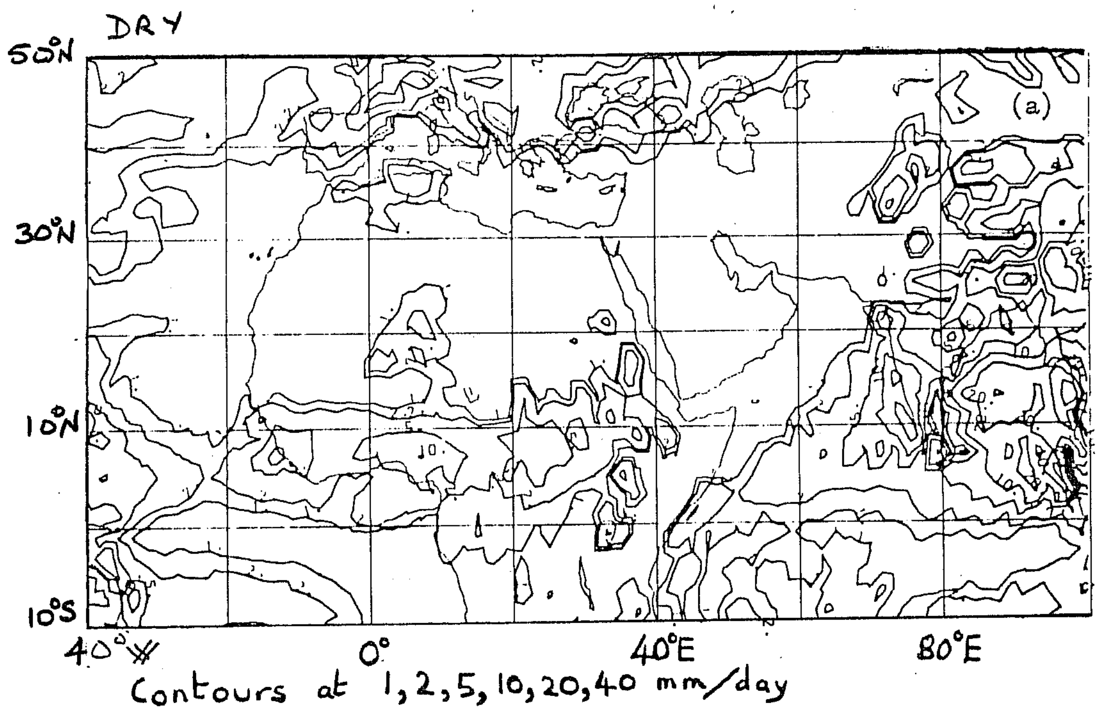
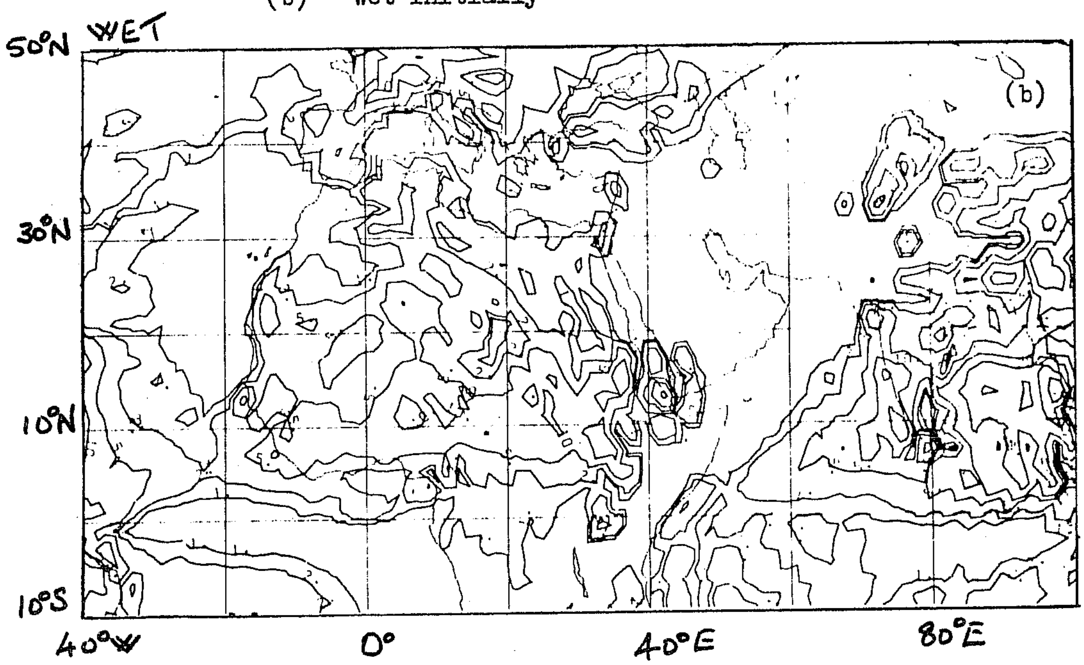


FIGURE 10 Precipitation averaged for Days 7-16 of experiments with dry initial atmosphere with soil
(a) dry
(b) wet initially



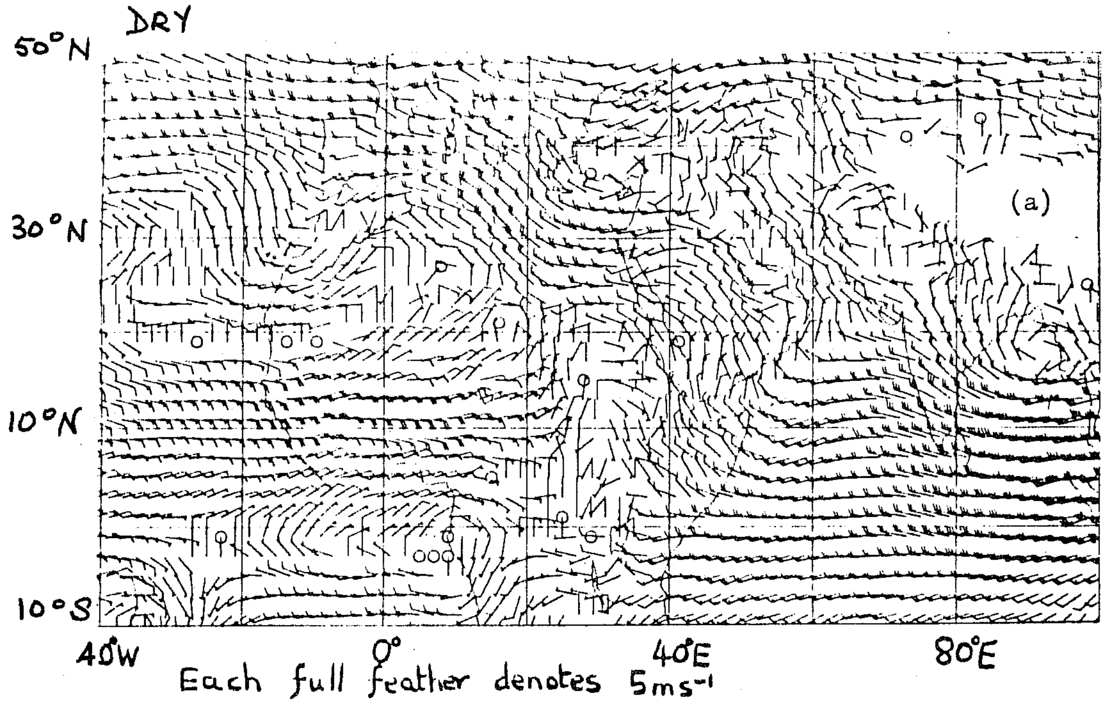
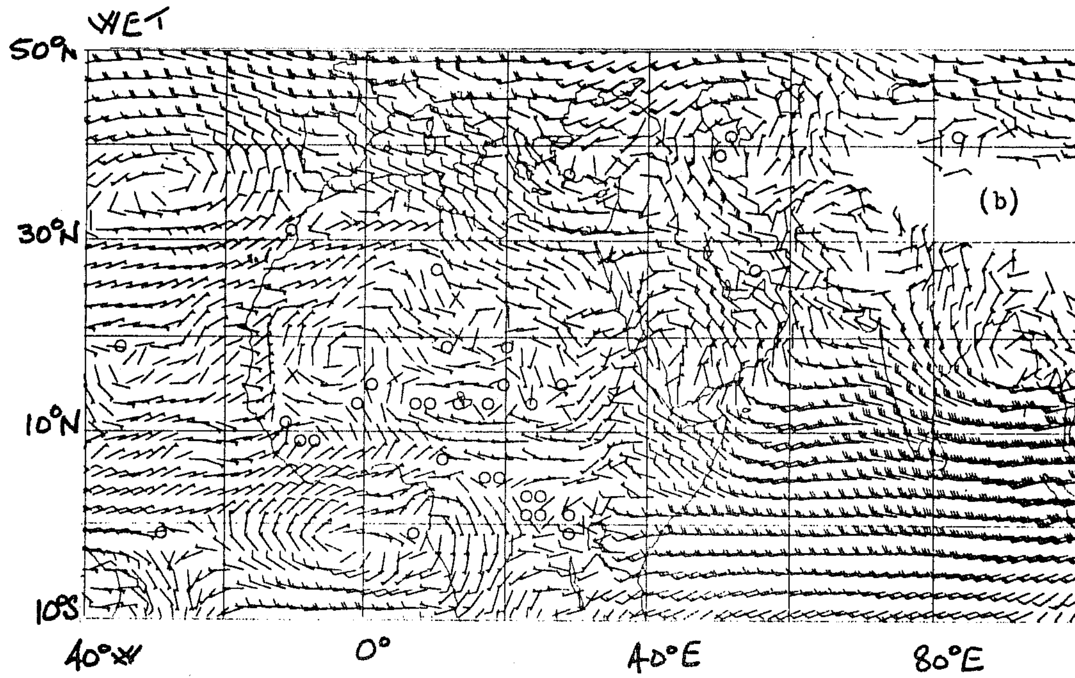


FIGURE 11 As Figure 10 but for winds at 700 mb



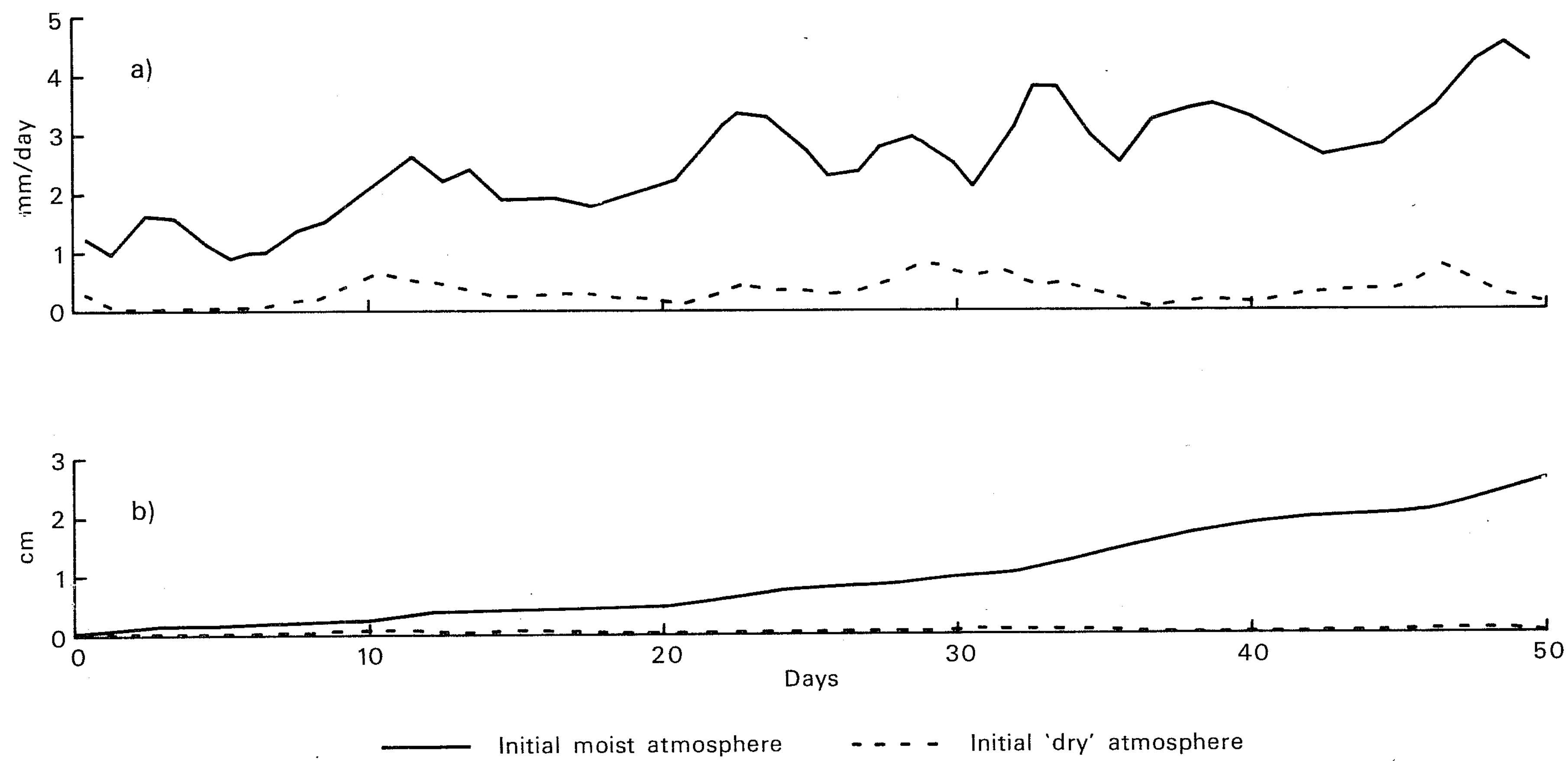


Fig. 12 Time series of a) rainfall and b) soil moisture content for experiments with initially wet and dry atmospheres. Means over N. Africa 10^{o}W-30^{o}E; 18^{o}-32^{o}N.

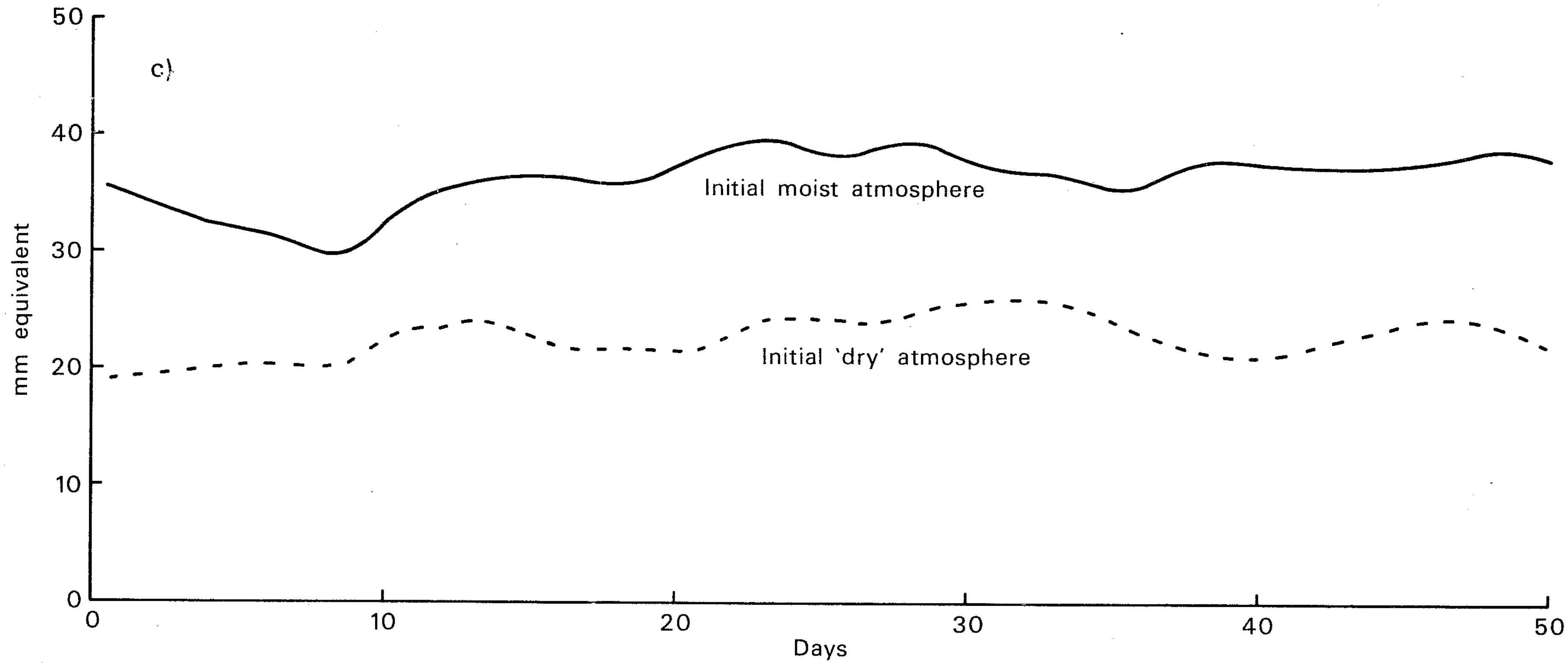


Fig. 12 (cont.) Times series of c) atmosphere water content for experiments with initially wet and dry atmospheres. Means over N. Africa 10°W-30°E; 18°-32°N.

TABLE 1

Averages for Days 7 to 16 and 21 to 50 for Sahara

	Initial conditions (10°-32°N, 15°W-35°E)				Time averaged (18°-32°N, 10°W-30°E)				
	Soil Moisture	Atmosphere	Water content Soil (g/cm^2)	Water content Air (g/cm^2)	Rain (mm/day)	Evap (mm/day)	Sensible Heat (Wm^{-2})	Net* Radiation (Wm^{-2})	Temp (σ=.987) (K)
Days 7-16	0 cm	Moist	0.32	3.38	1.92	1.63	81.8	128.9	302.0
	0 cm	Dry	0.04	2.25	0.38	0.36	99.0	109.5	303.9
	10 cm	Dry	5.93	3.76	2.90	5.12	16.0	164.0	292.3
Days 21-50	0 cm	Moist	1.52	3.84	3.12	2.35	72.9	140.7	299.9
	0 cm	Dry	0.04	2.38	0.37	0.37	101.9	112.7	304.1

* The net radiation has been estimated from the sum of sensible and latent heat fluxes, assuming the heat storage in the ground to be zero. Radiation data were not retained.

is no change in cloudiness since this is prescribed so the increased net upward longwave radiation must be due partly to an increase in temperature - over 4K at $\sigma = 0.987$ at Days 21 - 50(the surface temperature increase is similar)-and partly to reduced downward flux from a drier atmosphere.

The effects of the wetter soil at Days 7-16 (compare the second and third rows of Table 1) help to clarify the roles of the changes in temperature and moisture content. If the surface temperatures are equal to the temperature at 100 m the upward longwave flux is reduced in the wet case by 70 Wm^{-2}. Even with no reduction in net solar radiation, the downward longwave flux can only have decreased by 15 Wm^{-2}, the reduction due to the lower atmospheric temperatures being largely compensated by the considerable increase in atmospheric water content which is particularly pronounced near the ground, specific humidity of the lowest layer increasing from 6.3 to 13.0 g/kg.

It is important to note that this model, by omitting the negative cloud feedback but including the positive water vapour feedback, will have exaggerated the response of the radiative heating to the soil and atmospheric moisture changes. It is estimated from radiative calculations with this scheme in a single column model that the increase of 28 Wm^{-2} in the net downward surface radiation due to increased moisture and lower surface temperatures could be counteracted for average solar zenith angle by moderate increases in cloud cover e.g. by increasing high clouds by 14%, medium cloud by 7% and low cloud by 17½% (with random overlap). A realistic cloud parametrization scheme is needed in the model to make a proper assessment of the effects of soil moisture. It is possible that the present intransitive behaviour of this model would not be repeated if cloud feedback were realistically represented.

8. EUROPEAN SOIL MOISTURE ANOMALIES

As discussed in section 2.2.2, a soil moisture anomaly may have significant effects on the atmosphere quite quickly with relative humidity changing at 12%/day in the absence of changes in precipitation for the case discussed there. Since the surface heat and moisture flux over land is concentrated during the daytime, the timescale for this effect is of order 12 hours and the required horizontal scale for winds of order 5 ms^{-1} little over 200 km.

However an anomaly on such a small scale would be quite quickly eliminated because of the advection of well-moistened air from adjacent areas. The middle latitude anomaly, for which results are reported here, is on a larger scale, similar to that affected by the European drought of 1976 - about 2000 km from north to south and 1300 km from east to west (Fig. 13). To obtain a rough com-

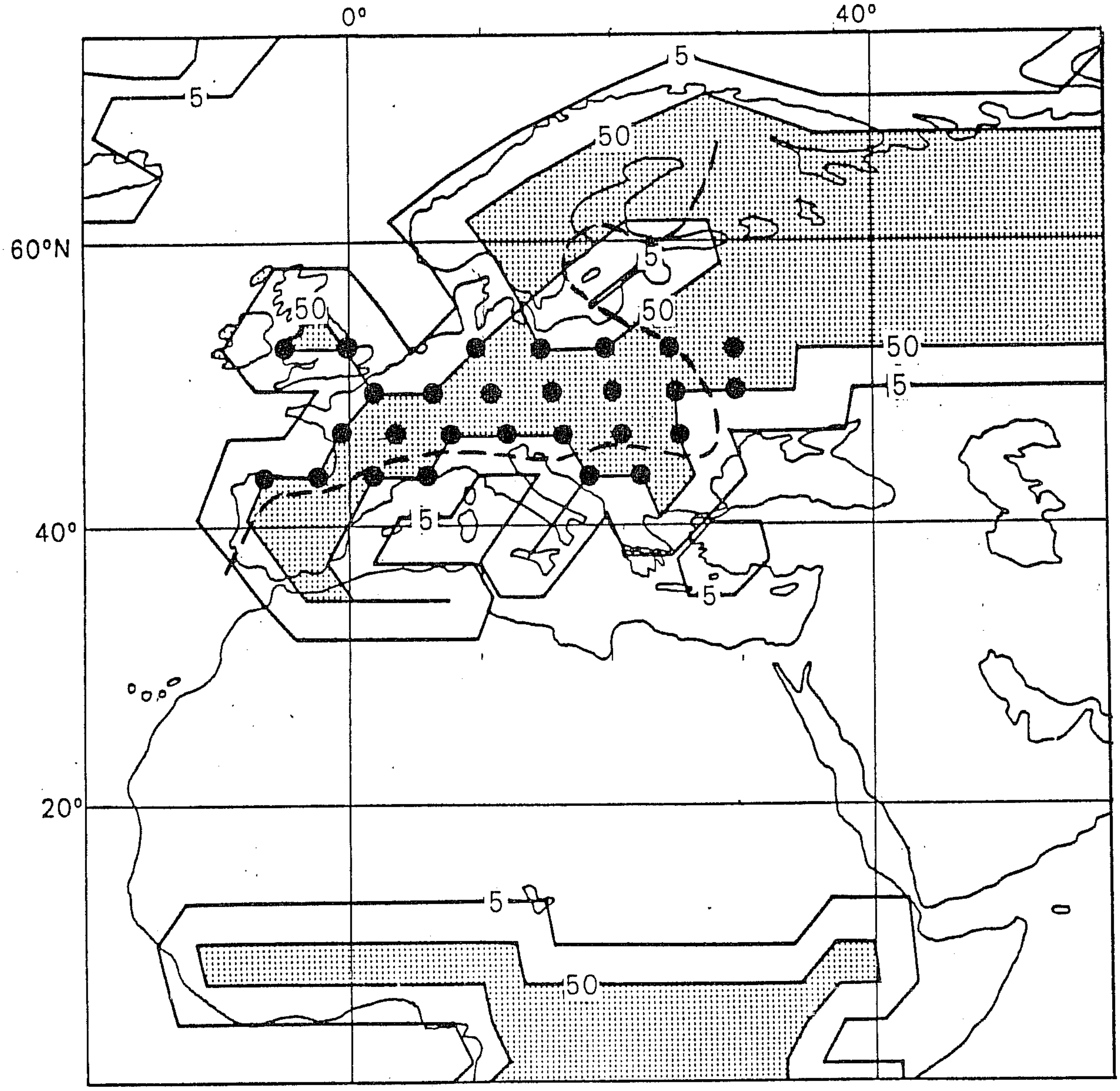


FIGURE 13 Initial soil moisture (mm) for the control experiment. The grid points at which anomalous values of initial soil moisture were applied are marked by dots. The rainfall data described in the text suggest that most of Europe north and west of the dashed line received less than 75% of normal rainfall in February to June 1976, much of it less than 50%.

parison to the horizontal scale of the 1976 drought, rainfall data were extracted from 'Witterung im Ubersee', and are shown in Fig. 13 averaged for February to June. Although the intensity of these was not as extreme as those imposed in the model, the horizontal scale is quite realistic. Experiments were run for 50 days with the 333 and 500 km mesh versions of the 5-layer model with initial soil moistures of 0 (dry), 5 (control) and 15 cm (wet) over the anomaly area. Only results from the dry and wet medium resolution experiments (MD, MW) are shown here. The results from the other experiments were consistent with these.

Fig. 14 shows the specific humidity and temperature differences at the model's bottom level ($\sigma = 0.9$, about 800m) after two days. The humidity differences exceed 2 g/kg over a large area with more than 3 g/kg over northern France. Coupled with the temperature changes, which exceed 4K over most of the anomaly area, there is a decrease in relative humidity - estimated as the ratio of the mean specific humidity to the saturation specific humidity corresponding to the mean temperature, both meaned over the anomaly area - from 90% in the wet case to 50% in the dry case. In the first two days there were slight changes in rainfall, as shown by the northern edge of an area of rain exceeding 1 mm/day centred over the Mediterranean. During the following 24 hours these small differences increase (Fig. 15), some even appearing outside the anomaly area over Spain and west of Britain (there was an easterly airstream over western Europe initially). In the dry case rainfall over the western half of the anomaly area is below 0.5 mm/day whereas in the wet case there is a large area over France and parts of neighbouring countries with over 1 mm/day associated with development of a trough from the Mediterranean. This trough is deeper in the dry case but gives little rain. These differences continue to expand, affecting southern Scandinavia by the sixth day (Fig. 16) when southerly winds advect the humidity and temperature anomalies north of the anomaly area and bring much more rain ahead of a trough in the wet than the dry case.

The Day 21 - 50 mean maps (Fig. 17) show that rainfall has remained light in the dry case whereas in the wet case much of western Europe including southern Scandinavia has rainfall in excess of that (about 3 mm/day) needed to maintain the soil moisture against potential evaporation. Even in the wet case the mean rainfall is not sufficient to balance evaporation over the whole anomaly area, and generally the model tends to dry out over Europe as demonstrated for the low resolution model in section 4. This is partly because the model's flow is less westerly than is usually observed. It is appropriate to consider what would happen if precipitation over Europe equalled or exceeded evaporation. In the annual cycle run of the model discussed by Mitchell (1983) the summer flow is generally westerly and Europe remains quite wet from the first autumn onwards.

Rowntree and Bolton (1982) therefore made a 50-day experiment with a dry anomaly initially in the area shown in Fig. 13. Fig. 18 shows the rainfall averaged for the first ten days of the control (effectively wet) and dry anomaly experiments. Generally rainfall in the dry case is between 30 and 70% of that in the wet case. Soon after Day 20 a wet spell affected much of Europe in the dry case and moistened most of the anomaly area sufficiently to bring evaporation close to the potential value. Only parts of southern France and northern Spain showed significantly less rainfall throughout the 50 days. This result confirms what one might expect - that soil moisture anomalies in a maritime region can persist only with some flow types.

9. Conclusions

An analysis of land surface processes indicates that atmospheric general circulation models should be most sensitive to the treatment of surface albedo and soil moisture availability, the first modifying the energy available for turbulent transfer into the atmosphere, the second the partitioning of this transfer between the sensible and latent forms. Because surface albedo is dependent on snow cover, surface moistness and vegetation, a further sensitivity to these features of the land surface is implied. A sensitivity of the circulation to variations in surface roughness through the momentum flux is also to be expected but is difficult to quantify in the absence of suitable numerical experiments.

Experiments with GCMs confirm these expectations with clearly defined responses to planetary scale variations of albedo and soil moisture and to regional anomalies in soil moisture. However, a lack of observational data has so far restricted experiments to rather idealized, enhanced perturbations and further experiments are needed to explore sensitivities to observed anomalies. In formulating such experiments, care is needed to avoid unrealistic restrictions or distortions of relevant interactive feedbacks.

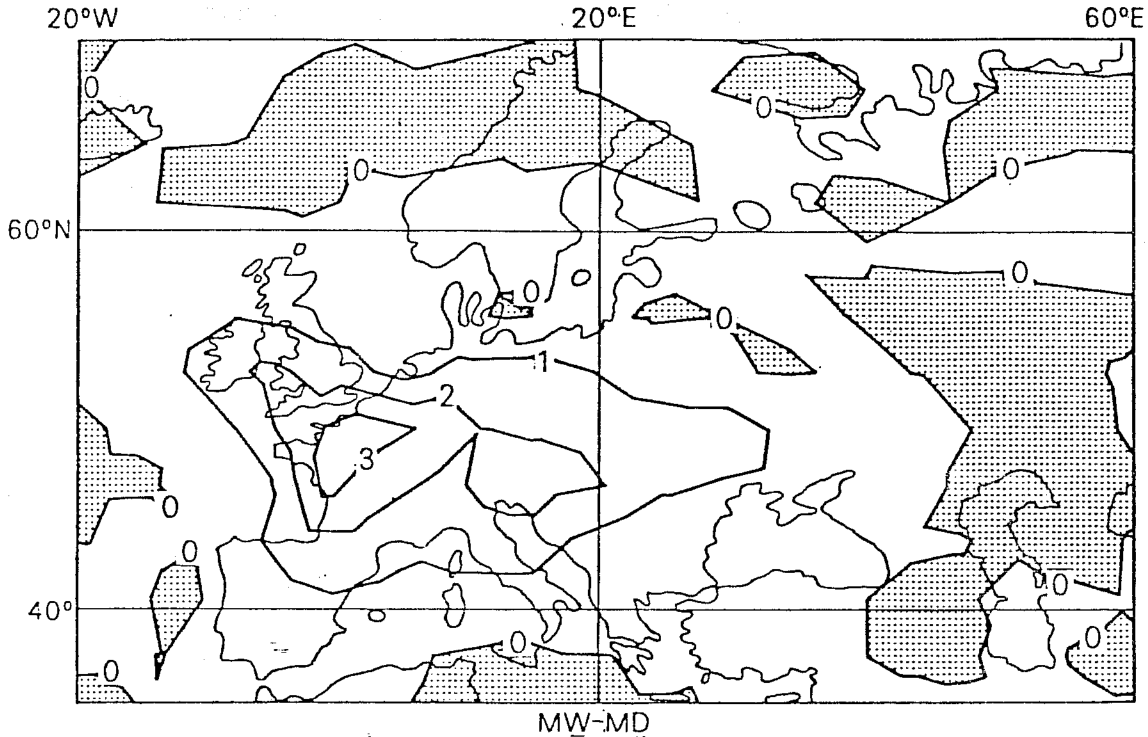
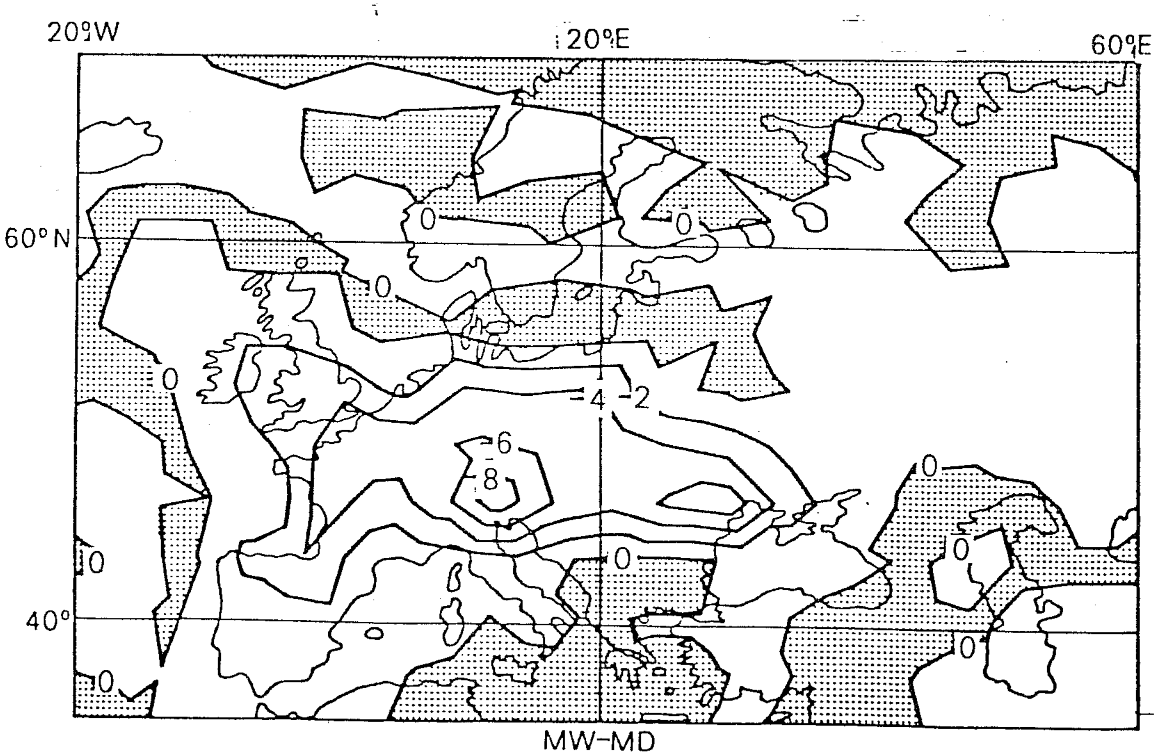


FIGURE 14 The differences between experiments (MW-MD) at Day 2 for
(a) specific humidity (g/kg), (b) temperature (K)
both at $\sigma = 0.9$.



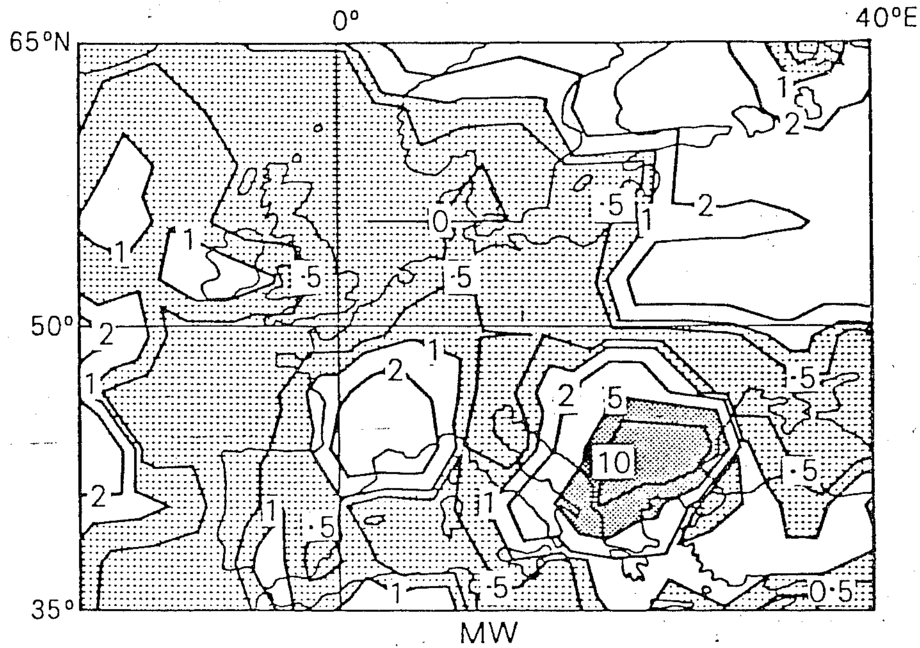
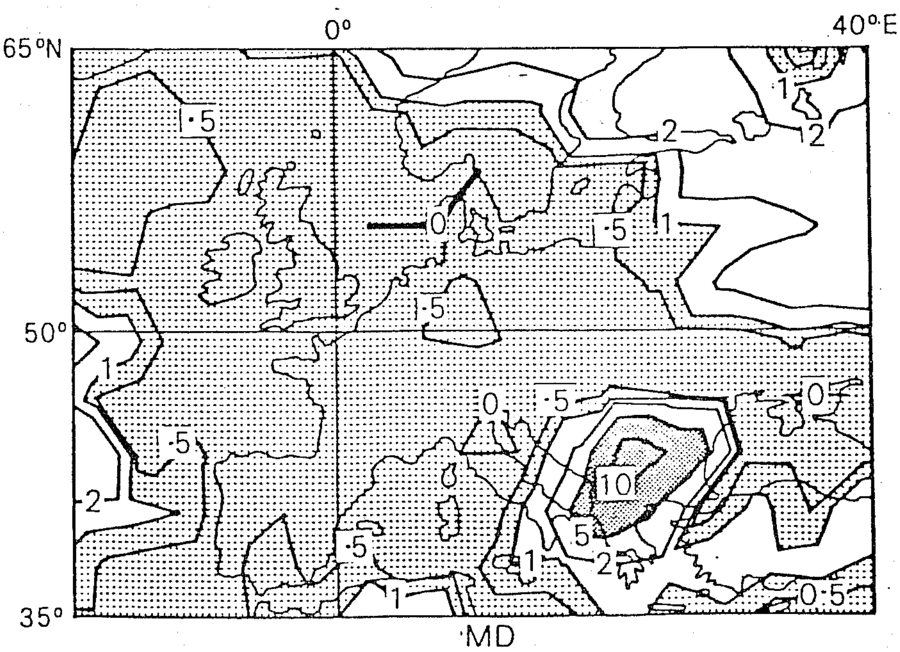


FIGURE 15 Rainfall (mm) during the third day of experiment
(a) MW, (b) MD.



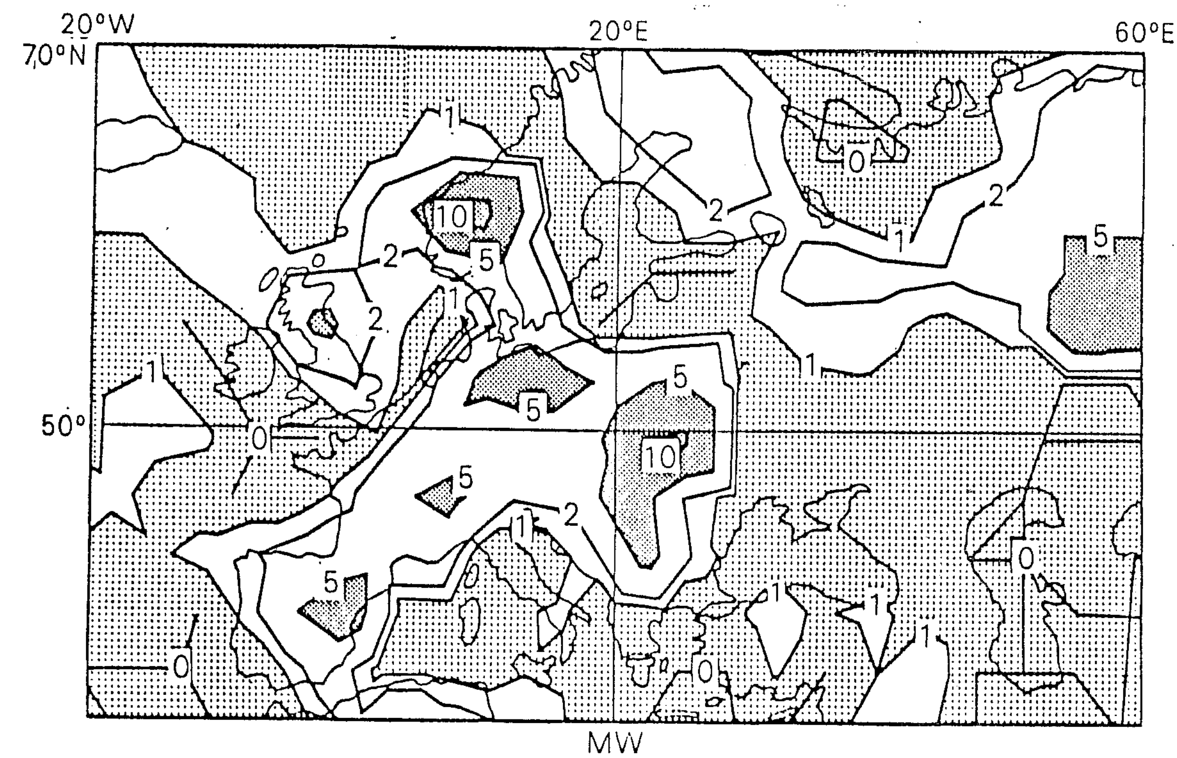
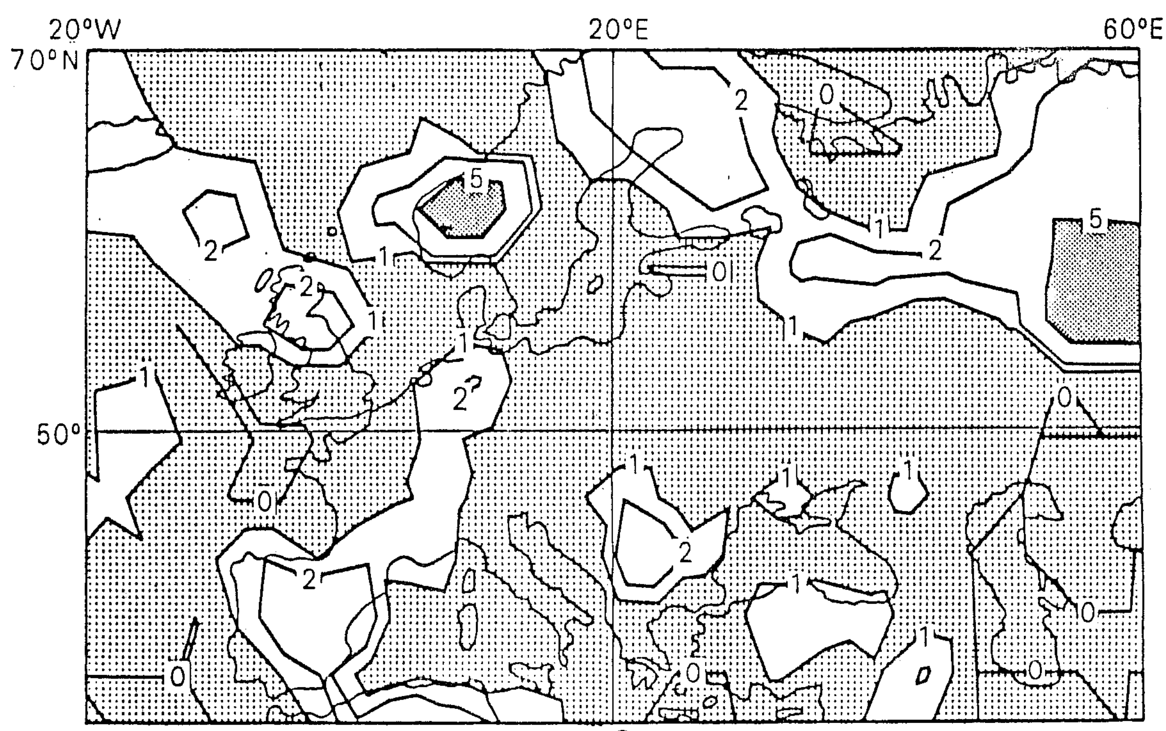


FIGURE 16 Rainfall (mm) during the sixth day of experiment
(a) MW, (b) MD.



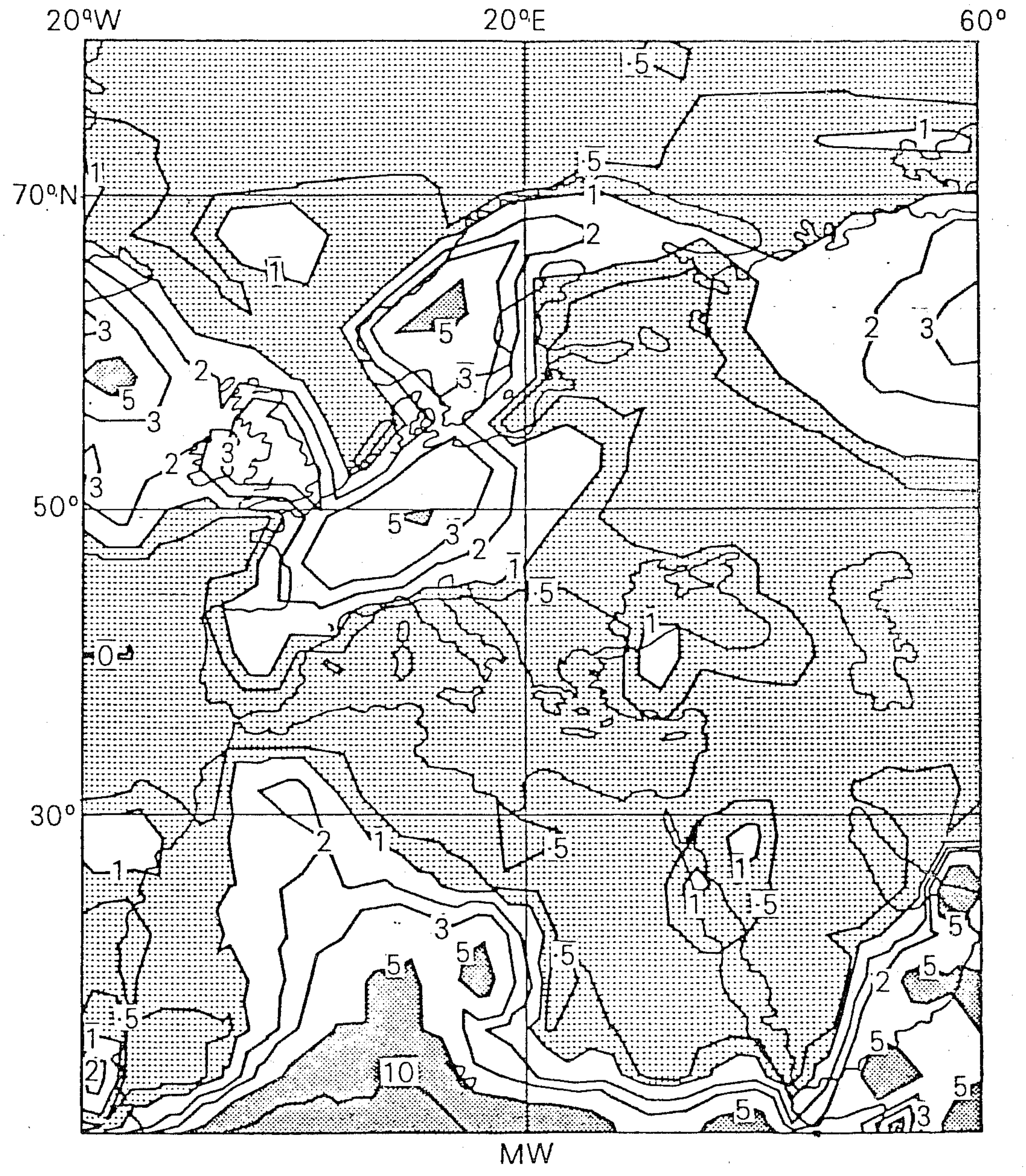


FIGURE 17 Rainfall (mm) averaged for Days 21 to 50 of experiment
(a) MW

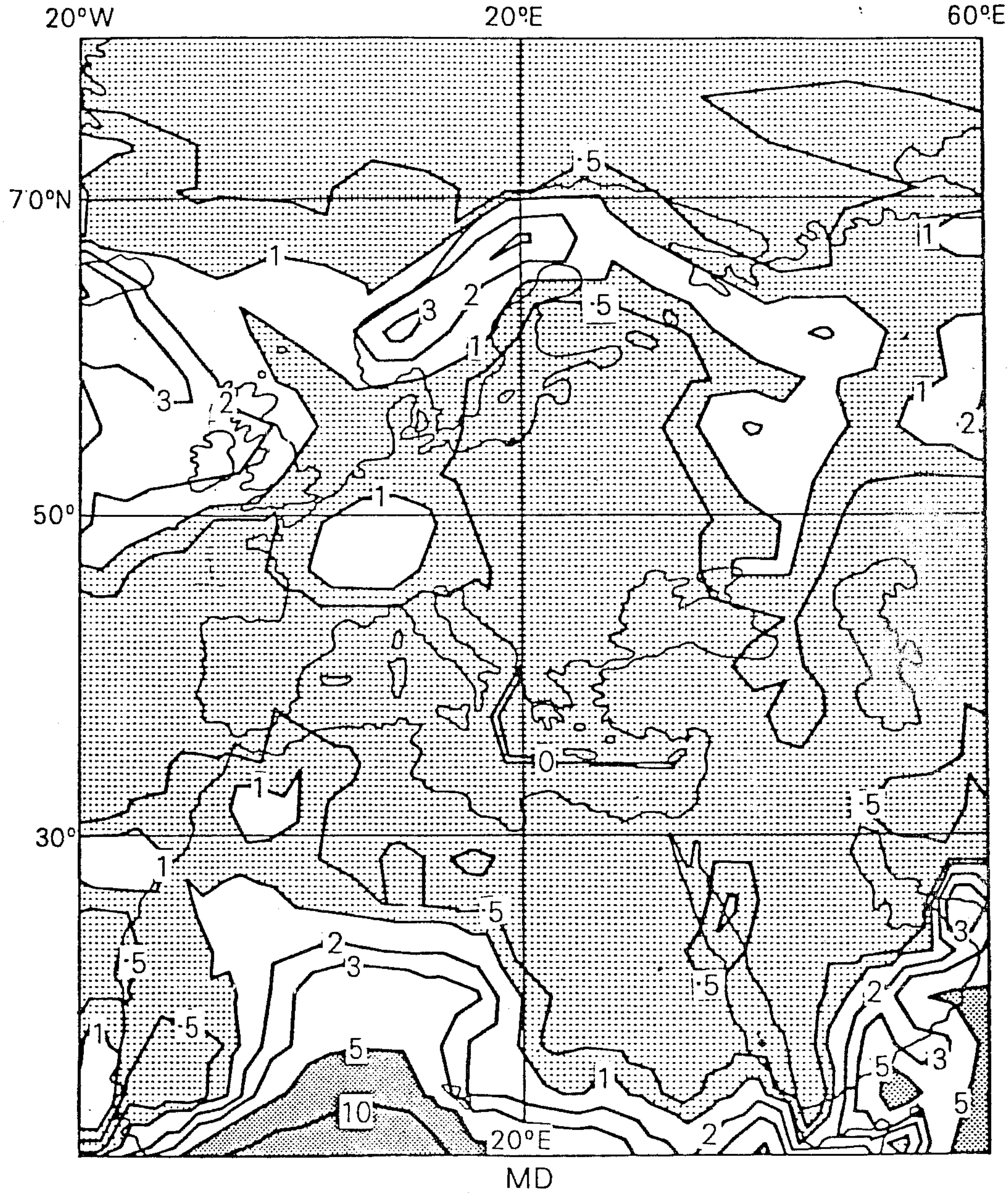


FIGURE 17 Rainfall (mm) averaged for Days 21 to 50 of experiment (b) MD

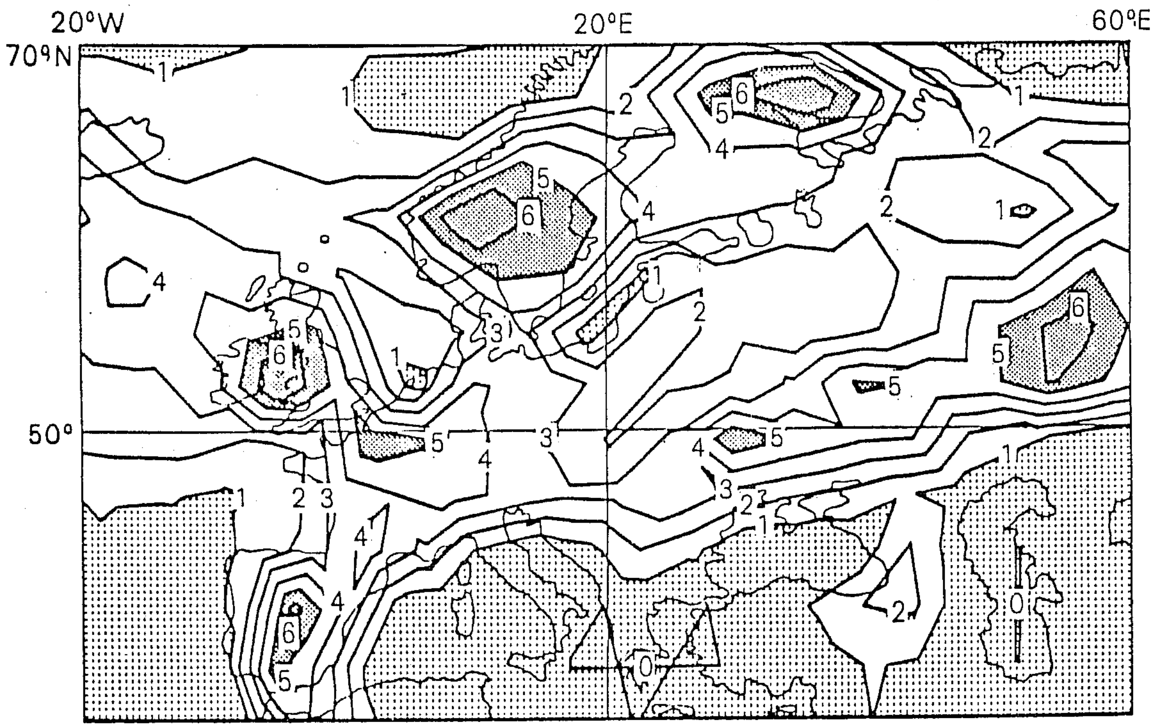
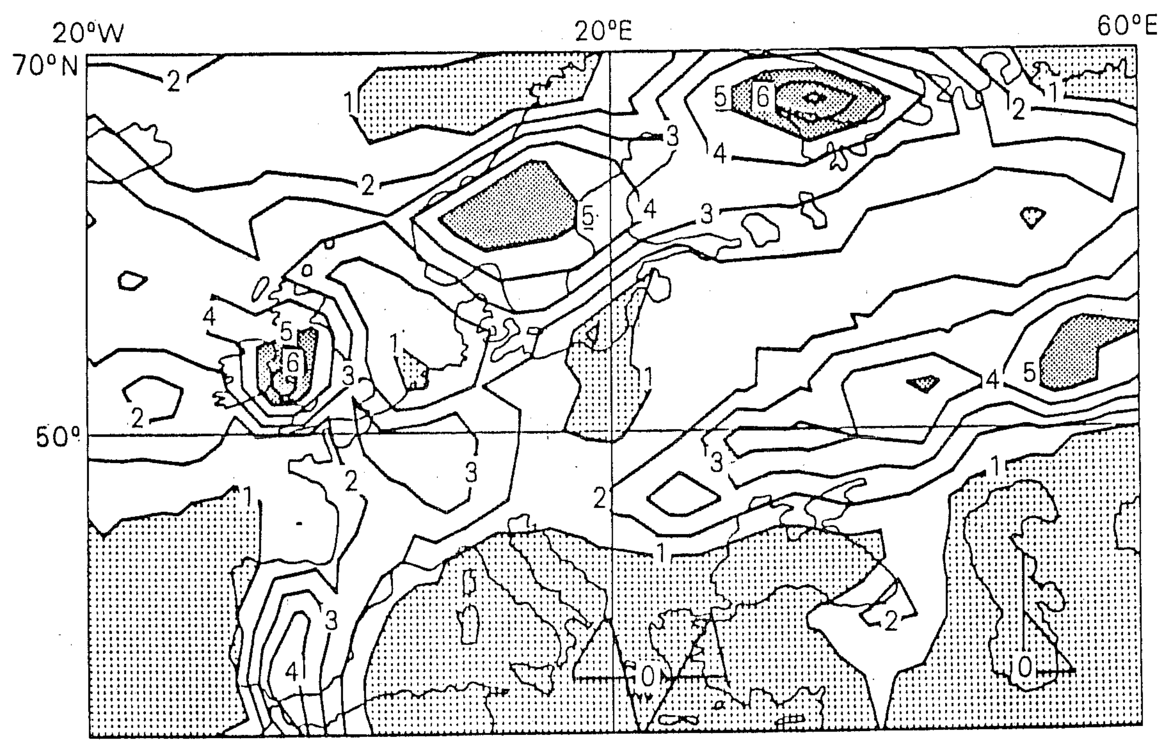


FIGURE 18 Rainfall (mm/day) averaged for Days 1-10 of the

(a) control,

(b) dry experiments with the annual cycle model.



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