SIGNIFICANCE OF LAND SURFACE PROCESSES AND THEIR ROLE IN THE ATMOSPHERIC GENERAL CIRCULATION

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Summary: The main ways in which land surface processes and characteristics influence the atmospheric circulation are considered. These are their modulation of the available energy via the surface radiative properties, their control of the partitioning of this energy between evaporation and sensible heat flux, and the frictional drag applied to the flow by surface roughness. Experiments demonstrating these sensitivities of the modelled circulation are described.

1. INTRODUCTION

The interactions between the land surface and the atmosphere have a major role in processes in both media. These interactions are strongly affected by land surface processes (including vegetation and soil characteristics). This paper focusses on the role of these processes in the atmospheric general circulation. In this context, we are mainly concerned with how they modify the fluxes of heat, moisture and momentum between the surface and the atmosphere, and how these modifications influence the atmospheric circulation. The topics we need to consider are thus how the land surface determines:

(i) the amount of solar radiation absorbed;

(ii) the surface temperature and the longwave radiation emitted into the lower atmosphere;

(iii) the partitioning of the net surface downward radiation between the upward sensible and latent turbulent heat fluxes and the heat flux into the ground.

(iv) the drag on the atmosphere by surface friction.

(The gravity wave drag generated at the surface is considered in a separate paper.)

2. SURFACE RADIATION AND ALBEDO

Table 1 describes the energy balance of the atmosphere and surface. The data are based on results from the Meteorological Office Hadley Centre (HC) atmospheric general circulation model

(AGCM) (strictly an atmosphere-land model) run in climate mode with climatological sea surface temperatures; they are within a few W/m^2 of the estimates from observations by Ramanathan et al. (1989), shown in parentheses in the table. In the global mean, the surface net radiative flux is composed of about 165-170 W/m^2 net solar flux (about 200 downward less 35 reflected) less 55-60 W/m^2 net longwave flux; these figures are also typical for land near 50^oN in summer. In the global mean, energy is provided by the surface fluxes to balance the atmospheric budget against the net atmospheric radiative cooling of about 110 W/m^2. The boundary layer and convective processes then distribute the energy through the atmosphere. A change in albedo of up to 10% such as can occur with a change in vegetation from forest to arable land or pasture can thus modify the available energy by about 20 W/m^2. More importantly, perhaps, it can change the top of atmosphere (TOA) radiative balance by a similar amount in what must on average be a zero net flux.

Table 1: Global heat balance based on 10 year annual mean from HC model (also, in parentheses, estimates from observations according to Ramanathan et al. (1989))

	RADIATION				TURBULENT FLUX		
	SOLAR		LONGWAVE		SENSIBLE	LATENT	SNOWMELT
	DOWN	UP	DOWN	UP			
TOA	342 *(342)*	108 *(105)*	0 *(0)*	235 *(237)*			
ATMOS HEATING	68 *(68)*		-181 *(-174)*		22 *(16)*	91 *(90)*	
SURFACE	166 *(169)*		-54 *(-63)*		-22 *(-16)*	-90 *(-90)*	-0.3

It was this realisation, of the role of surface albedo in the system energy balance for a region, which prompted Charney's seminal paper on the dynamics of deserts (Charney, 1975). Charney identified a potential positive feedback by which a decrease in rainfall could lead to a decline in vegetative cover so exposing highly reflective sandy soils; the consequent increase in albedo reduced the TOA net downward radiation and could result in a deficit in the column energy balance - indeed he had noticed that according to satellite observations the energy balance for the (surface+atmosphere) column over the Sahara was in deficit even in summer. To maintain temperature gradients small enough to be stable in the tropics, warming by descent is then required. This descent dries the air so reducing rainfall and maintaining the sparsely vegetated semi-arid surface conditions.

The reality of this mechanism in models has been demonstrated in many experiments since Charney's paper was published. Such effects are often most clearly shown by simple global change experiments designed to assess sensitivity. One such experiment was that reported by Carson and Sangster (1981). They investigated the precipitation changes caused by a global decrease in land surface albedo from 0.3 to 0.2. They found large changes, with increases exceeding 5 mm/day over some land and compensating reductions over the adjacent ocean. These were clearly associated with changes in vertical velocity in the sense expected by Charney (Rowntree, 1983). Similarly clear modelled responses to albedo changes have been found over the semi-arid tropics by Sud and Fennessy (1982), Laval and Picon (1986), Planton (1986) and Cunnington and Rowntree (1986).

Such changes may also be important in regions subject to deforestation. The tropical forests of Amazonia are the prime example of this as the largest remaining area of forest. Most recent experiments on the effects of deforestation have applied a set of land surface changes to represent several aspects of the change - reduced roughness and root depths as well as increased albedo. Lean and Warrilow (1989) and Mylne and Rowntree (1992) isolated the albedo effect in experiments of 8 or 9 months. Lean and Rowntree (1995) (see also Lean et al. (1995)) have recently run a similar experiment with the Hadley Centre (HC) model in association with our latest series of deforestation experiments and taken this to 5 years.

The albedo change was based on observations made as part of the ABRACOS (Anglo-Brazilian Amazonia Climate Observation Study) collaborative project; it is about 0.05 though with seasonal variations. The Charney mechanism is confirmed by the changes in vertical motion (Fig. 1) with a general trend towards decreased ascent over the deforested area. The map of annual mean precipitation changes (Fig. 2(a)) (with statistical significance in Figure 2(b)) shows a large area with over 0.5 mm/d decrease in rainfall. This is a change of about 10% in a region with over 5mm/d mean rainfall both in the modelled and the observed; a decrease of about 2A% for an A% albedo change has been typical of such albedo change experiments (Mylne and Rowntree, 1992).

Changes in forest cover in higher latitudes such as the boreal forests of Canada, Scandinavia and Siberia can also affect the albedo. Though the likely future changes in extent of these forests are less than for tropical forests, the potential changes in albedo are larger: because of windblow and sublimation, snow tends not to cover trees for long after a snowfall, so that typical albedos when snow is lying are much larger for grassland and tundra (0.8-0.9) than for forest (0.2 to 0.3) (Robinson and Kukla, 1985). Forests thus have a marked warming influence on higher latitude climates particularly in spring when there are large amounts of solar radiation and the snow cover is still extensive (Bonan et al, 1992; Thomas and Rowntree, 1992; Chalita and Le Treut, 1994). The impacts obtained were very large in some of these experiments. In one by Bonan et al., in which the model included an interactive ocean so that the ocean was free to respond to the changes induced in the atmosphere, the zonal mean cooling near 60^{o}N over land was 5K or more throughout the year and 12K in April. Effects on the atmospheric circulation have not been documented.

3. PARTITIONING OF AVAILABLE ENERGY

The second impact of land surface characteristics to be considered is the effect on the partitioning of the available energy. The energy balance (Table 1) shows that typically about 90 W/m2 of the 110 W/m^2 available enters the atmosphere as evaporation, the rest as sensible heat.

There is a clear case for expecting changes in evaporation to affect precipitation, as the consequent change in relative

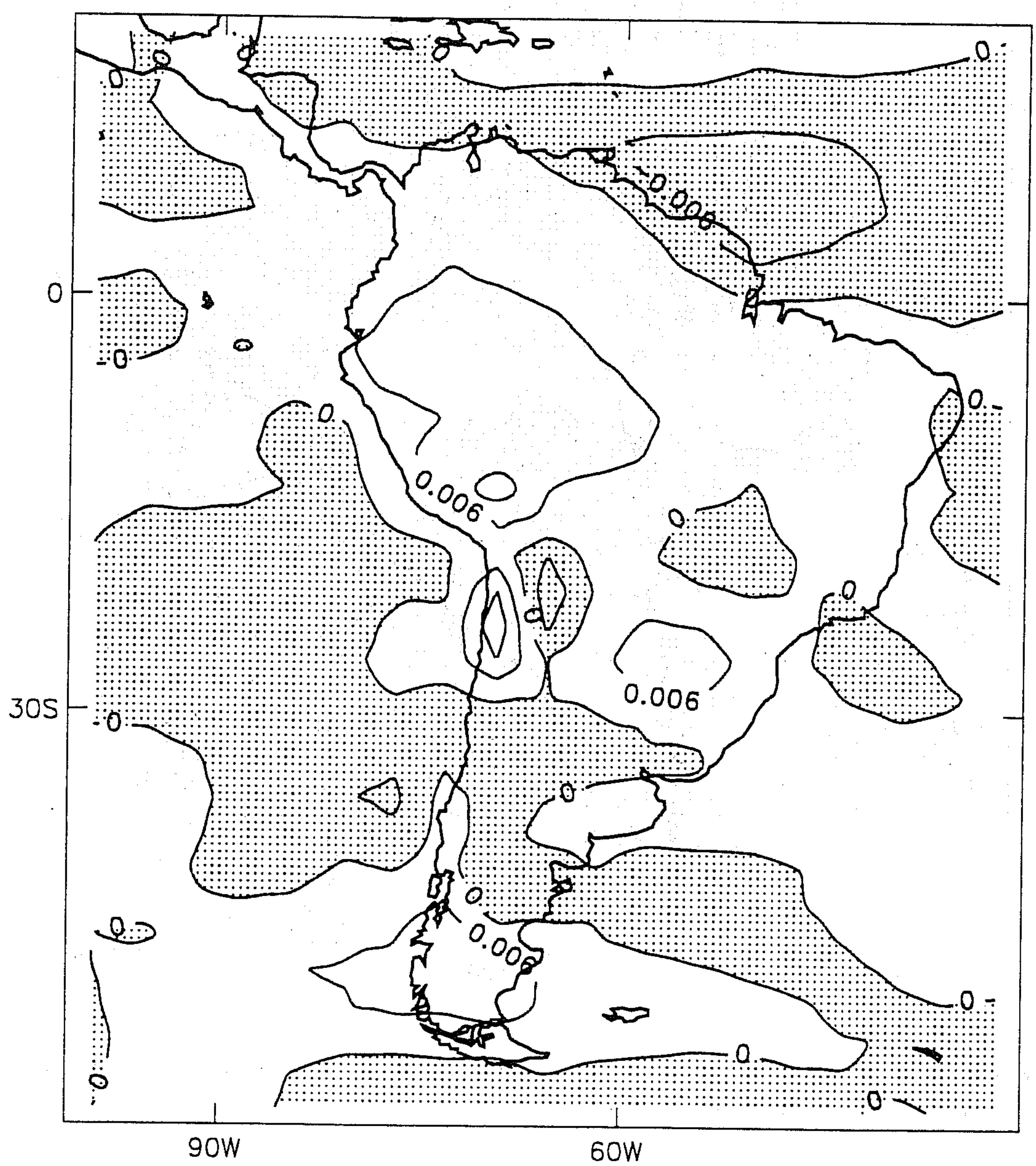

Figure 1: Changes in vertical motion at 500 hPa caused by increases in albedo (averaging 0.05) over the tropical forest regions of northern South America. Areas of increased ascent are shaded (contour interval: 0.006 Pa/s).

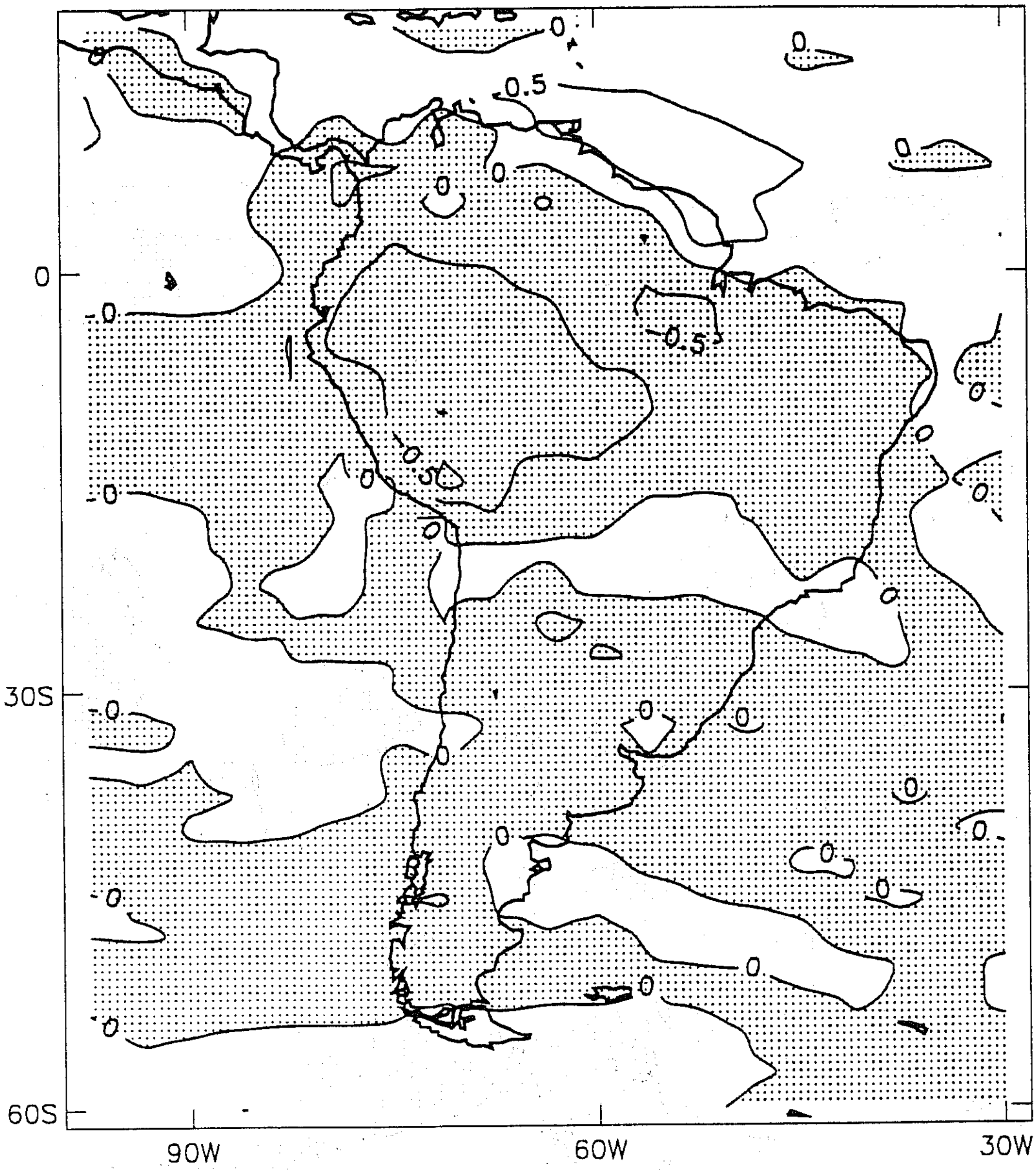

Figure 2(a): As Figure 1 for annual mean rainfall (contour interval 0.5 mm/day).

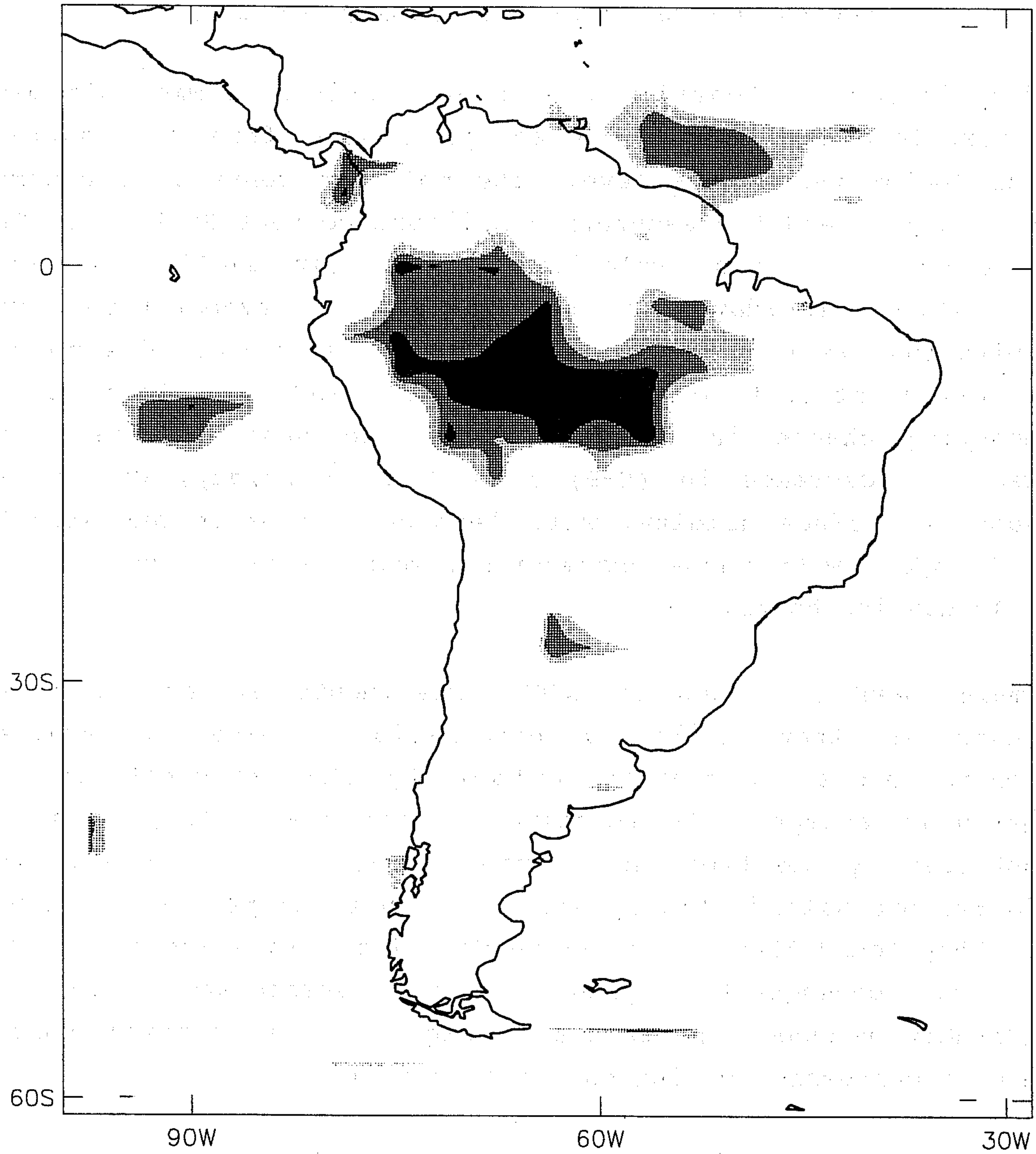

Figure 2(b) Statistical significance for rainfall changes shown in (a) (light, medium and heavy shading for areas significant at the 90, 95, 99% levels).

humidity is large. Consider a moist mid-latitude summer air mass with turbulent and convective energy fluxes from the surface confined below the 500hPa level; the moisture content is assumed to be 23 kg/m^2, and the saturation moisture content 30 kg/m^2. The rate of change in mean relative humidity (in day^{-1}) due to a change δ(E-P) (mm/day) in the excess of evaporation over precipitation, is 0.06*δ(E-P). About half of this is due to the direct effect of (E-P) on moisture content, and about half to the accompanying change in temperature as the sensible heat flux changes. A decrease in (E-P) from 3 to 1 mm/day, due to an increase in surface aridity, will decrease relative humidity by about 12 %/day with these assumptions, most of this occurring in about 12 daytime hours.

The main feedback found in AGCM experiments in extratropical latitudes is that increased evaporation gives a moister atmosphere and so a greater probability that precipitation is triggered by ascent. On relatively short timescales, this was demonstrated by Rowntree and Bolton (1983). They compared two forecasts, one with initially arid soil over Europe, one with wet soil. They found that on the sixth day of the forecast from a dry soil, the consequent drying of the atmosphere (Fig. 3) considerably reduced the rainfall (Fig. 4) over central Europe ahead of a depression moving east from the Atlantic.

A very clear rainfall response is also clear in many other sensitivity experiments including those comparing the effects of a dry and wet surface for all the continents (Shukla and Mintz, 1982). In their dry experiment, the surface temperature field exhibits a strong warming as all the available radiation goes into sensible rather than latent heat flux. With the wet soil, rainfall is increased over almost all land from generally very low values for the arid case. One exception is NE India where there is increased southwesterly flow onto the Himalayas from the ocean giving a large increase in rainfall. This increased flow - amounting to an intense Asian monsoon where there was none in the wet case - is a response to the circulation changes generated by a much warmer lower troposphere with a dry surface. Rainfall over

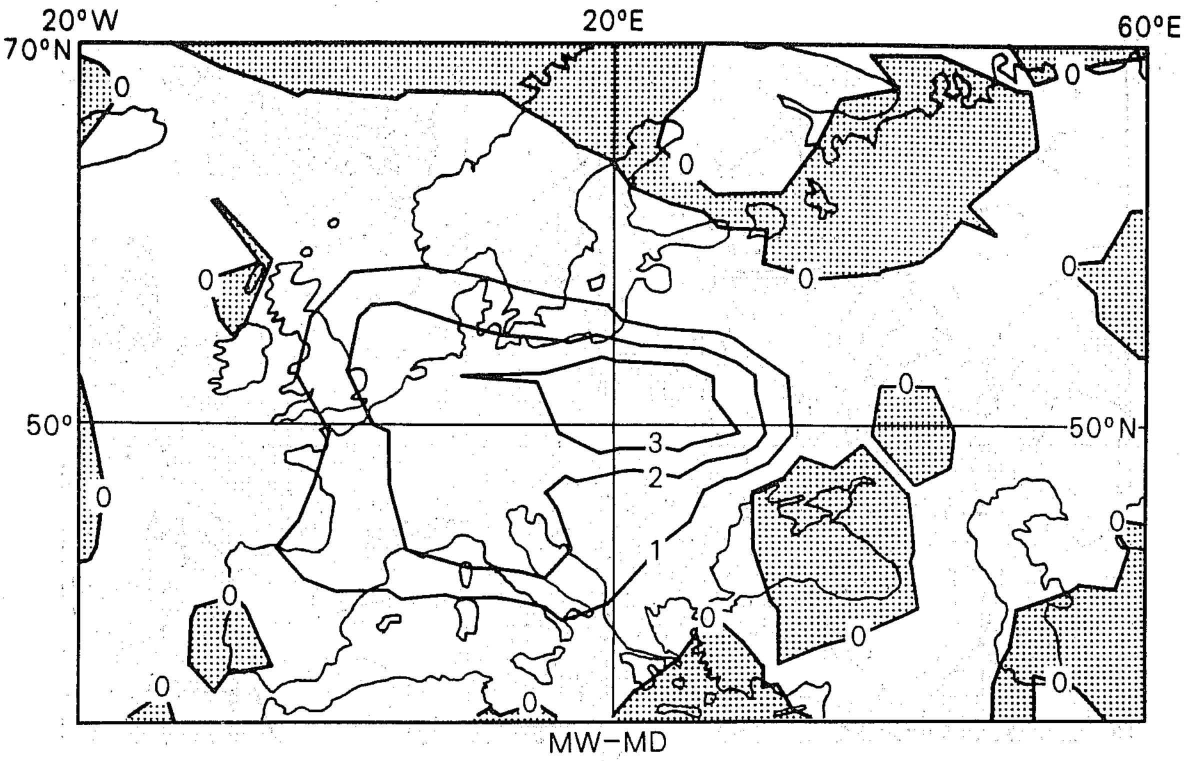

Figure 3: Difference (WET-DRY) in specific humidity at Day 5 at about 900 hPa between experiments for July starting from saturated (WET) and arid (DRY) conditions over Europe from 42 to 54°N west of 30°E. (contour interval 1 g/kg, with decreases shaded).

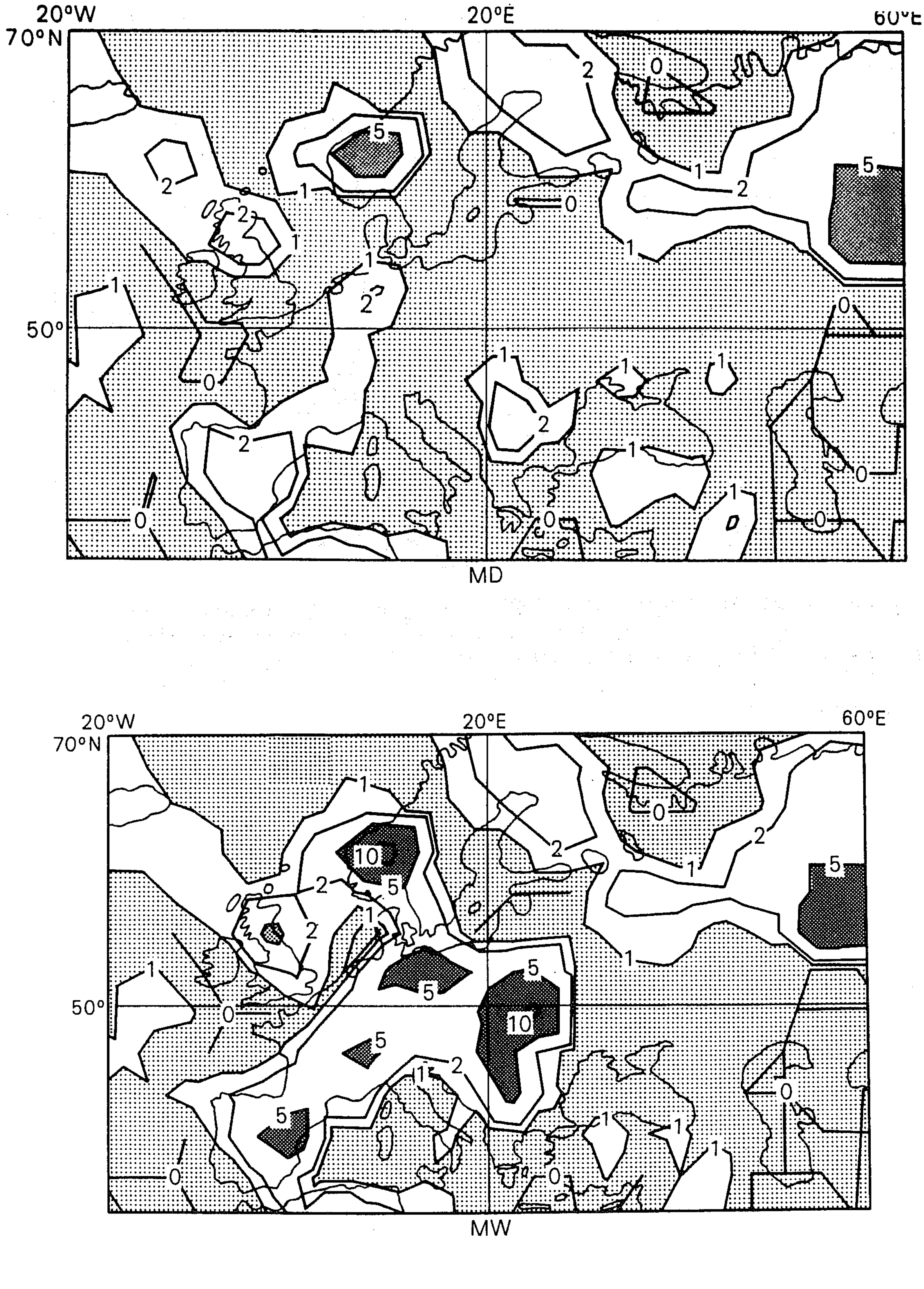

Figure 4: As Figure 3 but rainfall on the sixth day of experiments with initially (a) dry and (b) wet soils. (light stippling for <1 mm, heavy stippling for >5mm).

northern South America changes little, increased moisture convergence compensating for the decrease in evaporation. A similar compensation has been found in some recent experiments to assess the effects of deforestation of Amazonia (Polcher and Laval, 1994; Lean and Rowntree, 1995). Lean and Rowntree find that a reduction in evaporation, resulting from decreased surface infiltration, increased the rainfall in some areas.

The effects of partitioning on the vertical circulation can be very marked as one might anticipate from Shukla and Mintz's expt. The analysis by Walker and Rowntree (1977) of a zonally symmetric model of a desert - wet land - ocean geography similar to that of west Africa with cyclic eastern and western boundary conditions demonstrated this. The only difference between the two experiments was the initial soil moisture - dry for the "desert" region in one, wet in the other. The experiments showed a tendency for the initial dry-wet contrast to be self maintaining, with marked differences also in the vertical circulations (Fig. 5). In the dry case, there was ascent over the strongly heated desert but it was shallow with little vertical motion above $\sigma=0.5$ (about 500hPa); in contrast the wet experiment exhibited a deep circulation up to about $\sigma=0.2$ (about 200hPa). This can be understood in terms of an extended Charney mechanism as discussed by Cunnington and Rowntree (1986): radiative cooling of the upper troposphere must be balanced by warming through subsidence unless there is a heat source from condensation; without the surface moisture supply, the import of moisture does not moisten the atmosphere sufficiently to generate the deep convection needed to give latent heat release. (Note that the averages used by Walker and Rowntree are for a short period, only 5 or 10 days. One can learn a great deal from such short runs - something not always appreciated by those starting on studies of the general circulation in research institutes where 10 year experiments can be run in a week or so!)

Evidence on the impact on the horizontal circulation is provided by similar experiments (Cunnington and Rowntree, 1986) with a global model, starting with different atmospheric humidities (one

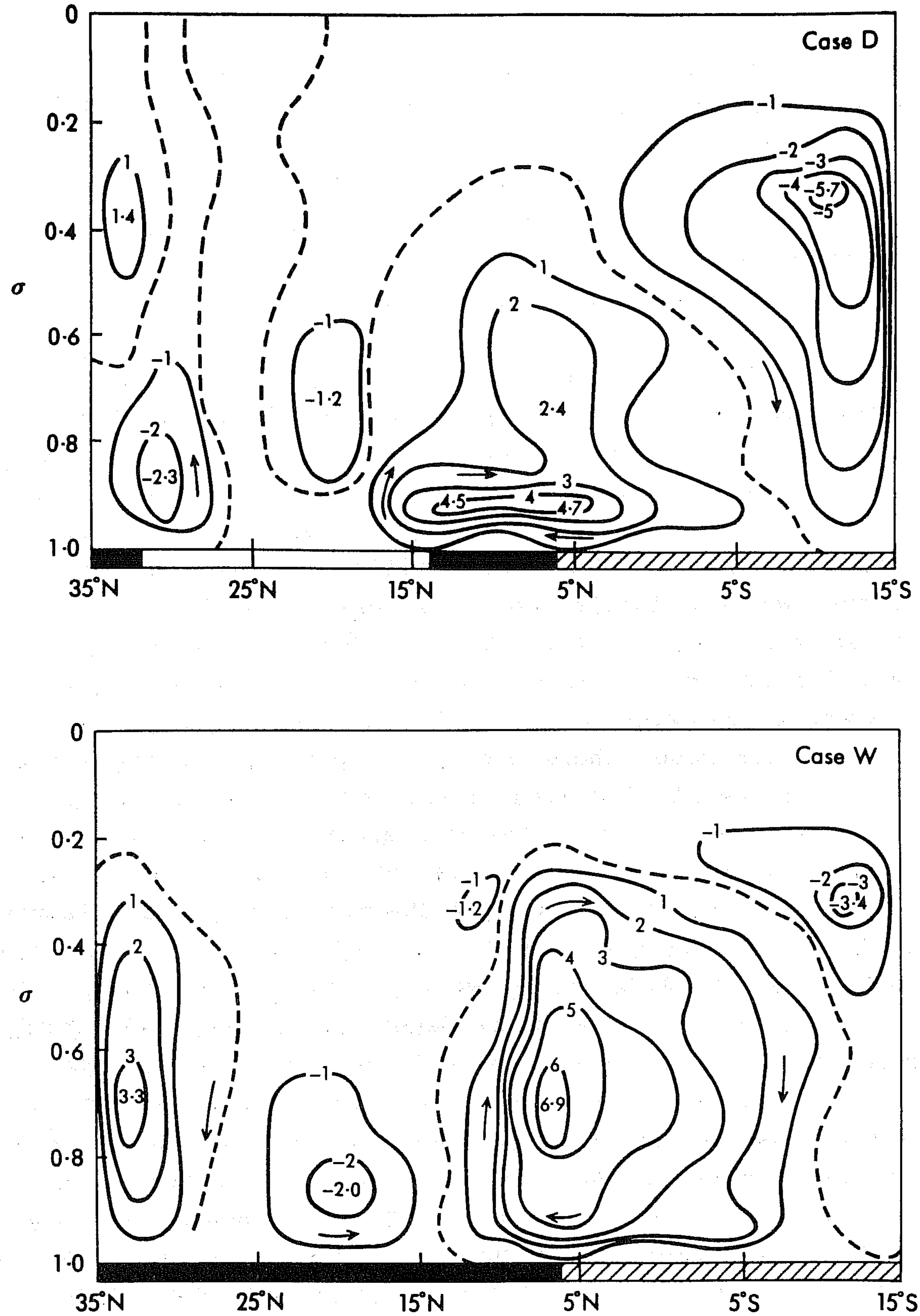

Figure 5: Streamlines of meridional circulation (10^3kg m^{-1} s^{-1}) averaged for Days 5 to 10 in a zonally symmetric model for initially dry (Case D) and wet (Case W) soil at 14-32^{O}N. Surface types are ocean south of 6^{O}N, land to north. The vertical coordinate σ is pressure normalised by surface pressure.

realistically dry, the other wet) over the Sahara and Sahel. The consequent enhancement of rainfall persisted throughout the 50 days of the experiment. Very different 700hPa circulations developed (Fig. 6), with a realistic anticyclonic circulation in the dry experiment, in contrast to a cyclonic circulation centred near 20^{O}N 0^{O}E in the wet experiment. The upper tropospheric circulation was also much changed (Fig. 7) with reduced easterlies over much of the African and Atlantic tropics, a change apparently typical of weakened tropical heat sources (Cunnington and Rowntree, 1986).

4. SURFACE ROUGHNESS CHANGES

The effects of decreases in surface roughness have been explored by Sud and Smith (1985a,b) and Sud et al. (1988). Sud et al (1988) found that cross isobaric flow was decreased giving less frictional convergence and less rainfall over continental cyclonic regions. In regions with divergence, one might expect the opposite effect, but such areas do not have much rainfall. Recently, Sud et al. (1993) have argued that the deepening of the boundary layer with increased roughness is likely to give increased rainfall.

Another relevant mechanism may be that as boundary layer flow is increased with decreased roughness, convergence is likely to be increased where flow is towards a topographic barrier; this is believed to be the explanation for the results from a recent Hadley Centre Amazon deforestation experiment where rainfall in western Amazonia is enhanced when roughness is decreased (Fig. 8) (Lean and Rowntree, 1995); it is also consistent with the result for the Bangladesh region obtained by Sud and Smith (1985b) where rainfall was increased where roughness was reduced in the region of upslope flow. However, Sud et al. (1988) obtained decreases in rainfall over northern S America, with increases over the ocean west of the Andes. A possible reason for the contrast with Lean and Rowntree (1995)'s result may be the higher elevation of the Andes in the HC model (from 1500m at 5^{O}N to 4100m at 15^{O}S) than in the coarser resolution model used by Sud et al. (1988).

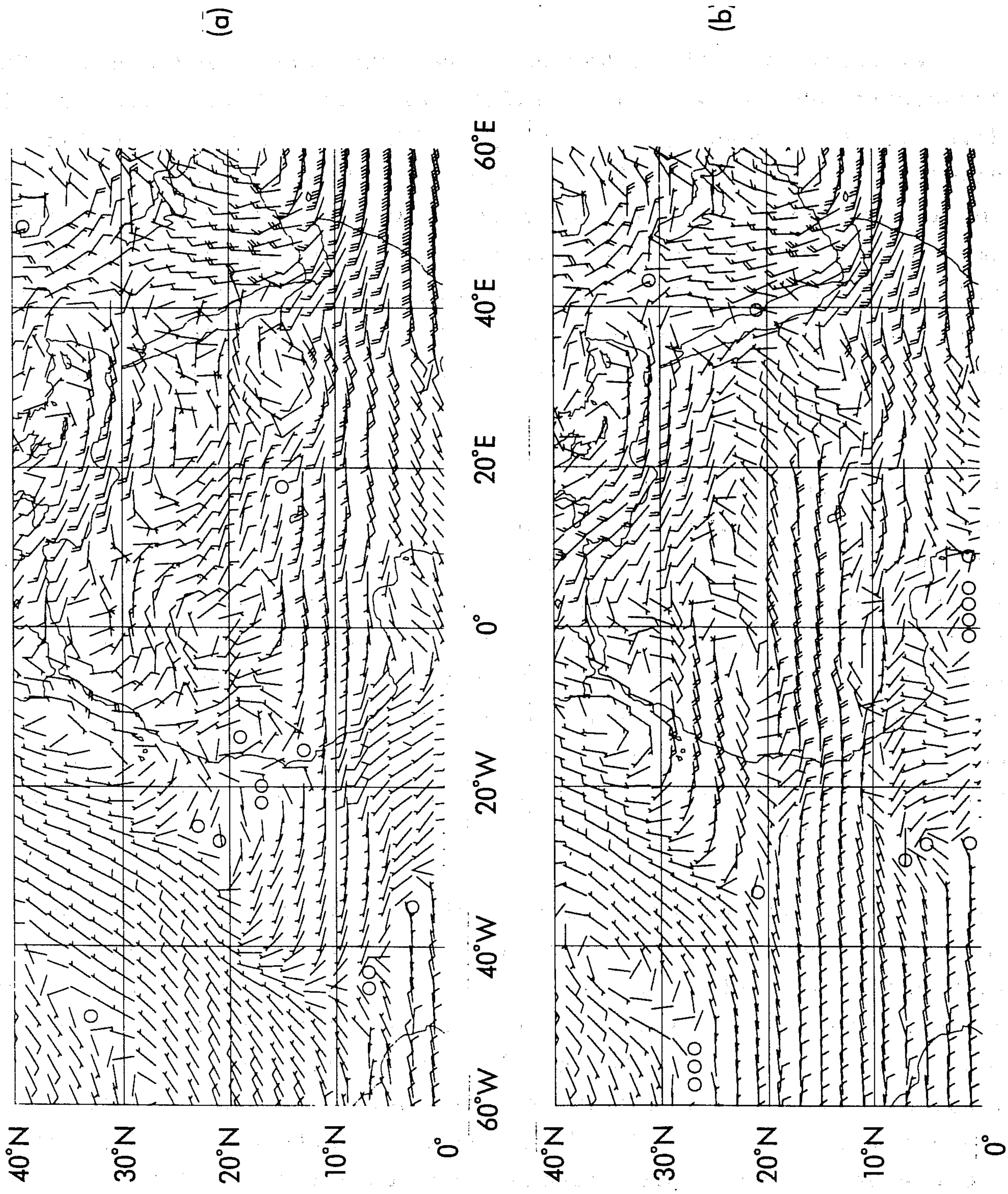

Figure 6: 700hPa circulations averaged over Days 21-50 for July experiments with initially (a) moist and (b) realistically dry atmosphere over Sahara. Each full feather denotes 5 m/s.

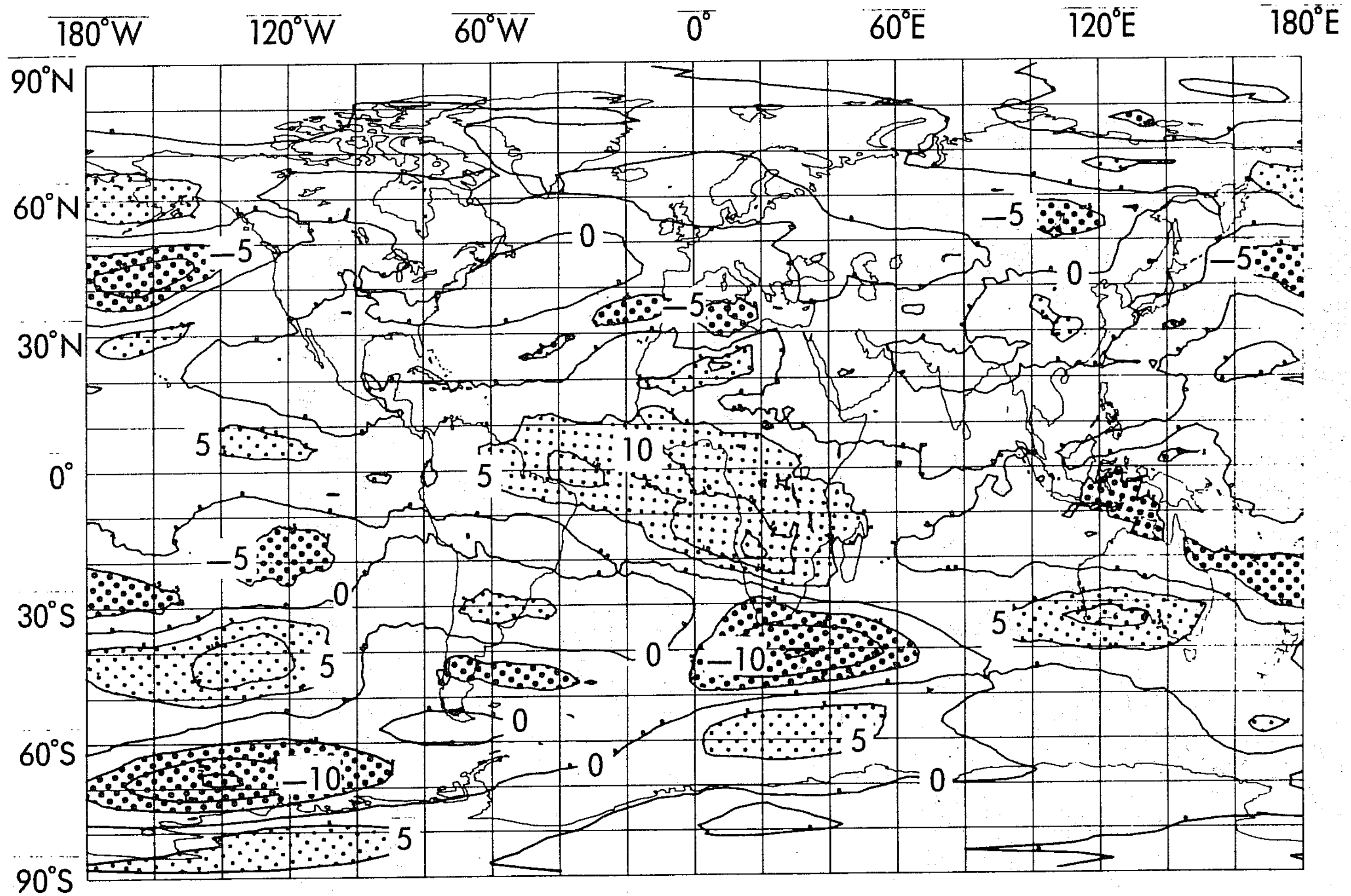

Figure 7: Difference of 200 hPa zonal wind in dry relative to wet case for same experiments as in Figure 6. Light stippling for increases of more than 5 m/s and heavy stippling for decreases of more than 5 m/s.

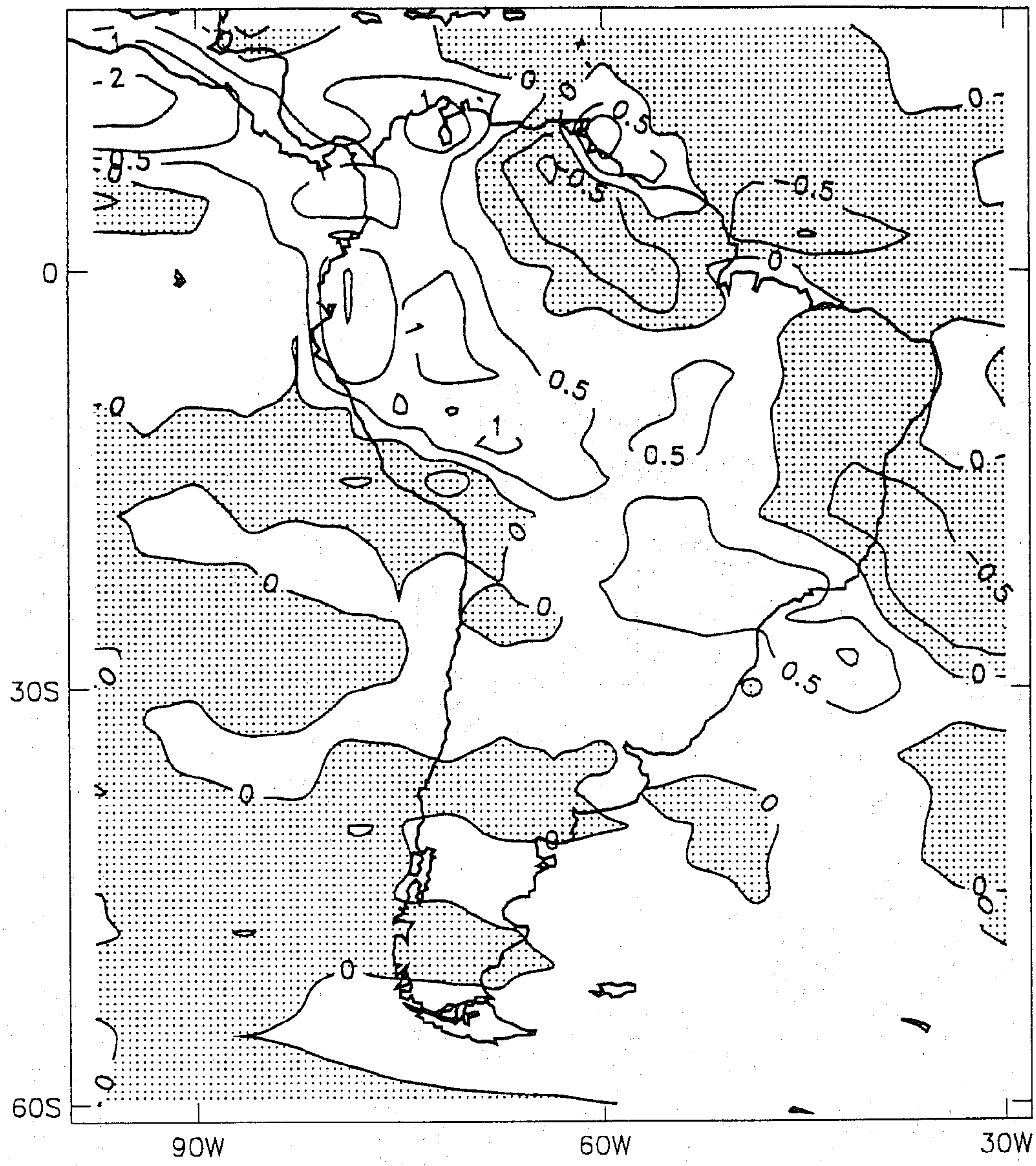

Figure 8: Five-year mean effect on rainfall (mm/day) of decrease in surface roughness from 2.1m to 2.6cm over rainforests of northern South America. Shading shows areas of decreased precipitation (contours at 0, ±0.5, 1, 2 mm/day).

5. RELATIVE IMPORTANCE OF LAND SURFACE CHARACTERISTICS

I have discussed three basic sensitivities of the circulation to the land surface. The first and last refer clearly to specific characteristics - albedo and roughness length for momentum transfer. The second however appears as a sensitivity to evaporation triggered by changes in surface moisture availability. These may be due to any of several different parameters or processes. These include soil hydrological characteristics affecting the soil moisture availability at the roots, the leaf structure affecting the water held on the canopy (important because it evaporates with zero r_S) and plant physiological factors and meteorological parameters affecting the transpiration.

Some insight into the mechanisms behind the partitioning can be gained from study of the Penman-Monteith equation as discussed by Rowntree (1991) . The equation can be derived from the equations controlling the sensible heat flux and evaporation:

$$H = \rho c_p \delta T \,/\, r_{AH} \qquad (1)$$

$$E = \rho\, \delta q \,/\, (r_{AE} + r_S) \qquad (2)$$

where r_{AH} and r_{AE} are aerodynamic resistances between the air at a reference height z_r and the air very near the surface, and r_S is an additional stomatal resistance between the interior of the leaf and the leaf space. δT is the temperature difference between the surface and a reference level (typically a few metres) and δq the difference in specific humidity between the leaf pore space and the same reference level. By eliminating the surface temperature between these equations and the surface heat balance equation, one can obtain the Penman-Monteith equation:

$$\frac{LE}{(R_N - G_*)} = \frac{1 + \rho\, c_p\, \delta' q \,/\, [(R_N - G_*)\, r_A\, \Delta]}{1 + (c_p / L\Delta)(1 + r_S/r_A)} \qquad (3)$$

Here, Δ is the gradient of the saturation specific humidity q_S with temperature and $\delta' q$ is the vapour pressure deficit $(q_S(T_A) - q_A)$ at the atmosphere's lowest or reference level. Figure 9 shows results of using equation (3) for a typical tropical or

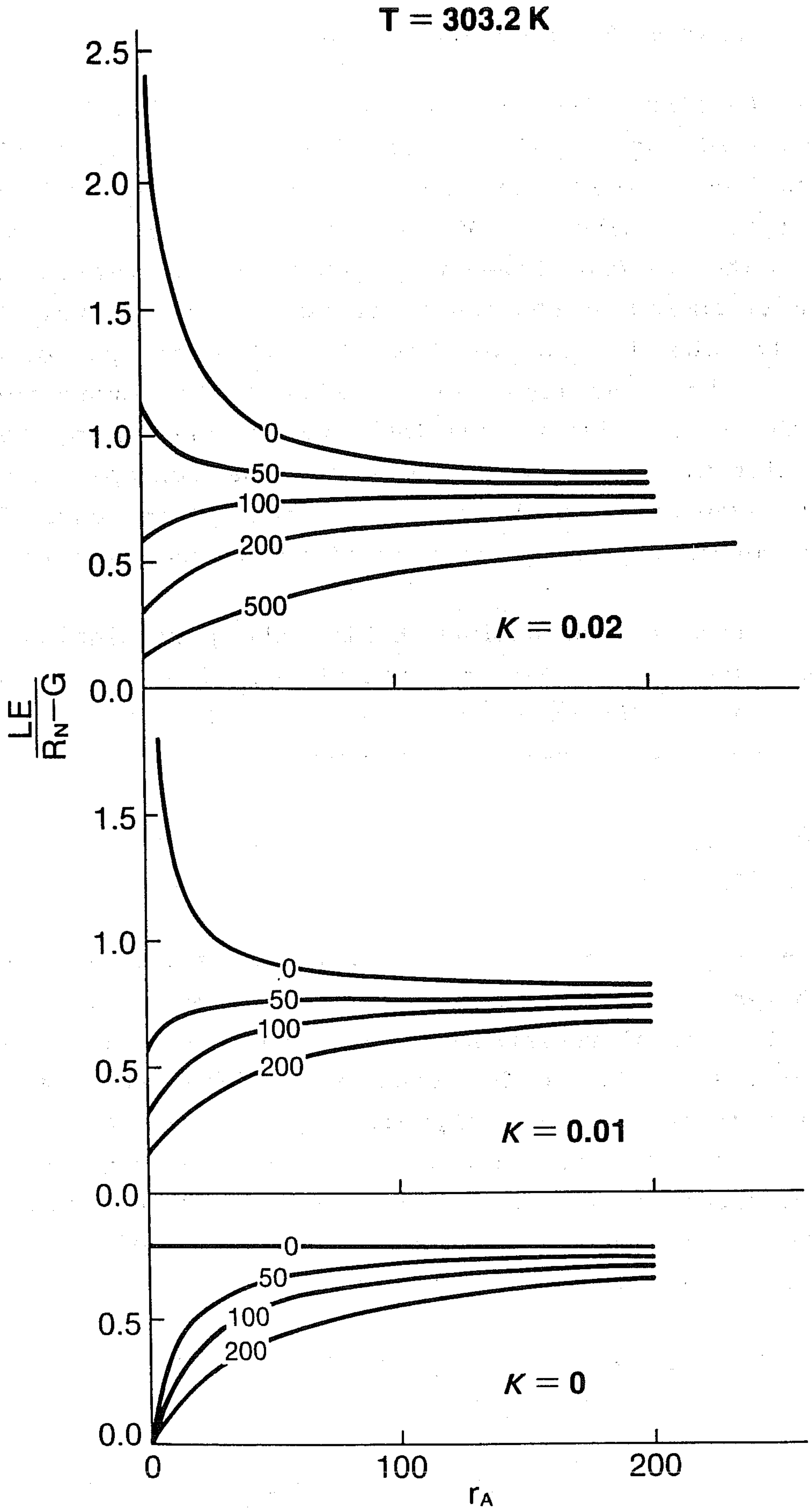

Figure 9: Dependence of evaporative fraction $LE/(R_N-G_*)$, for a temperature of 303.2K, on r_A from 0 to 200 s/m and r_S (0, 50, 100, 200, 500 s/m (500 s/m for κ=0.02 only)) for three values of $\kappa = \delta' q/(R_N-G_*)$ (κ in $10^{-3}W^{-1}m^2$).

mid-latitude summer daytime temperature of 303K, for three values of the parameter $\kappa = \delta'q/(R_N-G_*)$. The largest of these ($0.02*10^{-3}$) is typical of daytime conditions in the absence of limitation of transpiration by soil moisture deficits. Of particular interest is the existence of an asymptotic value of $LE/(R_N-G_*)$ approached in a number of contexts, including as r_A increases, and as $\delta'q$ tends to zero for $r_S=0$. This value is temperature-dependent, varying from about 0.56 at 283K to 0.79 at 303K.

Sensitivity of the partitioning may be expected from any surface parameter. However, it is not straightforward to infer the dependence of evaporation on particular parameters such as albedo or roughness length from the Penman-Monteith equation because the variations in other parameters such as temperature or vapour pressure deficit may not be known. A particular problem is that of boundary layer feedbacks; for example, an increase in evaporation will tend to moisten the boundary layer and so modify $\delta'q$, while a decrease in sensible heat flux will cool the boundary layer and so modify Δ and ρ.

This can be illustrated by comparing some results from experiments using a single column version of a general circulation model on the effects of a change in albedo, with the result from equ. (3) obtained assuming that temperature and vapour pressure deficit are unchanged, as in equation (2.26) of Rowntree (1991). The single column model (SCM) version of the HC model was used with radiative forcing appropriate for southern England in July, and results were taken from the mean of the first 10 days of experiments with different surface albedos (other details in Table 2). The results (Table 2) show that a change in albedo from 0.3 to 0.1, increasing the available energy, leads to a considerable increase in $\delta'q$ which enhances the increase in evaporation. The fraction $d(LE)/d(R_N-G_*)$ of the change in available energy which is used for evaporation is around 70% compared to the 44% estimated assuming $\delta'q$ to be constant. The figure of 70% is close to the evaporative fraction $LE/(R_n-G^*)$ of around 75% for these conditions.

Table 2: Effects on latent heat flux of a change in albedo estimated using single column version of HC model for southern England in July, compared with calculation assuming constant temperature and $\delta' q$ from Rowntree (1991) (R91).

ALBEDO	0.1	0.2	0.3
R_N (W/m^2)	167.7	140.6	113.7
$R_N - G_*$ (W/m^2)	163.0	137.6	112.5
LE (W/m^2)	120.2	102.3	85.4
H (W/m^2)	42.8	35.3	27.1
$\delta' q/10^{-3}$	3.12	2.73	2.41
LE/($R_N - G_*$)	.737	.743	.759

d(LE)/d($R_N - G_*$)	.705	.673
R91 (Eq 2.26)	.445	.433

Notes: Assumes geostrophic wind of 10 m/s, r_s = 79 s/m, z_0 = 0.02 m, zero cloud cover.

Another result from experiments with the HC SCM is that as the stomatal resistance increases, the vapour pressure deficit increases - by a factor of 4 from 1.1 g/kg for r_s=0 to 4.5 g/kg for r_s=200s/m for the conditions given in Table 2 with an albedo of about 0.2. The formulation of r_s used in these experiments did not include any dependence on meteorological parameters. If, as generally observed (e.g. Turner, 1991), r_s increases with vapour pressure deficit $\delta' q$, and $\delta' q$ increases, as these experiments show, with r_s, there is potential for a strong positive feedback.

6. EFFECTS OF REGIONAL CHANGES IN LAND SURFACE TYPE

Actual land surface changes are a combination of changes in most or all of the above. Regions for which changes have been given particular attention include the Sahel region of North Africa, and the tropical rain forests, especially Amazonia.

Effects on climate of the possible desertification of the Sahel, involving reduced moisture availability and increased albedo, were assessed by Rowntree and Sangster (1986), who found a widespread decrease in rainfall - that in the western Sahel was mainly due to the reduced evaporation, and that in the eastern Sahel to increased albedo. More recently Xue and Shukla (1993) modelled the effects of a change in vegetation from savanna and broadleaf shrubs with ground cover to "shrub with bare soil", and obtained a similar result with widespread reductions in rainfall between 10 and 20^{o}N; they demonstrated a strong resemblance to the differences between exceptionally dry and wet observed years in the Sahel except that observed decreases extended to the equator over east Africa. However, a strong case has also been made for a major role for sea surface temperature (SST) anomalies in such rainfall variations (Rowell et al., 1995), with a relatively minor role for the land surface. In the absence of strong evidence that variations in land surface characteristics in association with Sahel drought have actually occurred, it remains possible that both theories are correct, with the drought of recent decades caused primarily by SSTs.

There have been several experiments in which the effects on climate of deforestation of Amazonia have been estimated (recent ones include Nobre et al. (1991), Lean and Rowntree (1993), Pitman et al. (1993) and Polcher and Laval (1994)). Deforestation is expected to affect not only vegetation characteristics such as albedo, roughness, the capacity of the canopy to intercept rainfall, the stomatal resistance and the root depth, but also soil characteristics. The ABRACOS project has provided data on each of the vegetation-dependent parameters, and also an indication that maximum surface infiltration rate is decreased.

Experiments with the HC model have demonstrated strong sensitivity to albedo, roughness and the maximum infiltration rate, with each contributing to the total effect in quite different ways, as discussed in earlier sections. As with previous investigations, the experiment with all the changes included exhibits substantial decreases in rainfall, but there are also increases over the western part of the forested regions which can mainly be attributed to the decreased roughness (see Section 4) (Lean and

Rowntree, 1995). Unlike most such experiments, the moisture convergence is increased; this is attributable to the decrease in evaporation due to reductions in the maximum infiltration rate. These results are consistent with results recently reported by Dirmeyer and Shukla (1994) who found that increases in rainfall can result from deforestation scenarios in which the albedo increase is less than 3%. However, the regional distribution of rainfall changes differs substantially from those in the HC experiments. This difference is perhaps not surprising as there are large differences between the control rainfall simulations with the two models.

7. CONCLUSIONS

The results on the effects of albedo, moisture availability and roughness may be summarised as below.

(i) Increases of albedo during seasons with substantial insolation reduce ascent, moisture convergence and precipitation over the modified area. Evaporation is reduced because of the reduction in available energy and, where soil moisture is limiting, because precipitation is less.

(ii) Reduced moisture availability generally reduces precipitation. In the tropics, the replacement of mid-tropospheric latent by near-surface sensible heating modifies the atmospheric circulation, in some situations leading to increased moisture convergence and rainfall.

(iii) Reduced roughness decreases the boundary layer convergence into cyclonic regions. However, the associated increase in boundary layer flow can give increased precipitation if the flow is up the topographic slope.

The effects on climate of possible changes to the actual land surface have mainly been discussed for the Sahel and for tropical forests. These suggest that climate is sensitive to all three of the above types of change. As the nature of these effects can be quite different both in the direction in which they act, and in the region affected, the prediction of the combined effect will require realistic modelling of the actual climate and of the details of the land surface changes.

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