EXPERIMENTS WITH LAND SURFACE SCHEMES AT THE UK METEOROLOGICAL OFFICE

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1. INTRODUCTION

This paper reviews some experiments using the UK Meteorological Office Atmospheric General Circulation Model (AGCM) (Slingo et al.,1989) and a Single Column Model (SCM) which incorporates the physics of the AGCM. From the early stages of the development of these models, sensitivity experiments were carried out to determine the response of the simulated climate to changes in the specified land surface parameters and to the representation of land surface processes. The earliest versions of the model had uniform values of albedo, water capacity of soil, surface roughness and resistance to evaporation over all land points (except that the albedo had a different value for snow covered land).

Recent work has replaced the globally constant land surface parameters by geographically varying values. The representation of land surface processes has also been been changed to include a four layer soil temperature model, a somewhat more realistic soil hydrology scheme and a parametrization of the interception of rainfall by a vegetation canopy. Results to indicate the impact of some of these changes are presented. Further experiments to explore some of the effects of the land surface and its parametrization in simulations of climate and climate change are discussed.

2. EXPERIMENTS WITH EARLY VERSIONS OF THE MODEL

Soil moisture content has always been a predicted variable in the AGCM. Experiments with anomalies in the initial values of soil moisture in central Europe have been done by Rowntree and Bolton (1983). They found that the different partitioning between surface sensible and latent heat fluxes had major effects on simulated rainfall and atmospheric humidity both in the anomaly area and, by propagation, over adjacent land areas. The effects on rainfall were already important by the third day of the integration in the anomaly area and by the fifth day over regions outside this area. The results were dependent, as might be expected, on the prevailing flow: moist westerly flow led to a shorter period for the persistence of a dry anomaly.

The importance of surface albedo specification was demonstrated by Cunnington and Rowntree (1986). This work showed that a realistic simulation of Saharan climate was dependent on the specification of the surface albedo and the initial soil moisture content (and the initial atmospheric humidity). In particular too low a value for the albedo for desert led to the Sahara being too wet, in accordance with Charney's hypothesis.

3. THE DEVELOPMENT OF A MORE DETAILED LAND SURFACE PARAMETRIZATION

3.1 Recent changes to the treatment of land surface processes

Over recent years the representation of surface and sub-surface processes in the AGCM has been made more detailed so that both present day climate and possible changes in climate caused by human activity can be simulated with increased accuracy and physical realism. Four main aspects have been developed:

- Heat conduction within the soil is represented using a four-layer soil thermodynamics model (see the contribution of Rowntree in these Workshop Proceedings). The number of layers and their thicknesses have been chosen to give the best response of both amplitude and phase to forcing at natural periods (1 day to 2 to 3 years).

- The previous 'bucket model' treatment of soil hydrology has been replaced by a scheme which includes fast (surface) runoff and slow (deep) runoff by gravitational drainage of soil water (Warrilow, 1986a).

- A vegetation canopy of specified, spatially varying, capacity has been introduced (Warrilow, 1986b). This intercepts and stores some of the rainfall and provides a mechanism for this to re-evaporate at zero resistance. The water which is not intercepted falls through to the underlying soil.

- Globally constant land surface parameters have been replaced by spatially varying values. 17 soil and vegetation parameters are specified

Table 1

List of parameters defined from soil and vegetation datasets

Soil parameters
Soil moisture concentration at wilting point
Soil moisture concentration at the critical point
Soil moisture concentration at field capacity
Soil moisture concentration at saturation
Saturated conductivity for soil moisture flow
Exponent in soil conductivity relationship
Volumetric heat capacity
Thermal conductivity
Vegetation parameters
Vegetation fraction
Root depth
Deep snow albedo
Snow free albedo
Surface resistance to evaporation
Canopy capacity
Surface infiltration capacity enhancement factor
Surface roughness

(see table 1). Only surface albedo had been treated as geographically varying in the AGCM previously.

3.2 The derivation of global land surface datasets

The procedure for deriving the parameters from the Wilson and Henderson-Sellers (1985) global 1°x1° resolution soil and vegetation type dataset is described in Warrilow and Buckley (1989). Figure 1 shows the three main stages in the process of converting the 1°x1° vegetation and soil type data to m°xn° parameter datasets. Firstly, parameter values were assigned to a subset of basic vegetation and soil types; secondly 1°x1° parameter datasets were produced by linear combination of the basic types; finally the m°xn° parameter datasets were derived by weighted averaging of the 1°x1° values. Averaging is therefore done on parameter values rather than by attempting to find the average soil and vegetation types at coarser resolution. Surface albedo was the only parameter which was derived from both the soil and vegetation datasets.

Maps showing the global distribution of some of the land surface parameters at 2.5°x3.75° resolution are shown in figures 2 to 6. Figure 2 shows that the derived surface albedo of much of the Sahara is about 0.3 with a maximum of 0.35. Maximum albedos of 0.75 are found over the land-ice regions of Greenland and Antarctica and minimum land albedos of 0.12 are specified for equatorial rainforest. The albedo of the ocean is set to 0.06 everywhere.

In previous versions of the model deep snow albedo was set to 0.6 over all land points where the water equivalent snowdepth exceeded 1.1 cm. Following Robinson and Kukla (1984 and 1985) a global variation of deep snow albedo with vegetation type was derived. This is shown in figure 3, with values varying from 0.8 for snow over land-ice, tundra and deserts to as low as 0.2 for evergreen forest where snow soon falls from the canopy to the ground.

In figure 4, z_0, the surface roughness length, ranges from the order of 10^{-4} to 10^{-3} m over the relatively smooth ice caps and deserts to near 1 m in forested regions. The previous uniform value for land areas was 10^{-1} m.

Canopy capacity, shown in figure 5 ranges from zero over the ice caps and deserts where there is no vegetation to 2.5 mm in tropical forests.

Field capacity, shown in figure 6, is not used explicitly in the

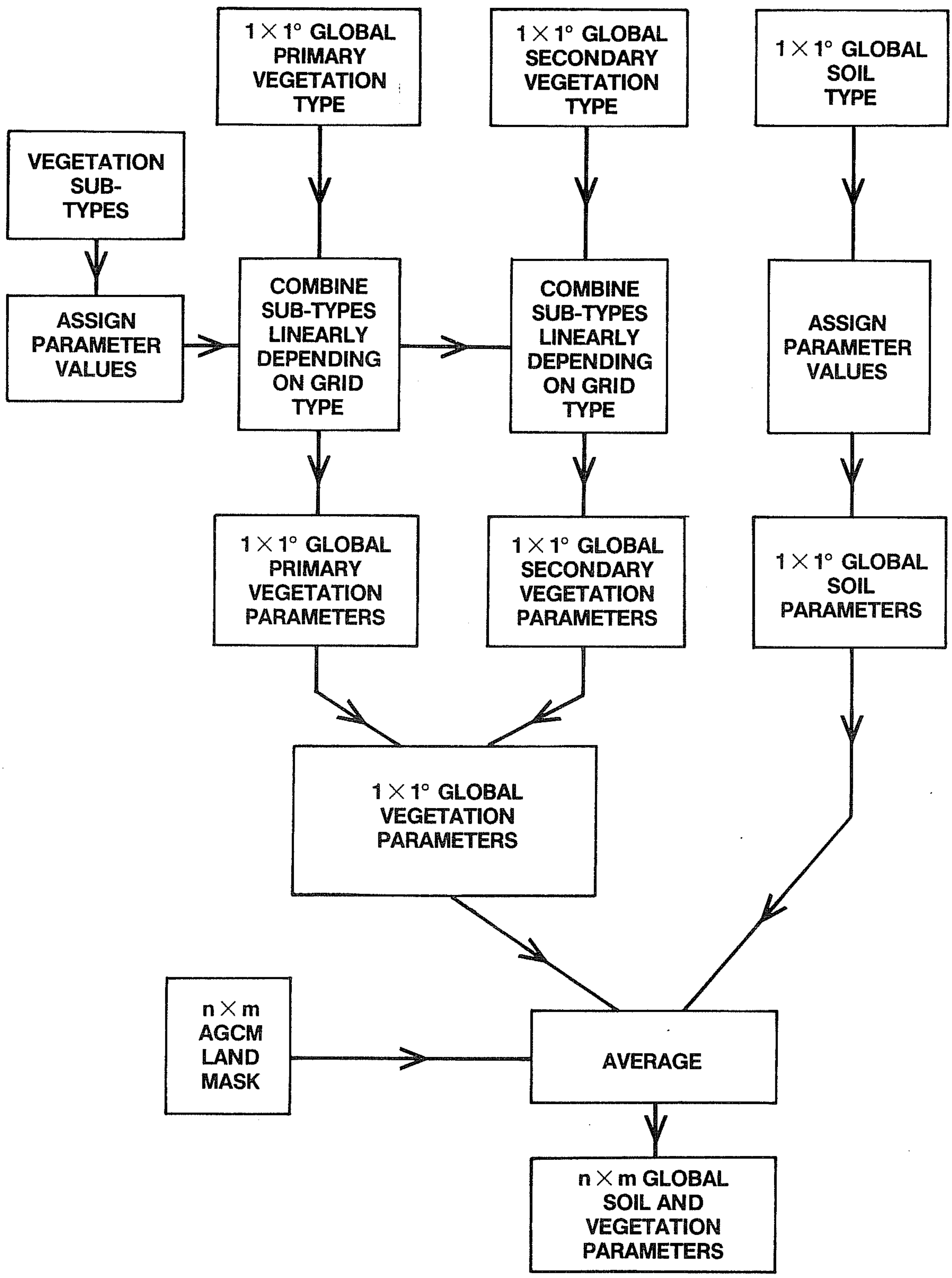

Figure 1 The steps taken in deriving land surface datasets for the AGCM

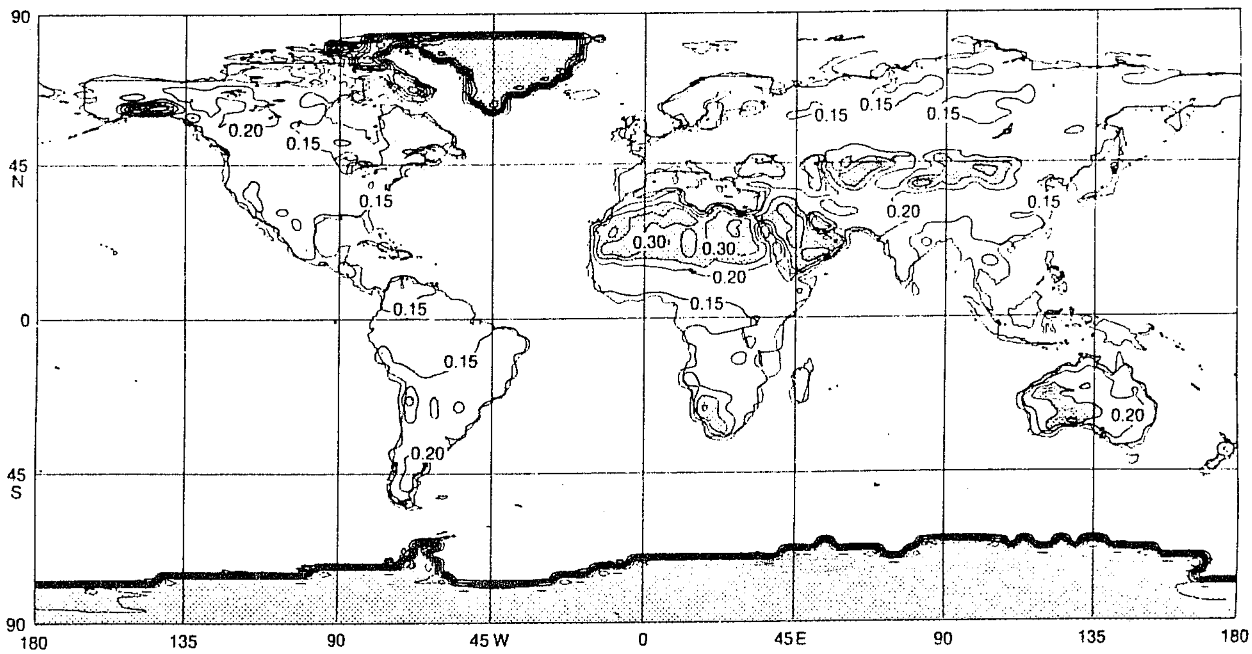

Figure 2 Snow free albedo. Contours every 0.05 from 0.15 to 0.35 and shaded above 0.25.

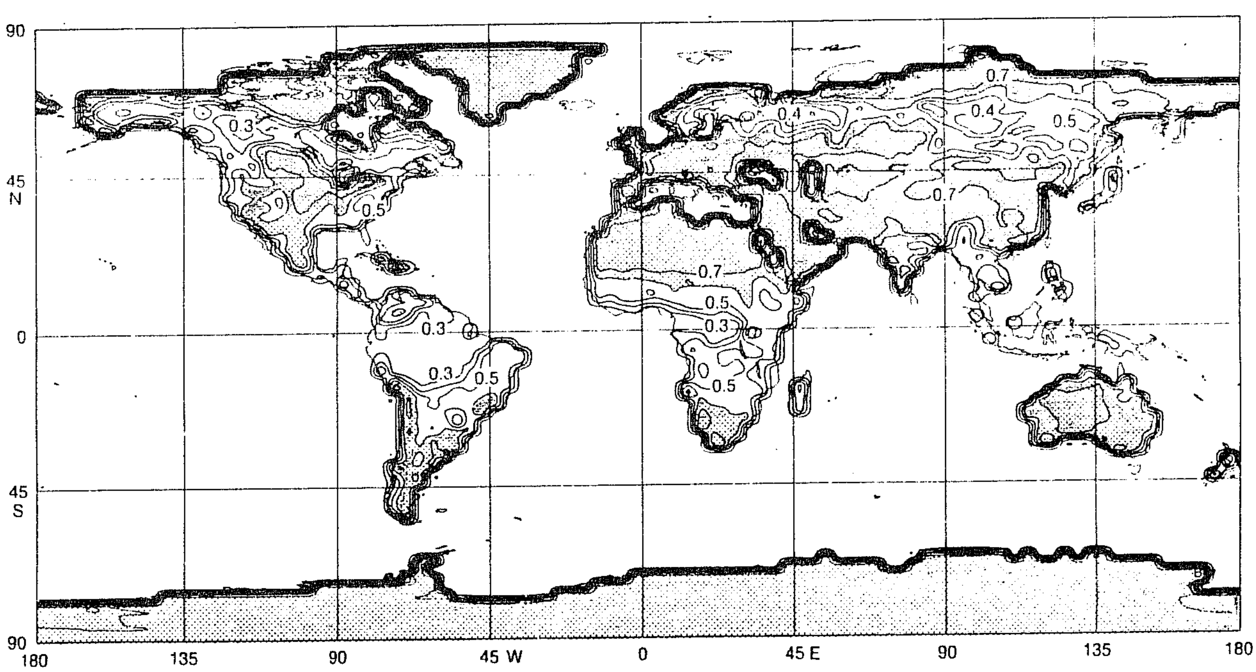

Figure 3 Maximum deep snow albedo.
Contours every 0.1 and shaded above 0.6.

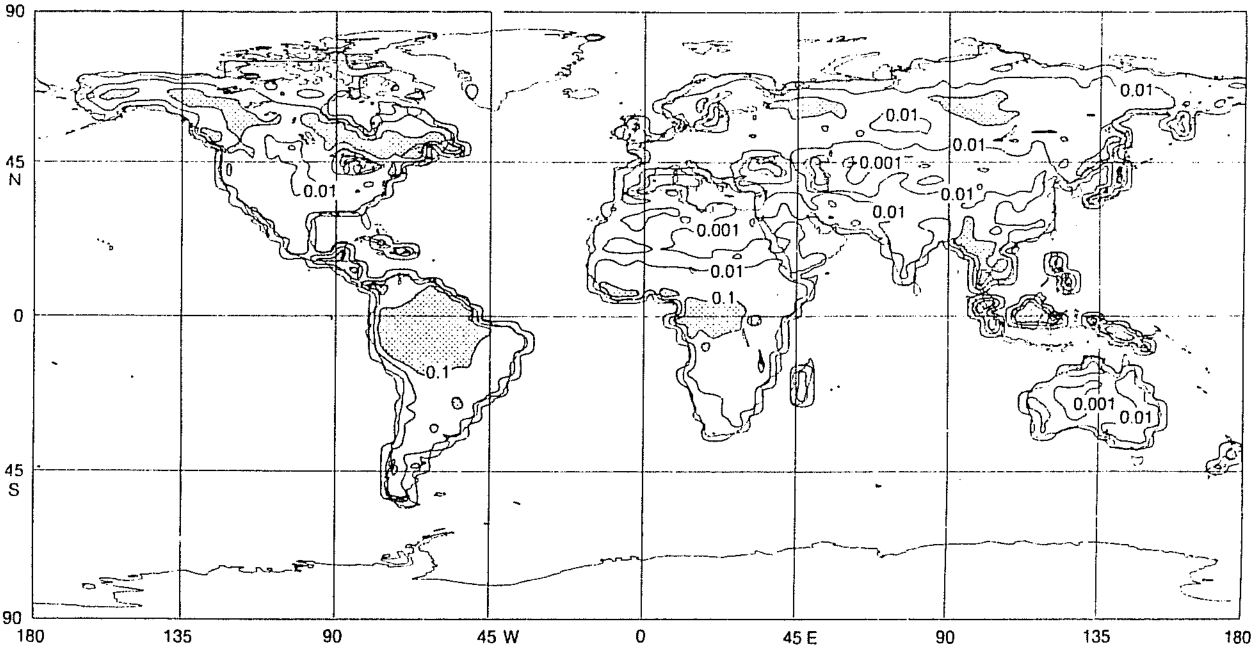

Figure 4 Surface roughness. Contours at 0.001, 0.01, 0.1 m and shaded above 0.1 m.

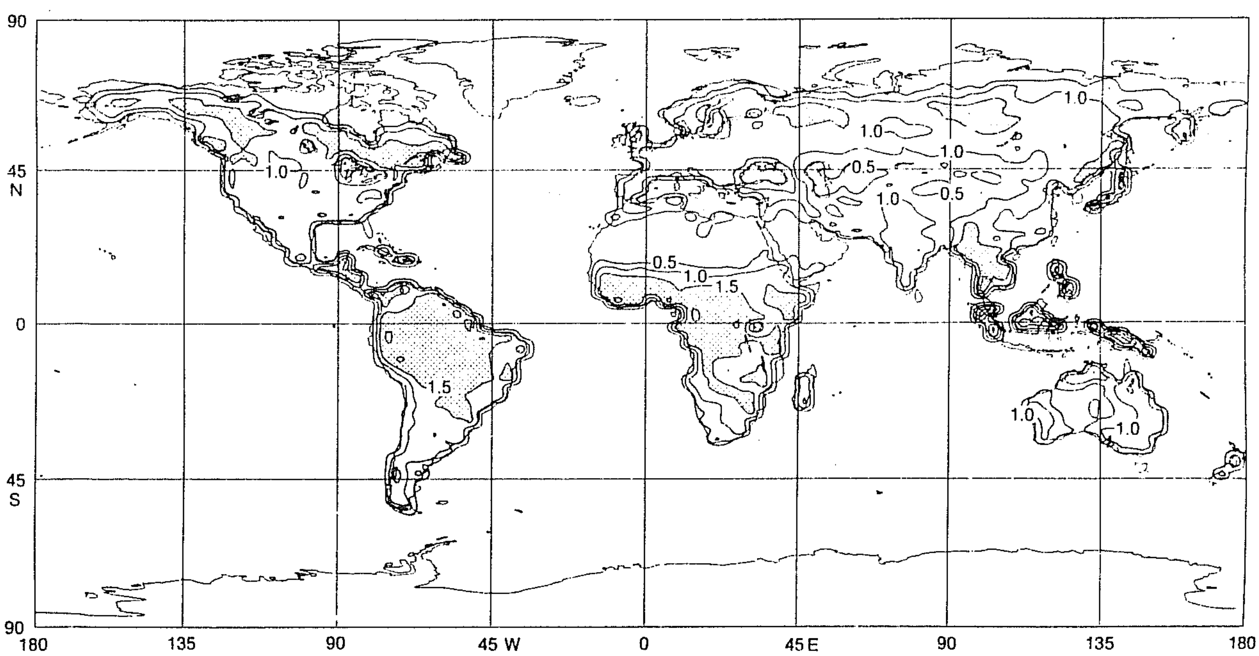

Figure 5 Canopy moisture capacity. Contours every 0.5 mm and shaded above 1.5 mm.

model, but since it is based on root depth and soil type it gives a good idea of the water holding capacity of the soil. Values range from zero over the ice caps to as high as 25 cm in the tropical forests. Arid regions have a capacity less than 2cm. Much of the land has a lower field capacity than the previous global value of 15cm.

4. AN EXPERIMENT TO DETERMINE THE IMPACT OF NEW FEATURES OF THE LAND SURFACE SCHEME

4.1 The experiment

Two integrations of the AGCM are compared in this section. The model used for the control integration ("EXPC") was a slightly modified version of that described by Slingo et al. (1989): it included the four-layer soil thermodynamics scheme and the more complex soil hydrology parametrization outlined above (with the spatially uniform parameters of the "test soil" used by Warrilow (1986a)). The modifications included a correction to surface heat capacity during snowmelt and an early version of a snow insulation parametrization. The anomaly integration ("EXPV") was the same as the control apart from inclusion of vegetation canopy interception and evaporation and globally varying soil and vegetation parameters.

4.2 The impact of canopy interception and evaporation

Canopy interception and evaporation are important new processes which have been added to the model in EXPV; the existence of a canopy allows immediate evaporation at zero resistance. Its importance has as much to do with the frequency and intensity of rainfall as the size of the canopy store.

Figures 7a,b show the impact of canopy moisture processes on total evaporation over Africa and South America. These results are very similar to those obtained in another experiment, not shown here, in which only canopy processes were introduced. Increases in evaporation of between 10 and 20% are typical over much of the moist regions.

Figure 8 shows canopy evaporation alone. Clearly canopy processes can have a significant regional impact. Of particular interest is the change in evaporation over equatorial rainforest regions where canopy capacity is large and rainfall events are often frequent. Shuttleworth (1988) in an observational study of evaporation at a site at (3°S, 60°W) in the Amazon over a period of two years found an average July precipitation of 4.42 mm

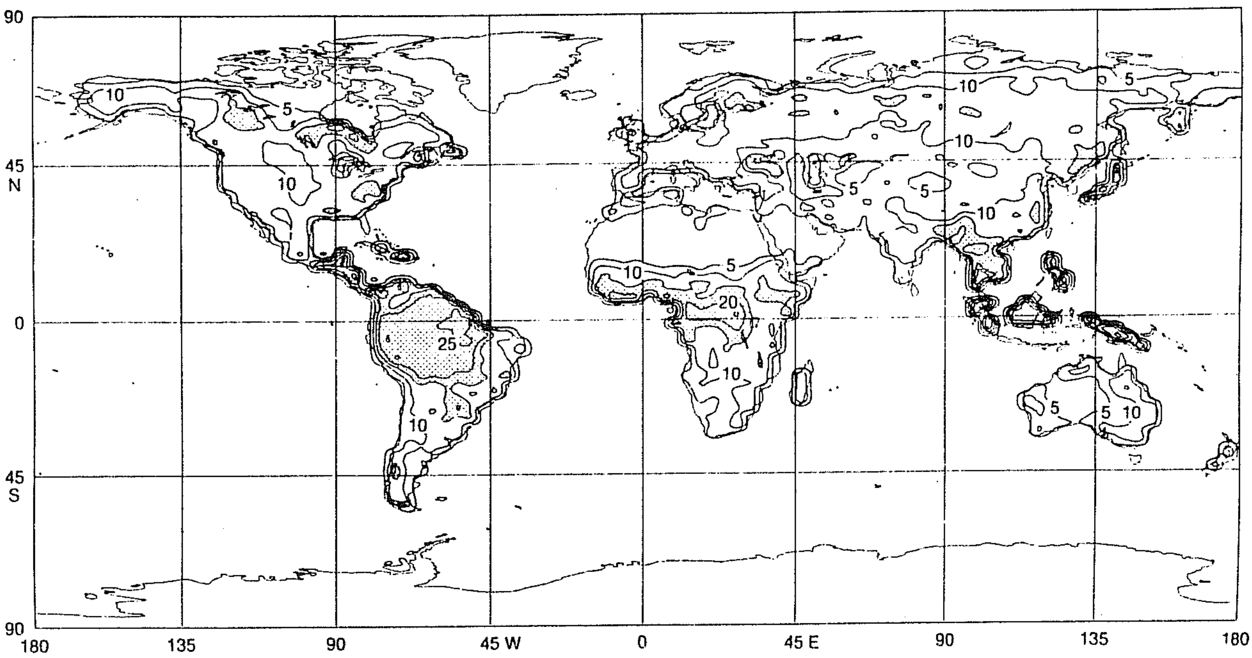

Figure 6 Soil moisture content at field capacity.
Contours every 5 cm and shaded above 15 cm.

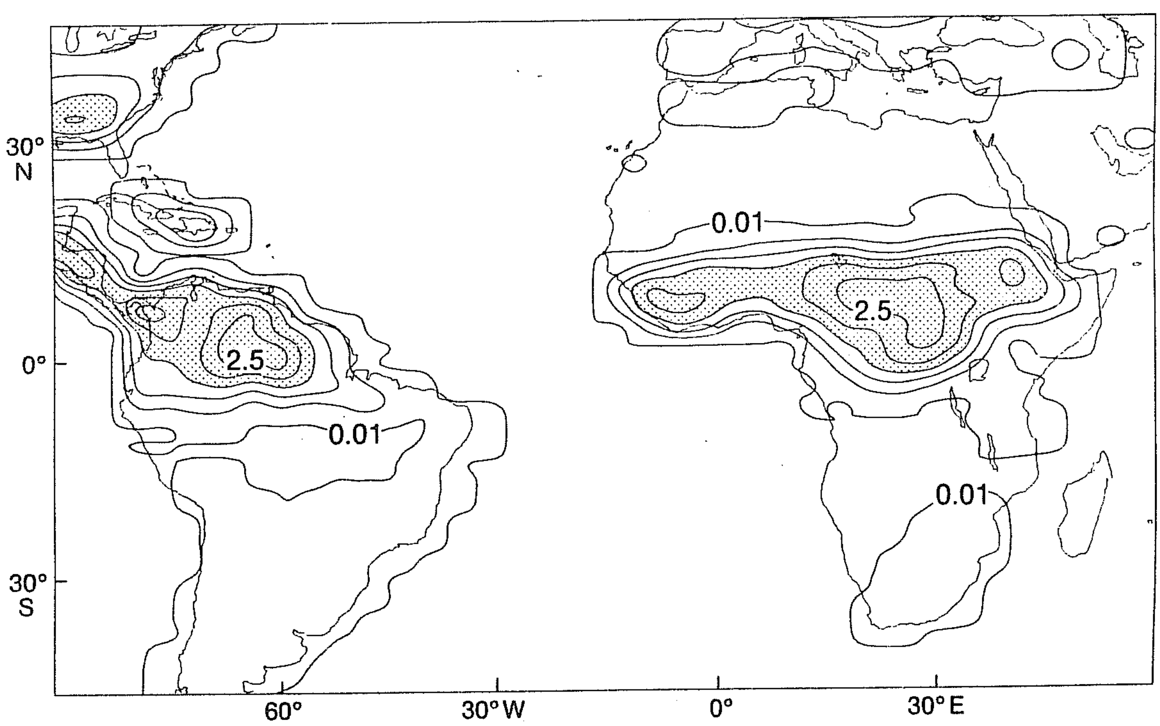

Figure 8 Canopy evaporation in July from EXPV. Contours at 0.01, 0.5 and every 0.5 mm day^{-1} and shaded above 1.5 mm day^{-1}.

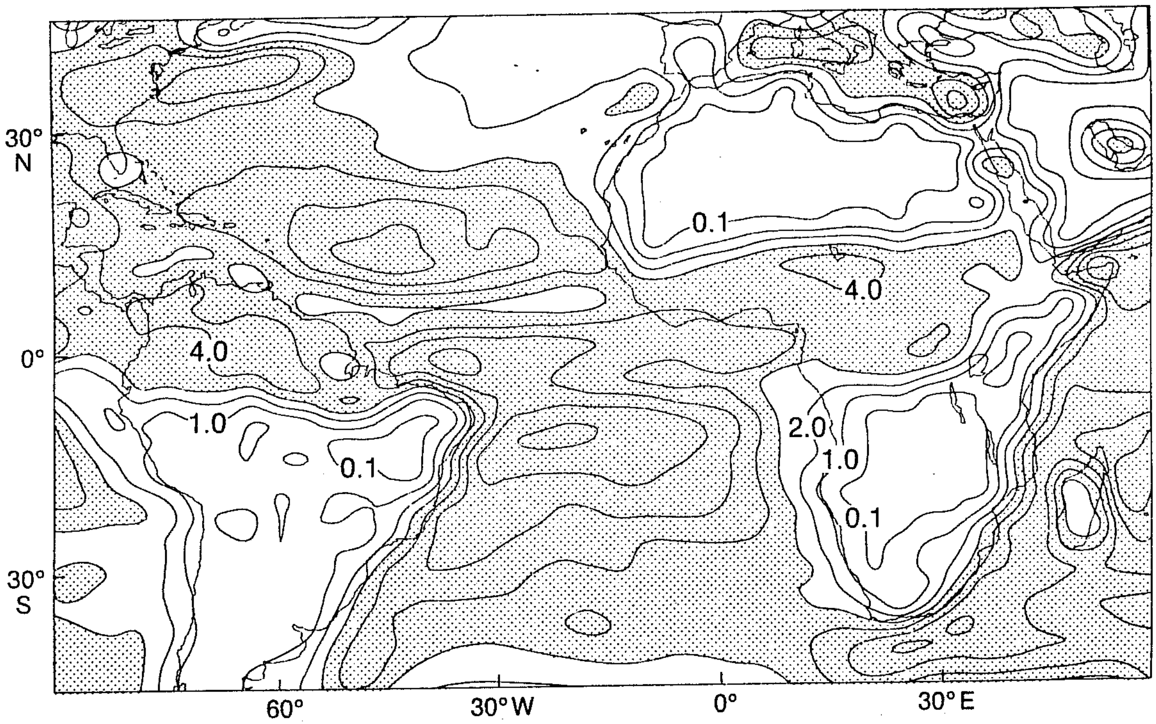

Figure 7a Evaporation in July from EXPV. Contours at 0.1, 1.0 and then every 1.0 mm day^{-1} and shaded above 3.0 mm day^{-1}.

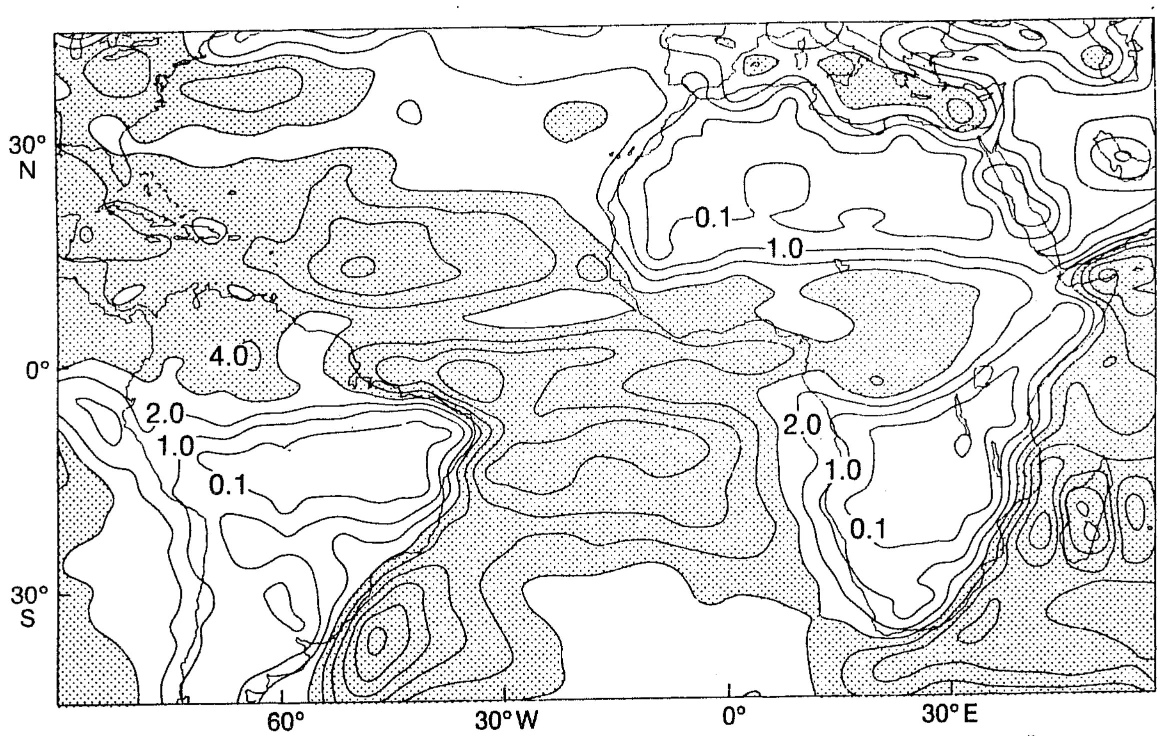

Figure 7b As 7a but for EXPC.

day^{-1}, evapotranspiration of 3.84 mm day^{-1} of which canopy evaporation was 0.87 mm day^{-1}. The maximum evapotranspiration occurred in September when the total was 4.33 mm day^{-1} and the canopy evaporation was 1.0 mm day^{-1}. Shuttleworth et al. (1984) reported average transpiration from a dry canopy over a period of 8 days at the same site to be 3.37 mm day^{-1}. Walsh (1987) in a summary of tropical forest evaporation quotes total annual evaporation varying from 3.01 - 4.93 mm day^{-1} with interception losses varying from 0.55 - 3.84 mm day^{-1} (this is approximately 20% of the annual rainfall (range 12 - 35%)). The rate of canopy evaporation depends on the frequency of rainfall and so tends to be correlated with the rainfall total. The model values are not unrealistic although the Shuttleworth (1988) observations suggest that they may be too high. The grid-averaged canopy capacity may therefore be too high.

Figures 9a,b show the impact of canopy processes on simulated rainfall. Although there is an increase in rainfall in EXPV there is also a greater spatially random response. Comparison with a rainfall climatology in figure 10 (Jaeger, 1976 and 1983) suggests that the simulation of the spatial distribution of Amazonian rainfall has improved slightly but there is relatively little significant change over Africa. However, rainfall remains too low in EXPV to the south of the Amazon, despite the existence of a large canopy store (figure 5) and therefore the potential for evaporation of freely available water. Thus it is clear that where the simulated general circulation, as determined largely by the model's dynamics, gives inadequate rainfall, changes to the land surface scheme will have little effect. (Failure of the EXPV simulation to show any improvement in the Indian monsoon corroborates this). The rainfall simulated by EXPV in the Amazon is too large at other times of the year. This supports the view, discussed above, that canopy evaporation may be overestimated.

4.3 The impact of surface roughness

The surface roughness length is an important parameter in the calculation of surface turbulent fluxes and therefore affects the heat and moisture (and momentum) exchange at the land surface. From figure 4 it is evident that in EXPV z_0 is lower than the control value of 10^{-1} m over much of the land surface except in forested regions. Over deserts the differences are largest with z_0 reduced to about 3×10^{-4} m and here the

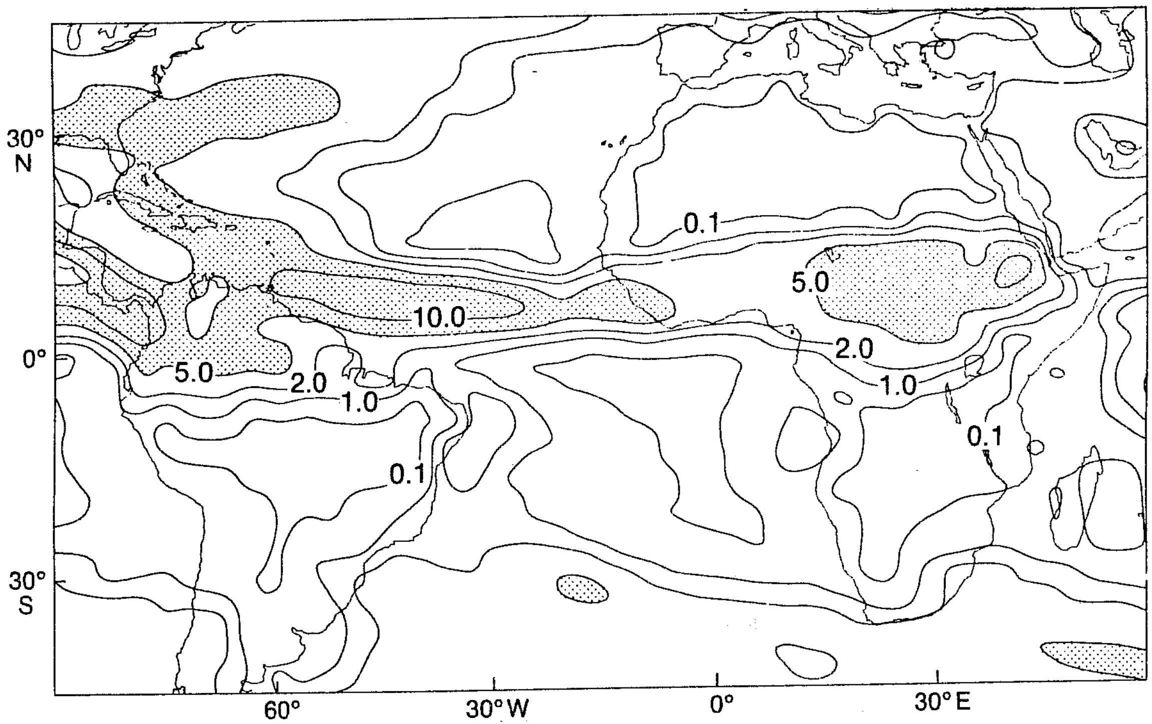

Figure 9a Rainfall in July from EXPV. Contours at 0.1, 1.0, 2.0, 5.0, 10.0 mm day^{-1} and shaded above 5.0 mm day^{-1}.

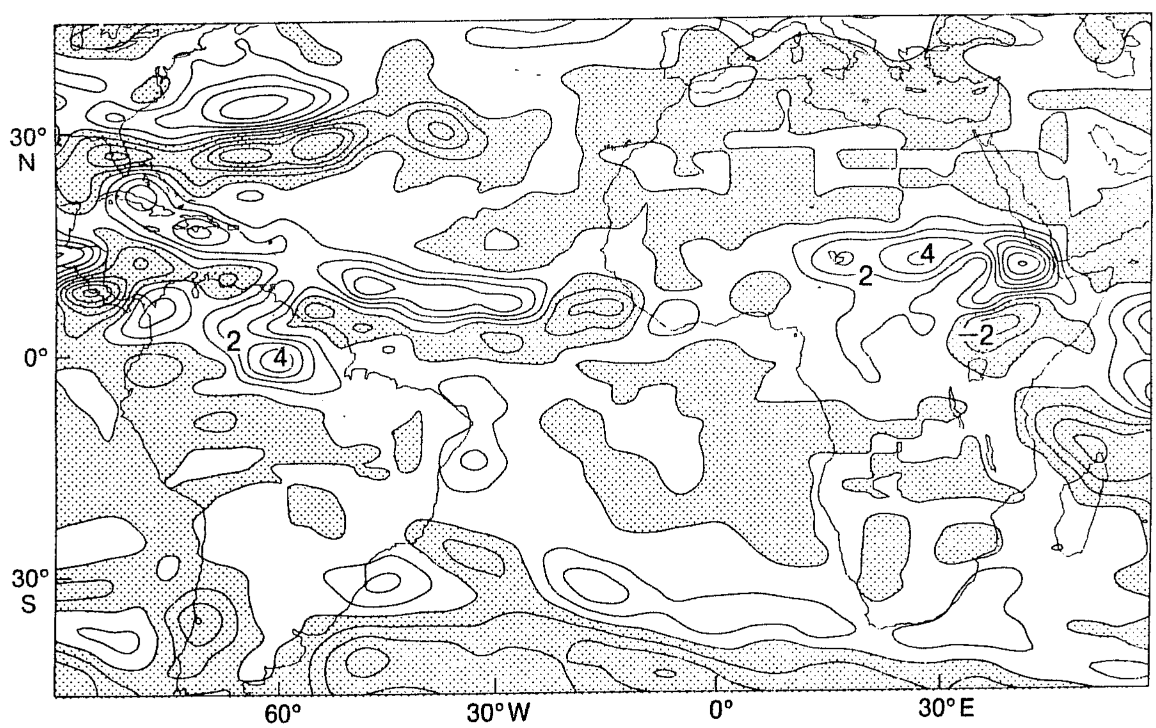

Figure 9b Difference in rainfall (EXPV - EXPC) for July. Contours every 1.0 mm day^{-1} and shaded negative.

largest impact is found. Reduced roughness leads to a decrease in the turbulent exchange coefficients. This does not necessarily imply a decrease in the surface fluxes of the same proportions because there is an increase in the low level wind speed as a result of reduced drag. Nevertheless the heat flux from the dessert surface is lower in EXPV than in EXPC which has a much higher roughness. Since deserts are dry, evaporation is negligible and heat can only be removed from the surface by the sensible heat flux and by longwave radiative cooling. The latter only balances the reduced sensible heat flux at a higher surface temperature.

Figure 11 shows the difference in maximimum temperatures between EXPV and EXPC. Maximum surface (skin) temperatures in EXPV reach values locally in excess of 60°C, in accordance with locally observed values in desert regions. The maximum observed values quoted by Cloudsley-Thompson (1965) and Chang (1958) are about 80-85°C.

It is not clear from these experiments if there is a significant increase in evaporation over the tropical forest regions due to increased roughness since the changes caused by the introduction of a canopy in EXPV are dominant.

5. SOME CLIMATE CHANGE EXPERIMENTS AND THEIR SENSITIVITY TO LAND SURFACE PROCESES

5.1 An experiment showing the sensitivity of doubled CO_2 model simulations to the treatment of runoff

Results obtained by Mitchell and Warrilow (1987) have shown that the summer drying in northern middle and high latitudes which had been simulated by some GCMs when run with increased carbon dioxide can be reduced or even reversed by an alternative, equally plausible, treatment of runoff over frozen ground.

Figure 12a shows that in spring the model's soil moisture reaches its maximum possible value; this is due mainly to snowmelt. (The simulated maximum for present CO_2 concentrations occurs late as the version of the AGCM used in this study is too cold in spring.) On increasing CO_2 the soil moisture increases in winter and spring due to increased precipitation. Snowmelt and therefore saturation of the ground and subsequent summer drying occur earlier, producing a reduction in summer soil moisture. In this simulation it was assumed that all precipitation and snowmelt go to increase soil moisture unless this is at its maximum possible value. In

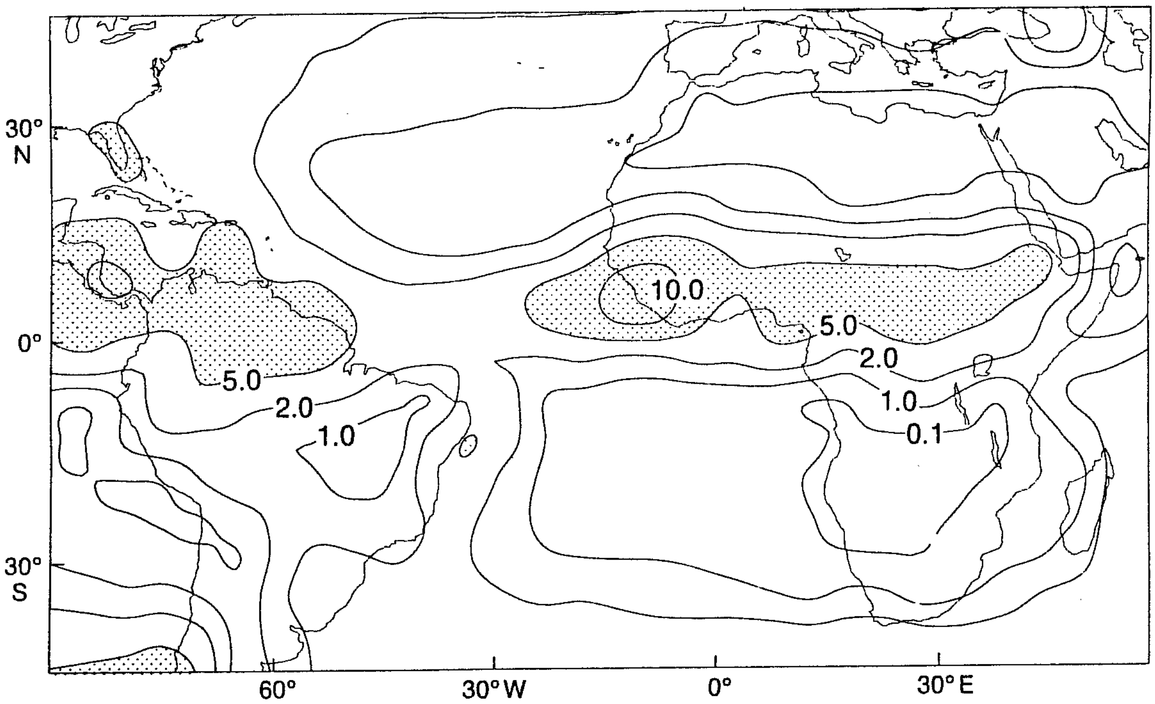

Figure 10 Rainfall climatology in July from Jaeger (1976). Contours at 0.1, 1.0, 2.0, 5.0, 10.0 mm day^{-1} and shaded above 5.0 mm day^{-1}.

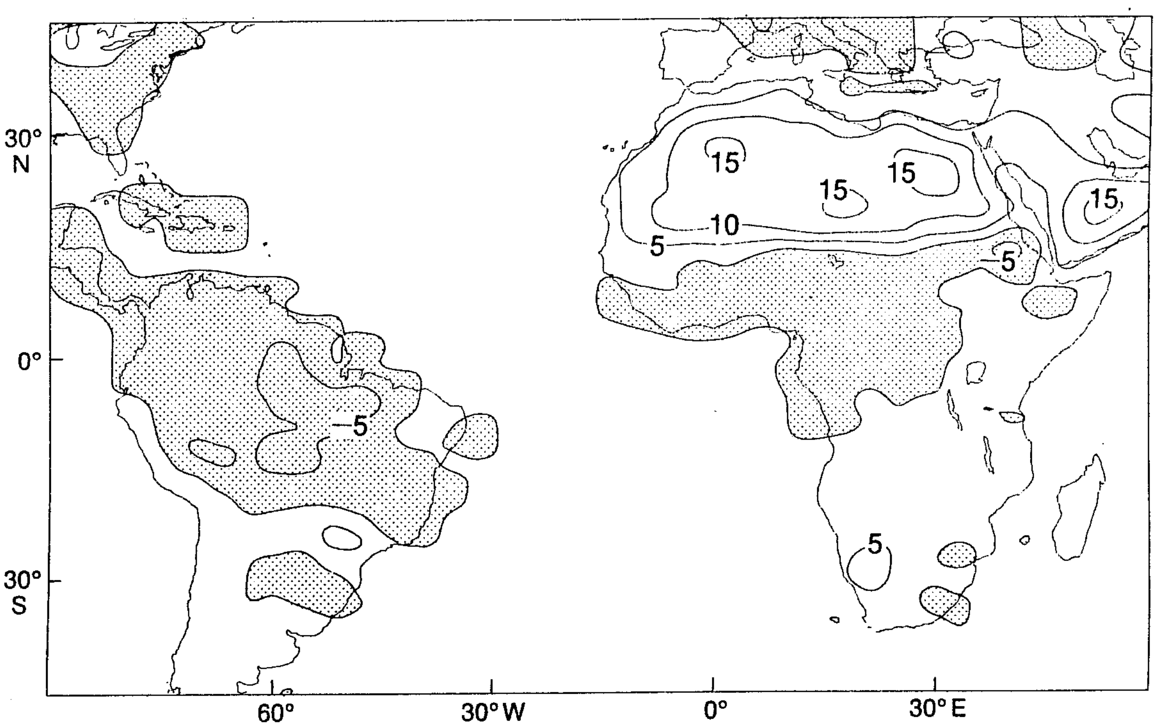

Figure 11 Difference in maximum surface temperature in July (EXPV - EXPC). Contours every 5 K and shaded negative.

reality, infiltration will be limited if the soil is frozen. The release of snowmelt-water may be sufficiently rapid that little infiltrates the soil, most running off directly.

In a second pair of experiments all rain or snowmelt is run off immediately if either of the middle two soil layers used for soil temperature calculations (centred at 0.1 and 0.41 m) are sub-freezing. This is an extreme assumption but it is appropriate for a sensitivity study such as this. Figure 12b shows that this change reduces soil moisture amounts in the control (present CO_2) simulation; indeed the soil no longer becomes saturated in spring. Thus in the increased CO_2 simulation, increased moisture accumulated in autumn persists, and although summer drying starts earlier and is faster than in the control, it starts from a higher value. This contrasts with the original case in which both simulations start from similar saturated values but at different times (figure 10a). Hence, with the revised formulation, the soil does not become drier until late summer and then only slightly so.

These experiments show that the nature of the simulated changes in soil moisture between 45° and 60°N in summer due to increased CO_2 are sensitive to the formulation of surface runoff. Other factors including the formulation of evaporation and the effects of vegetation may also be important but have yet to be studied with the Met Office GCM.

5.2 Some studies with a single column model of the effects of deforestation

A Single Column Model (SCM) incorporating the "physics" schemes of the AGCM has been used for a preliminary study of the effects of deforestation. The large scale dynamical processes which the SCM cannot represent are imposed as forcing terms in the prognostic equations. The fluctuations in the heat and moisture convergences about a climatic mean for the grid-point are treated statistically. The details of the way the statistics are worked out are given by Warrilow et al. (1986). The experiments descibed below used forcing data for a point (70° W, 0°N) in the Amazon basin in February.

An experiment with surface vegetation parameters appropriate for tropical forest was used as the control. In a second experiment the parameters were replaced by those for short grass. A series of four further experiments was done with just one of the vegetation parameters

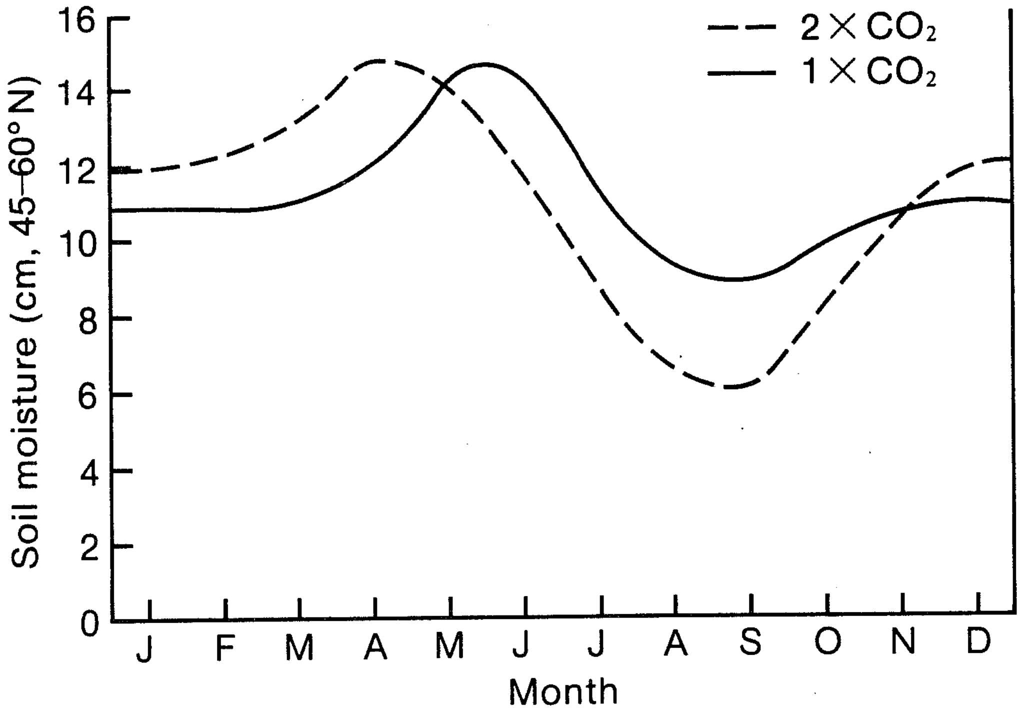

Figure 12a Simulated monthly mean soil moisture (cm) averaged over land between 45° and 60° N. Solid line, present CO_2 concentrations; dashed line, doubled CO_2. (After Mitchell and Warrilow, 1987)

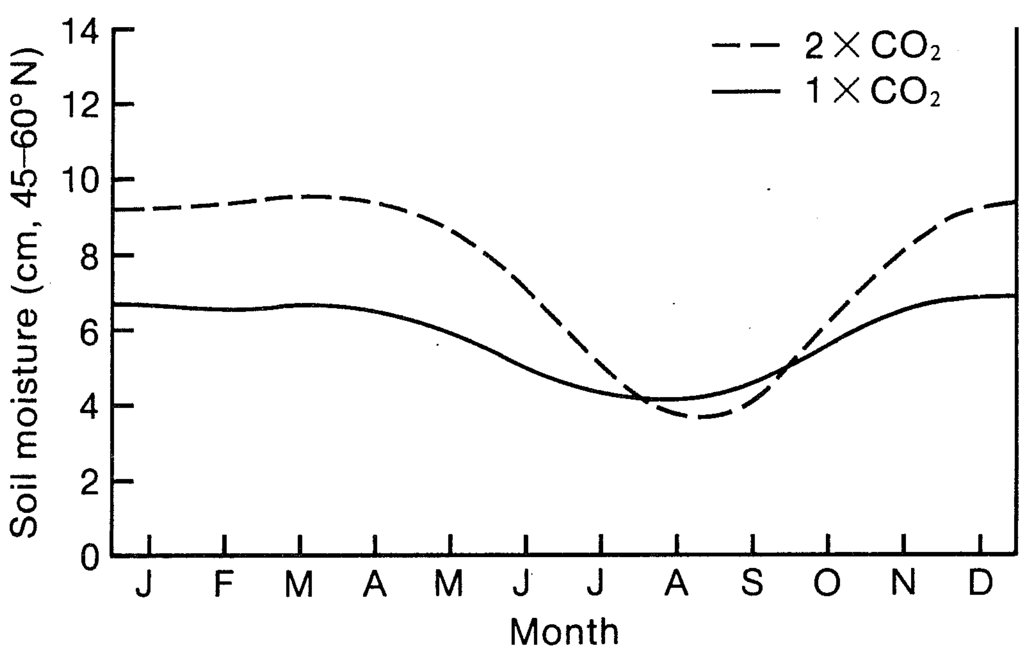

Figure 12b As Fig. 12a, but with 100% runoff when the second and/or third soil layers are frozen

changed in turn to its grass value. The values of the parameters used in each of these six experiments are given in Table 2. The parameters representing the soil type were the same in all the experiments and had the values of the "test soil" used by Warrilow (1986a). In order to reduce feedbacks from processes not directly connected with the surface scheme the amounts of convective and layer cloud were held fixed in all the experiments: this ensures the incoming solar radiation at the surface is the same for all the experiments.

Table 3 gives the values of the main surface quantities meaned over the month of the simulation for all six experiments. Also shown are the changes from the control. The differences in the surface heat fluxes between the grass and forest cases (G-F) are negative, the change in latent heat flux being numerically larger. As required for an approximate balance of surface fluxes, the net radiative flux change is also negative and of similar magnitude to the total turbulent heat flux change. The surface temperature for grassland is higher and there is a corresponding increase in the surface outgoing longwave radiation.

Some indication of the causes of the G-F differences can be sought from the experiments A, Z, R and C with only one surface parameter changed in turn. The reduced latent heat flux in G can be attributed to the lower surface roughness, the higher albedo and (rather less importantly) the lower canopy capacity. The lower resistance to evaporation partly offsets these.

The surface temperature increase for the grassland simulation can be seen to be due to the lower surface roughness partly offset by the lower resistance to evaporation. The reduced net downward radiative flux for grassland is caused, in approximately equal measures, by the higher albedo and the lower surface roughness. The lower roughness affects the radiative flux through the increased surface temperature and correspondingly increased upward longwave flux.

The experiments show that a change in one surface parameter alone could give a misleading indication of the effects of a change in surface vegetation. The scheme outlined above in section 3.1 for deriving a consistent set of parameter values from soil and vegetation type data is therefore necessary for realistic simulations of the effects of changes in land use on climate. The ability to change just one parameter in turn in the model is, however, useful for trying to explain the net effects of a

TABLE 2: The specified surface parameters used in single column model experiments.

Expt.	Description	Albedo	Surface roughness length (m)	Resistance to evaporation (sm^{-1})	Canopy capacity (m)	Root depth (m)
F	Tropical Forest	0.14	2.00	150.0	0.0025	1.5
G	Grassland	0.20	0.02	50.0	0.001	0.6
A	Forest with increased Albedo	0.20	2.00	150.00	0.0025	1.5
Z	Forest with reduced surface roughness length, Z_o	0.14	0.02	150.00	0.0025	1.5
R	Forest with reduced resistance to evaporation	0.14	2.00	50.00	0.0025	1.5
C	Forest with reduced canopy capacity	0.14	2.00	150.00	0.001	1.5

TABLE 3: Surface quantities simulated by a single column model with various land surface types.

Expt.	Description	T_*(°C)	Heat Flux (Wm^{-2}) Sensible	Latent	Total	Rnet (Wm^{-2})	RSnet (Wm^{-2})	RL (Wm^{-2})	Evaporation (mm d^{-1})	PPN (mm d^{-1})	Total Runoff (mm d^{-1})
F	Tropical Forest	24.9	18.0	117.6	135.6	134.0	160.3	448.1	4.08	6.98	1.68
G	Grassland	25.9	14.1	107.1	121.2	120.5	151.7	453.9	3.72	6.65	2.43
G-F	difference	1.0	-3.9	-10.5	-14.4	-13.5	-8.6	5.8	-0.36	-0.33	0.75
A	F with increased Albedo	24.7	15.6	111.8	127.4	125.6	151.7	447.1	3.88	6.91	1.74
A-F	difference	-0.2	-2.4	-5.8	-8.2	-8.4	-8.6	-1.0	-0.20	-0.07	0.06
Z	F with reduced Roughness, Z_o	26.9	20.5	105.6	126.1	126.1	160.3	460.4	3.67	6.60	1.69
Z-F	difference	2.0	2.5	-12.0	-9.5	-7.9	0	12.3	-0.41	-0.38	0.01
R	F with reduced Resistance	23.9	8.1	130.1	138.2	136.1	160.3	442.1	4.52	7.22	1.56
R-F	difference	-1.0	-9.9	12.5	2.6	2.1	0	-6.0	0.44	0.24	-0.12
C	F with reduced Capacity	25.1	19.9	115.2	135.1	133.7	160.3	449.2	4.00	6.92	1.70
C-F	difference	0.2	1.9	-2.4	-0.5	-0.3	0	1.1	-0.08	-0.06	0.02

surface vegetation change.

The use of a single column model for this kind of study has limitations: any changes in the large scale flow and associated feedbacks on surface quantities will not be represented. The importance of such changes was demonstrated in a recent study by Sud et al. (1988). They showed that a reduction in the surface roughness length can produce a large change in the horizontal convergence of water vapour in the boundary layer and a corresponding large change in the rainfall distribution. An experiment with the Met. Office AGCM is underway to study the effects of deforestation in the Amazon basin. A control simulation with the standard surface parameters for the Amazon region is being compared with an anomaly simulation in which tropical forest and savannah in the control are changed to pasture. Some further simulations to investigate the impact of the individual surface parameter changes are also being done.

6. SUMMARY AND CONCLUSIONS

The importance of a realistic treatment of land surface processes for simulating present day climate and possible climate change senarios has been demonstated in a series of experiments using the UK Meteorological Office AGCM. The use of spatially varying land surface parameters derived from a detailed soil and vegetation type dataset has improved aspects of the model's simulation.

There has been much interest recently in the climatic effects of human activities such as burning fossil fuels and cutting down tropical rainforests. The latter obviously affects the climate directly through changes in the properties of the land surface. Experiments using both a single column model and the AGCM with different surface parameter specifications have shown the substantial impact these can have on rainfall, soil moisture, surface temperature and the surface fluxes of radiation and sensible and latent heat. The response of the soil moisture to increased amounts of atmospheric CO_2 has also been demonstrated to be sensitive to the representation of runoff. Other land surface processes may also be important in determining the local, if not the global, changes in surface and boundary layer climate; further model experimentation and maybe development are required.

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