A GCM simulation of the impact of Amazonian deforestation on climate using an improved canopy representation

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SUMMARY

To obtain a good estimate of the impact of Amazonian deforestation on local climate it is critical that the representation of the forest canopy within general circulation models (GCMs) is as realistic as possible. Recent measurements from the Amazonian forest have highlighted major weaknesses in the Meteorological Office GCM simulation of the interception of rainfall from the forest canopy. Here, we present results for Amazonia from a new 3-year control experiment which incorporates an improved representation of micrometeorological processes within the forest. A detailed assessment of the control simulation reveals that the adjusted GCM provides a realistic description of the climate of Amazonia. In determining the impact of Amazonian deforestation on climate we present a comprehensive analysis of the simulated climate following the replacement of Amazonian forest by pasture.

A comparison of the new results with those from the earlier deforestation experiment carried out by Lean and Warrilow (1989) suggests that the reductions in local rainfall (14%) and evaporation (24%) are smaller than those obtained with the previous formulation of interception. It was concluded by Lean and Warrilow that with a wet canopy, decreases in roughness in the deforested case reduce evaporation. With the introduction of the new interception formulation the canopy is less often wet, and so the effect of deforestation is reduced.

1. INTRODUCTION

In the last two decades general circulation models (GCMs) have been used to address the climatic consequences of one of the most serious threats to the global environmental system, namely tropical deforestation. Three recent GCM experiments to assess the likely impact of Amazonian deforestation on climate (Dickinson and Henderson-Sellers 1988; Lean and Warrilow 1989; Shukla et al. 1990) have improved the credibility of such experiments by incorporating a more physically realistic treatment of the land surface. The rationale of these experiments was to represent more satisfactorily the changes in vegetation and soil parameters and the hydrology that accompany such land degradation. However, the applicability of the results obtained from some of these studies is uncertain, since aspects of the simulated forested climate have been shown to be flawed (Dickinson 1989; Shuttleworth and Dickinson 1989).

The version of the Meteorological Office GCM used by Lean and Warrilow (1989) has since been verified against observational data from the Amazonian forest (Shuttleworth 1988); key deficiencies in the simulation of a canopy description for tropical forest were identified. This is particularly pertinent in studying possible climatic implications of deforestation, as the forest canopy has an integral role to play in the land surface/atmosphere moisture exchange. The forest canopy intercepts rainfall and thus regulates the amount of water reaching the surface. The fraction of the canopy that is covered by water then has zero resistance to evaporation, and can evaporate water rapidly into the atmosphere. Evaporation from the dry parts of the canopy will be subjected to stomatal control as well as possibly being limited by available soil moisture.

Following on from this important validation exercise, we report here on the use of an improved canopy formulation which takes some account of the spatial variability of rainfall. Results are presented from a 3-year control experiment. Several aspects of the surface hydrology together with important atmospheric variables are analysed, as it is recognized that deficiencies in atmospheric variables such as net radiation and rainfall

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may conceal improvements made to the treatment of the land surface (Dickinson and Kennedy 1991). The control simulation was validated against a number of independent observations for Amazonia and use was made of some of the most up-to-date data sets.

In assessing the impact of tropical deforestation on climate we will focus on Amazonia, as this region with its expansive forest accounting for some 25% of all tropical forests has an important influence on the atmospheric circulation. Precise evaluations of the fraction of Amazonian forest destroyed to date are disputed, but typical values are modest. For example, Fearnside (1990) estimated that the area cleared by 1988 had amounted to some 8% of the total 5 million km². However, more alarming evidence suggests that explosive rates of clearing are currently taking place over vast areas of Amazonia, which could lead to the total loss of the forest (e.g. Fearnside 1990; Salati et al. 1989).

In order to evaluate the climatic consequences of the complete removal of the Amazonian forest, a 3-year GCM simulation was carried out in which tropical forest and savanna were replaced by pasture in South America, north of 30°S. In a detailed assessment of the results we will consider the influence of this land transformation on several aspects of local climate.

2. DESCRIPTION OF THE MODEL

The version of the Meteorological Office GCM used here is an atmospheric global model (Slingo et al. 1989) which uses the primitive equations to describe and predict the surface pressure, and the atmospheric variables of temperature, wind and humidity at 11 levels between the surface and the top of the atmosphere. The 11 sigma layers (sigma is equal to the pressure divided by its surface value) are spaced unevenly in the vertical to allow for enhanced resolution near the surface and in the upper troposphere. The horizontal resolution at each level is 21° × 31°. The sea surface temperatures are prescribed as the model is run without an interactive ocean; these are taken from climatological values and are updated every 5 days.

Sub-grid-scale processes are treated as follows. Radiative processes are computed interactively every three model hours. The radiation scheme computes diurnally and seasonally varying solar radiation. Both long-wave and short-wave radiation interact with model diagnosed cloud, which has fixed optical properties. A cloud-water variable is now included and large-scale rainfall is diagnosed from the rate of depletion of cloud water (Smith 1990). A mass-flux convection scheme (Gregory and Rowntree 1990) is included that uses a 'bulk' model to represent an ensemble of convective clouds, and aims to describe shallow, deep and mid-level convection. The number of model levels in the boundary layer is allowed to vary up to a maximum of four. Surface turbulent transfer of heat, moisture and momentum are calculated using Monin–Obukhov similarity theory.

Geographical variations of land surface and soil types are represented within the model (Warrilow and Buckley 1989). The vegetation and soil types are based on the work of Wilson and Henderson-Sellers (1985). Their original 1° × 1° resolution primary and secondary vegetation and soil data sets were used to create spatially varying data sets of 15 soil and vegetation parameters which were interpolated onto the 21° × 31° model grid (Warrilow and Buckley 1989). The vegetation parameters and soil parameters (as shown in Table 1) are specified globally in terms of locally derived parameters that were obtained from a review of the literature. Soil-moisture concentration is defined as the ratio of the total volume of moisture to the total soil volume. Available soil-moisture concentration is then the soil-moisture concentration above a threshold value, given by the soil-moisture concentration at wilting point, the point at which it becomes virtually
impossible for plants to remove water from the soil. All parameters are derived at the start of a model experiment and are then assumed to be fixed throughout the run; thus, seasonal variations in the parameters are ignored.

Details of all other aspects of the current land-surface description are given in Warrilow et al. (1986). The surface-hydrology scheme is characterized by a single soil-moisture store. Following the important realization that evaporation of rainfall intercepted by vegetation must be treated independently of transpiration of water vapour (Shuttleworth and Calder 1979), a canopy is also represented. Precipitation is intercepted by the canopy and, while the canopy remains wet, water vapour evaporates from it at the potential or maximum rate. Bare soil is given a canopy capacity to represent ponding of water on a bare soil surface. Transpiration from the dry parts of the canopy is subject to stomatal control and, together with evaporation from the soil, can be limited when the available soil-moisture concentration falls below a critical value. Water falling through the canopy onto the soil infiltrates into the soil at a rate equal to the saturated conductivity, modified to take account of the presence of vegetation. When throughfall exceeds the infiltration rate the surplus of water runs off. Surface runoff takes account of spatial variations in rainfall by assuming that rainfall is only falling over a fraction of the grid box. Within this fractional area rainfall is assumed to be exponentially distributed. Sub-surface runoff is also diagnosed and is controlled by the gravitational drainage of soil moisture. Thermal processes are treated using a four-layer soil-temperature model.

### TABLE 1. Vegetation and soil parameters for the control (NC) and deforested (DF) simulations averaged over Region 1

<table>
<thead>
<tr>
<th>Vegetation parameter</th>
<th>NC</th>
<th>DF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Root depth (cm)</td>
<td>127.4</td>
<td>61.9</td>
</tr>
<tr>
<td>Albedo</td>
<td>0.136</td>
<td>0.188</td>
</tr>
<tr>
<td>Surface resistance (s m⁻¹)</td>
<td>119.7</td>
<td>82.2</td>
</tr>
<tr>
<td>Roughness (m)</td>
<td>0.79</td>
<td>0.04</td>
</tr>
<tr>
<td>Canopy capacity (mm)</td>
<td>0.682</td>
<td>0.633</td>
</tr>
<tr>
<td>Vegetation fraction</td>
<td>0.93</td>
<td>0.84</td>
</tr>
<tr>
<td>Infiltration factor</td>
<td>5.00</td>
<td>1.93</td>
</tr>
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</table>

<table>
<thead>
<tr>
<th>Soil parameter</th>
<th>NC and DF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil-moisture concentration:</td>
<td></td>
</tr>
<tr>
<td>at wilting point</td>
<td>0.143</td>
</tr>
<tr>
<td>at critical point</td>
<td>0.210</td>
</tr>
<tr>
<td>at field capacity</td>
<td>0.345</td>
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<tr>
<td>at saturation</td>
<td>0.475</td>
</tr>
<tr>
<td>Exponent in soil-conductivity</td>
<td></td>
</tr>
<tr>
<td>relationship</td>
<td>9.51</td>
</tr>
<tr>
<td>Saturated conductivity for soil-moisture flow (cm s⁻¹)</td>
<td>$1.111 \times 10^{-4}$</td>
</tr>
<tr>
<td>Volumetric heat capacity (J m⁻³ K⁻¹)</td>
<td>$2.549 \times 10^6$</td>
</tr>
<tr>
<td>Thermal conductivity (J m⁻¹ K⁻¹ s⁻¹)</td>
<td>0.603</td>
</tr>
</tbody>
</table>

### 3. Recent improvements to the model

Measurements, including dry-canopy evaporation and rainfall interception (the water that evaporates from the canopy without reaching the soil), from a tropical rainforest at a site in central Amazonia (Shuttleworth 1988) have provided an unparalleled source of data with which to improve and validate the representation of micrometeorology
within the model. In the light of this study we report on some recent developments to the representation of rainfall interception.

In Fig. 1 we compare monthly interception from a 3-year control experiment (OC) as presented in Lean and Warrilow (1989) with values reported by Shuttleworth (1988). It is apparent that the model significantly over-estimates interception from September to April. This may be partially attributed to the assumption that precipitation falling onto the canopy does so uniformly over the whole of the grid square. Hence, the large spatial variability in rainfall over the area of a grid square, most prevalent during convective rain storms, is neglected.

![Graph showing interception over months]

Figure 1. Simulated monthly mean interception, averaged over the four grid boxes centred near Manaus (60°W, 3°S)(i.e. 56.25-63.75°W, 0-5°S). Observations are for Manaus from Shuttleworth (1988).

Another probable source of error is the value of 2.5 mm chosen for the maximum amount of water that the canopy can hold, but measured as 0.74 mm in the Amazonian forest (Lloyd et al. 1988).

A more realistic formulation was devised (Dolman and Gregory 1992) which is more consistent with Warrilow et al.'s (1986) treatment of surface runoff. Rainfall that is intercepted by the canopy is assumed to fall over a fractional area, \( \mu \), of the total area of the grid square. Within this area the distribution of rainfall is governed by an exponential probability distribution function. For convective and large-scale rainfall \( \mu \) was set to 0.1 and 0.5 respectively. These values were derived from geometrical considerations (Eagleson and Wang 1985).

The model was modified to include this formulation, together with a reduced canopy capacity equal to the measured value of 0.74 mm, and a new control integration (NC) was run for three years. In the following section the performance of this new control simulation is assessed.

4. ASSESSMENT OF THE NEW CONTROL SIMULATION

In this section a comparison is made between some important climatological variables from NC, OC, and observations for Amazonia. This will enable a thorough assessment of the impact of the new canopy formulation together with an assessment of how well the new integration captures reality, and in turn how much confidence we should place in the modelled impact of deforestation on climate. The observations presented in this section represent both climatological averages and data from the 2-year campaign in central Amazonia (Shuttleworth 1988).
(a) Precipitation

It is acknowledged that precipitation is the most important meteorological variable in the tropics and hence arguably the most important variable to model correctly for this type of study. Figure 2 compares the model rainfall with that estimated from observations (Figueroa and Nobre (1990)).

Southern summer. The mean position of the South Atlantic Convergence Zone, situated in western Amazonia and central Brazil at this time of the year, results in a region of high precipitation with a NW–SE orientation. The model captures the position of this well, but magnitudes are over-estimated.

The precipitation maxima on the eastern slopes of the Andes Cordillera in Peru and Ecuador and western Colombia, caused by mechanical lifting of low-level winds by the mountain barriers, are only partly simulated by the model.

Southern autumn. The modelled highest precipitation regions (> 900 mm) agree fairly well with observations. These regions occur along the equator; there are two maxima, one along the Atlantic coast and the eastern portion of Amazonia, and the other over western Amazonia. The modelled western maximum is too intense and the eastern maximum is rather weak, especially along the Atlantic coast, and does not extend far enough west. South of the maxima at the equator, modelled precipitation falls off rapidly and agrees favourably with observations. The modelled precipitation over western Colombia agrees more closely with observations than in southern summer but, probably because of the limitations of resolution, does not have a separate maximum over this area.

Southern winter. The region of maximum precipitation is now displaced even further north, centred at 66°W, 3°N. The model broadly defines this area but the maximum (>1050 mm) is not as extensive as the observed. The decrease in rainfall south of the equator is exaggerated by the model.

Southern spring. The intense precipitation over western Colombia is not as well simulated by the model as in the previous season. On the whole the model tends to present a much noisier pattern of precipitation over northern South America compared with the smoother observed field, although precipitation magnitudes agree broadly. The emergence of the NW–SE orientated band of precipitation from western Amazonia through central Brazil is evident, although this extends too far south. A minimum in precipitation is simulated correctly near the mouth of the Amazon but totals here are too low.

(b) Interception and evaporation

Figure 1 compares monthly values of interception. There is a marked improvement in the simulation of interception relative to OC, although the seasonality is exaggerated by the model, with too little interception from July to September and too much in November and December. These deficiencies can be attributed to an underestimation in rainfall during the southern winter and an over-estimation during early southern summer. The annual mean value of 377 mm now compares much more favourably with the measured value of 328 mm (OC: 709 mm).

Figure 3 shows monthly values of total evaporation. Although the ratio of interception to total evaporation has been improved (now 36% compared with a previous value of 71% for OC and the observed 25%), evaporation is too low from August to November. This is again largely associated with a lack of rainfall during these months, although the low evaporation may itself contribute towards the rainfall deficiency.
Figure 2.  (a) Simulated precipitation pattern of the new control experiment over South America for the southern hemisphere summer (Dec., Jan., Feb.) averaged over the 3 years. Contours are at 0, 30, 60, 150, 300, 450, 600, 750, 900, 1050, 1200 mm, then every 300 mm month$^{-1}$; shading above 900 mm month$^{-1}$. (b) Observed precipitation pattern over South America (Figueroa and Nobre 1990) for the southern hemisphere summer. Contours as for (a) except where labelled differently on the map. (c) As for (a) but for southern hemisphere autumn (March, April, May). (d) As for (b) but for southern hemisphere autumn.
Figure 2. Continued. (e) As for (a) but for southern hemisphere winter (June, July, August). (f) As for (h) but for southern hemisphere winter. (g) As for (a) but for southern hemisphere spring (Sept., Oct., Nov.). (h) As for (b) but for southern hemisphere spring.
(c) Radiation and surface temperature

Figures 4 and 5 show, respectively, the seasonal variation of the net surface radiation and surface temperature for the model compared with observations. The model largely captures the seasonal cycle in net radiation but over-estimates the magnitude by about 30%. This appears to be a common shortcoming of a number of GCMs; for example the version of the community climate model used in the Dickinson and Henderson-Sellers (1988) deforestation experiment over-estimates net radiation by 40% (Dickinson 1989 and Shuttleworth and Dickinson 1989) and it is over-estimated by 20% in the National Meteorological Center's model used in the Shukla et al. (1990) deforestation experiment.

The modelled annual surface temperature of 27.5 °C agrees well with the observed 27.1 °C, but peak values in August and September are exaggerated by the model as a result of the excessive drying of the surface at this time of the year.

Figure 6 compares the modelled radiation at the top of the atmosphere for January and July with data from the Earth Radiation Budget Experiment (ERBE) (Barkstrom 1984) for January 1986 and July 1986 respectively. The model produces a good simulation.
Figure 6. (a) Simulated net radiation at the top of the atmosphere for the new control experiment over South America for January averaged over the 3 years. (b) Observed ERBE net radiation at the top of the atmosphere over South America for January 1986. (c) As for (a) but for July. (d) As for (b) but for July 1986. Contours are at every 20 W m$^{-2}$. 
of the top-of-the-atmosphere radiation for both months. Hence, as the radiation reaching
the surface is over-estimated, this suggests that there is too little atmospheric attenuation
by the model. One probable reason for this is the omission of aerosols that can absorb
a significant proportion of radiation in the boundary layer.

(d) Summary

In summary, the performance of the new control simulation shows a substantial
improvement in modelling interception when compared with the old control experiment
as presented in Lean and Warrilow (1989). We have demonstrated that the model can
produce a fairly realistic simulation of precipitation both spatially and temporally.
Monthly averages of surface temperature and net surface radiation also provide adequate
descriptions of single-site measurements in central Amazonia. Simulated values of net
radiation absorbed at the top of the atmosphere for South America compare favourably
with observations. However, modelled rainfall and evaporation are deficient during the
dry season.

5. DEFORESTED EXPERIMENT

In order to simulate the gross effects of total removal of the Amazonian forest all
global boundary conditions were kept constant except over South America north of 30°S.
Here all tropical forest and savannah were replaced by tropical pasture. Table 1 lists the
vegetation parameters averaged over Region 1 (see Figure 7) for the control (NC) and
deforested simulations (DF). The vegetation parameters were derived from a number of
(1981); albedos from Wilson and Henderson-Sellers (1985); surface resistance from
Shuttleworth et al. (1984) and Monteith (1976); canopy capacities from Dolman (personal

Figure 7. Imposed change in albedo over South America. The box encompassed by the solid black line shows
the area used for averaging purposes, Region 1.
communication); roughness length from Brutsaert (1982), Thompson et al. (1981) and Eagleson (1970); and infiltration factor from Warrilow et al. (1986). The derivation of the soil parameters is described in Warrilow et al. (1986).

In summary, comparing the deforested with the control simulation: root depth (i.e. the average depth of soil from which moisture is available to plant roots) is reduced; albedo is increased (see Fig. 7), as pasture reflects more solar radiation than forests which are very efficient absorbers and scatterers of short-wave radiation; surface resistance to evaporation is less for pasture under conditions of freely available soil moisture; roughness is reduced as forests offer a significant resistance to the wind in the lower layers of the atmosphere; and the infiltration factor is larger for forests as tree root systems and forest litter enhance the infiltration rate. All soil characteristics were kept constant, except for the infiltration capacity which is dependent on the infiltration factor. The possible effects of soil changes associated with deforestation are discussed in section 7(e).

6. Results

In the following section we present results from the 3-year control (NC) and deforested (DF) simulations. The integrations were started at the end of June, using initial data from an earlier simulation. All results are displayed either for South America or as the average over an area, Region 1 (see Fig. 7) in which the imposed change in surface characteristics was a maximum. All tabulated results are averaged over the three years, excluding the first month when the model is still adjusting to the boundary conditions.

(a) Seasonal changes in important climatological variables

Figure 8 shows the monthly mean values for a number of important climatological variables for the control and deforested simulations, and the change caused by deforestation.

Rainfall has decreased in almost every month throughout the three years, with the exception of four months when small increases are apparent that can be attributed to natural climate variability. The decrease in rainfall ranges between 0.1 and 3 mm d\(^{-1}\) with no seasonal bias.

There has been a marked reduction in evaporation in almost all months. The largest decreases occur when soil moisture limits evaporation—for typical soils in this region this occurs when the available soil-moisture concentration falls below the critical value of 0.07. Figure 8(e) shows that in a deforested climate the surface layer is drier in this sense for a larger proportion of the year.

Monthly values of surface temperature have mostly increased, by between 1 degC and 4 degC throughout the three years.

Total runoff shows increases and decreases which over the three years almost cancel, leaving a small negative residual. In contrast, surface runoff shows substantial increases during the three years, primarily as a result of the reduction in infiltration capacity. This is most apparent during the wet season in the first and the third years. In some months there is a small fall in surface runoff, which implies that either the magnitude of rainfall is less, or that rainfall is occurring in less intensive bursts.

(b) Seasonal and spatial changes in rainfall

Figure 9 displays maps of the change in rainfall for each season and the year and highlights areas that are statistically significant at the 90% confidence level or above. With samples of only three years (4 degrees of freedom), the magnitudes of differences
Figure 8.  (a) Simulated monthly mean rainfall averaged over Region 1. Dashed line represents the deforested simulation (DF), solid line the control simulation (NC). (b) Simulated monthly mean difference of deforested (DF) minus control (NC) simulated rainfall averaged over the Region 1. (c) As for (a) but for evaporation. (d) As for (b) but for evaporation. (e) As for (a) but for soil-moisture concentration. (f) As for (b) but for soil-moisture concentration. (g) As for (a) but for surface temperature. (h) As for (b) but for surface temperature. (i) As for (a) but for total runoff. (j) As for (b) but for total runoff. (k) As for (b) but for surface runoff.
Figure 8. Continued.
Figure 9. (a) Simulated change in precipitation pattern over South America due to deforestation for the southern hemisphere summer (December, January, February) averaged over the 3 years. Contours are at 2 mm d$^{-1}$, areas of decrease are stippled, and changes locally significant at the 90% confidence level or above are heavily shaded. (b) As for (a) but for southern hemisphere autumn (March, April, May). (c) As for (a) but for southern hemisphere winter (June, July, August). (d) As for (a) but for southern hemisphere spring (September, October, November). (e) As for (a) but for the year and contours at 1 mm d$^{-1}$. 
need to be relatively large to be detectable against the model's variability; the 90% confidence level on Student's 't' test requires a difference of 2.13 s.d. (s.d. = standard deviation) compared with 1.65 s.d. for an infinite population. The null hypothesis we shall be testing for rainfall is that rainfall decreases owing to deforestation. We can, therefore, use a 'one-tailed' test on negative differences over the deforested area, defining this for simplicity as Region 1; the probability of changes in rainfall exceeding the 90% confidence level by chance and being negative is 5%. The frequency of decreases giving a larger value of t is a valuable indicator of the overall statistical significance. It has, therefore, been monitored, and is used in the ensuing discussion. It should be borne in mind that by chance 5% of the area should be expected to have 'significant' differences of each sign.

Southern summer. There is an expansive area of reduced rainfall over the southwestern Amazon basin. The largest reductions over this area represent a 40–50% shortfall in rainfall. The region is dominated by reduced rainfall with significant decreases covering 18% of Region 1 (compared with the 5% expected). Some areas, such as the Andes near 5–10°S, experience significant increases in rainfall, covering 11% of Region 1, twice the chance expectation.

It is also interesting to note that there are changes in rainfall outside the immediate area in which the land surface changes were imposed. These differences are most notable in eastern Brazil, where there are considerable reductions and increases in rainfall, which are comparable with differences within the perturbed area.

Southern autumn. The area of maximum decrease in rainfall centred over north-eastern Peru broadly coincides with the area of maximum precipitation at this time of year. The largest reductions in rainfall represent a fall of 40% (or 4 mm d⁻¹) and lie within the deforested area. This area of reduced rainfall now extends into the Andes, in sharp contrast to the previous season. Significant decreases cover 15% of Region 1. There are also widespread changes in rainfall over most of eastern Brazil; these show
some similarity in pattern to the southern summer with increases over much of the region, but decreases over a large area centred on the Bahia province.

Southern winter. Significant rainfall reductions are now concentrated over the Andean regions from Peru to Colombia and then eastward along the edge of the deforested area, and are again broadly linked to the area of greatest precipitation. Maximum reductions reach up to 65% (or 6 mm d\(^{-1}\)). Changes in rainfall over the rest of South America are mostly small decreases. However, it is difficult to draw any meaningful conclusions about changes over the south-western Amazon basin as the model is unable to capture realistic totals here for the control simulation during this season. The fraction of Region 1 with significant decreases is only 8%, though there are a number of points just beyond the north-west of Region 1 that have significant decreases.

Southern spring. There is an extensive area of significantly reduced rainfall that affects a broad strip of Brazil near 60°W southward to northern Paraguay, 23% of Region 1 having significant decreases. The maximum reductions in rainfall represent a 30% (or 4 mm d\(^{-1}\)) shortfall. Some increases in rainfall totals are also apparent, most notably in Ecuador and Colombia, but these cover only 5% of Region 1.

Annual mean. When the whole year is considered almost all of northern South America experiences reductions in rainfall. Over the degraded area the largest reductions in rainfall are concentrated in north-western and central South America; the decreases are significant over 29% of Region 1. Outside this area there are comparable reductions over much of eastern Brazil. There are only two small areas within South America that show significant increases in rainfall; the larger of the two is over the southernmost part of eastern Brazil and the other is over the southern coastal part of Peru.

The areas of reduced rainfall over the central Amazon basin coincide with parts of the region over which the largest changes in surface characteristics were imposed (see Fig. 7). It is mainly the influence of the higher surface albedo that is expected to reduce rainfall, predominantly through changes in atmospheric circulation over South America (Charney 1975; Mylne and Rowntree 1992). A measure of the vertical circulations is the divergent flow and velocity potential, shown in Fig. 10, for the control simulation together with the annual mean difference between the deforested and control simulations. The divergence at this level, associated with ascent below, is of importance for its effect on the planetary distribution of vorticity at a level where eastward flow and associated wave propagation is a maximum (see, for example, Webster (1983)). Figures 10(a) and 10(b) indeed demonstrate that there is reduced divergence at upper levels, implying less convergence at lower levels and less ascent through the atmosphere. This in effect corroborates the Charney (1975) mechanism.

(c) Annual changes in the radiative balance at the surface

In Table 2 we consider the net radiation balance at the surface for Region 1. The downward component of solar radiation reaching the surface has increased owing to a reduction in the cloud amount. However, the effect of the imposed increase in surface albedo dominates, and so the net solar radiation absorbed at the surface is less. The decrease in cloud amount and atmospheric moisture in the lower layers of the atmosphere (not shown here) reduce the amount of long-wave radiation that is absorbed by the atmosphere and re-radiated back to the surface. The increase in net long-wave radiation lost from the surface is enhanced by higher surface temperatures. Overall the net radiation absorbed at the surface is less, the decrease being mostly due to the greater loss in long-wave radiation. As a result of the decrease in net radiation absorbed at the surface, the turbulent exchanges of heat and moisture between the surface and the atmosphere are diminished.
Figure 10. (a) Simulated velocity potential and divergent wind field at 200 mb over South America for the year, averaged over the 3 years. (b) Simulated change in velocity potential and divergent wind field at 200 mb over South America due to deforestation for the year, averaged over the 3 years. Contours every $0.5 \times 10^6 \text{m}^2\text{s}^{-1}$.

<table>
<thead>
<tr>
<th>TABLE 2. RADIATION BALANCE AT THE SURFACE FOR THE DEFORESTED (DF) AND CONTROL (NC) SIMULATIONS, AND THE DIFFERENCE BETWEEN THE TWO SIMULATIONS (DF–NC) FOR 3-YEAR MEANS AVERAGED OVER REGION 1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Radiation variable</td>
</tr>
<tr>
<td>Solar downward component</td>
</tr>
<tr>
<td>Solar upward component</td>
</tr>
<tr>
<td>Net solar</td>
</tr>
<tr>
<td>Long-wave downward component</td>
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<tr>
<td>Long-wave upward component</td>
</tr>
<tr>
<td>Net long wave</td>
</tr>
<tr>
<td>Net surface</td>
</tr>
</tbody>
</table>

(d) Summary of annual changes in important climatological variables

Table 3 summarizes the results of the annual changes and includes average values of a number of other important atmospheric and surface variables. Deforestation has caused a depletion in all components of the moisture balance at the surface. There has also been a reduction in the convergence of moisture into the basin (estimated as the difference between the precipitation and the evaporation). Although surface pressure is less over the basin, there has been little change relative to the adjacent ocean and, therefore, this has little impact on the moisture flux from the ocean. The reduction in cloud amount is consistent with less rainfall. The surface soil layer is now much drier mainly because of the imposed reduction in the water-holding capacity. The drier surface restricts evaporation and so warms the surface; this warming allows more long-wave
TABLE 3. SELECTED CLIMATOLOGICAL VARIABLES FOR THE DEFORESTED (DF) AND CONTROL (NC) SIMULATIONS, AND THE DIFFERENCES BETWEEN THE TWO SIMULATIONS (DF–NC) FOR 3-YEAR MEANS AVERAGED OVER REGION 1

<table>
<thead>
<tr>
<th>Climatological variable</th>
<th>NC</th>
<th>DF</th>
<th>DF–NC</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total evaporation (E) (mm d$^{-1}$)</td>
<td>2.28</td>
<td>1.73</td>
<td>-24%</td>
</tr>
<tr>
<td>Interception (mm d$^{-1}$)</td>
<td>0.77</td>
<td>0.69</td>
<td>-10%</td>
</tr>
<tr>
<td>Precipitation (P) (mm d$^{-1}$)</td>
<td>5.69</td>
<td>4.88</td>
<td>-14%</td>
</tr>
<tr>
<td>Moisture convergence (P–E) (mm d$^{-1}$)</td>
<td>3.41</td>
<td>3.15</td>
<td>-8%</td>
</tr>
<tr>
<td>Soil moisture (cm)</td>
<td>13.42</td>
<td>4.63</td>
<td>-65%</td>
</tr>
<tr>
<td>Total runoff (mm d$^{-1}$)</td>
<td>3.34</td>
<td>3.14</td>
<td>-6%</td>
</tr>
<tr>
<td>Surface runoff (mm d$^{-1}$)</td>
<td>1.72</td>
<td>2.19</td>
<td>+27%</td>
</tr>
<tr>
<td>Cloud amount (fraction)</td>
<td>0.59</td>
<td>0.55</td>
<td>-7%</td>
</tr>
<tr>
<td>Net radiation (W m$^{-2}$)</td>
<td>141.25</td>
<td>122.80</td>
<td>-13%</td>
</tr>
<tr>
<td>Surface temperature (°C)</td>
<td>25.0</td>
<td>27.13</td>
<td>+2.1°C</td>
</tr>
<tr>
<td>Sensible heat (W m$^{-2}$)</td>
<td>75.33</td>
<td>72.44</td>
<td>-4%</td>
</tr>
<tr>
<td>Horizontal wind for bottom model level (m s$^{-1}$)</td>
<td>2.36</td>
<td>3.39</td>
<td>+44%</td>
</tr>
</tbody>
</table>
| Surface pressure (mb)                       | 1008.35| 1007.96| -0.4 mb

cooling and so reduces the net radiation absorbed at the surface. The horizontal wind has increased in response to the imposed decrease in roughness.

7. DISCUSSION

In this paper we have used a more physically based solution to the representation of interception from the forest canopy. Most importantly, we have demonstrated that the model captures a realistic description of the present-day climate of Amazonia. Potentially, this allows us to make a more confident statement concerning the simulated characteristics of the climate following large-scale removal of the forest. However, to substantiate our findings, we must provide plausible physical explanations for the changes.

The key to understanding the complex interactions that occur between the forest and the atmosphere lies in appreciating that changes in any one of the hydrological components can set into motion a cycle of events. This is primarily due to the integral role that the forest plays in sustaining the moisture cycle. This has been verified from observational studies that have shown that at least half of the rainfall falling in the Amazon basin is evaporated or transpired back into the atmosphere and is then available for further rainfall events (Salati et al. 1979).

Let us now consider the likely reasons for the changes in some of the most important variables.

(a) Evaporation

The mechanisms controlling changes in evaporation are primarily driven by the imposed changes in albedo, roughness and the depth of water available to plant roots. Increased albedo allows less absorption of the incoming solar radiation and, hence, less available energy for latent-heat exchanges. The response of evaporation to reduced roughness is more complex. Some insight can be gained by examining the Penman–Monteith equation (Rowntree 1991). Under wet conditions when the surface resistance is zero, the evaporation loss resulting from interception is reduced as roughness decreases. When dry conditions prevail, evaporation will only decrease if the surface resistance remains below a critical value, whilst above this value small increases in evaporation are expected. Hence it is extremely important to model the process of canopy-wetting correctly. Results here indicate that evaporation is reduced in almost all months. The largest reductions in evaporation occur at the driest periods of the year, when evaporation is also constrained by decreased soil-moisture availability due to reduced root depths.
At this point it is relevant to consider the earlier similar deforestation experiment carried out by Lean and Warrilow (1989). Results from their simulation predicted a more severe impact on local climate, with rainfall and evaporation reduced by 20% and 27% respectively, and surface temperature increased by 2.5 degC. However, the rather crude treatment of the wetting of the canopy adopted in their experiment may have over-emphasised the role of the change in roughness. This would result in unrealistically high reductions in evaporation, and in turn would influence other components of the moisture and heat balance at the surface.

(b) Rainfall

Reduced evaporation implies that there will be a reduction in the net amount of water recycled into the atmosphere, and hence less rainfall. If this were the only mechanism operating then temporal and spatial changes in rainfall could be simply correlated with the corresponding changes in evaporation. However, this picture is complicated by changes in the atmospheric circulation and in the moisture transport from the adjacent oceans.

Changes in the pattern of circulation are most clearly defined during the southern hemisphere summer, when a heat low is firmly established over central Amazonia, and are readily explained by the Charney (1975) mechanism. The imposed increase in albedo together with the associated changes in long-wave radiation lead to the radiative heat balance of the basin becoming less positive. In order to maintain a thermal equilibrium, there will be a compensatory increase in subsidence. This will tend to suppress ascent within the atmospheric column and, by implication, the convergence of moisture into the centre of the low-pressure system near the surface. Reductions in the frictional drag as a result of reduced roughness will also tend to reduce moisture convergence (Sud et al. 1988).

In response to these highly complex mechanisms, changes in rainfall are variable both on a seasonal time-scale and spatially throughout South America. Within the area in which the forest was removed there are widespread changes throughout the year that are statistically significant at the 90% confidence level or above. This area tends to be dominated by intense patches of reduced rainfall. A longer simulation would be expected to give a more uniform distribution of average rainfall changes as well as a higher statistical significance for a given rainfall change (see section 6(b)). The average reduction in rainfall amounts to some 14% of the total, but rainfall totals at particular locations can fall by as much as two or three times the spatial average. The most severe deficits are felt during the southern winter with up to 65% reductions locally.

Although reduced rainfall is certainly the dominant signal, some areas do experience increases in rainfall. These tend to be located over the mountainous regions of Peru and Ecuador. However, it is impossible to make a quantitative estimate of the likely change in rainfall as the model underestimates orographic rain.

As a result of the change in circulation pattern within South America, the impact of deforestation appears to extend well beyond the immediate area in which the forest was removed. For example, during the southern summer and autumn there are large fluctuations in rainfall over eastern Brazil that are statistically significant and of comparable magnitude to changes within the perturbed area. Over the year, 20% (12%) of the area of Brazil east of 50°W has significant decreases (increases) compared with the 5% expected by chance.

It is the highly non-uniform nature of rainfall modulation that could have the most serious consequences for the livelihood of the region. At present agricultural crops in Brazil are partly maintained by moisture exported from the Amazon basin (Marques et
For these crops it is not the overall reduction in rainfall that is important but more the denial of rainfall at critical stages in their development. The regrowth of forest within the perturbed area and the survival of forest in outlying areas may also be threatened by a lengthening of the dry season together with less rainfall at other times of the year (Salati and Vose 1984).

(c) Temperature

The increase in the spatially averaged surface temperature is consistent with the reduction in the energy used in evaporating water at the canopy and soil surface, and the imposed reduction in roughness. The decrease in aerodynamic roughness will reduce the efficiency of the turbulent transfer of energy from the surface so that a higher surface temperature is then needed to remove this excess of energy.

(d) Runoff

Surface runoff is controlled by changes in the spatial distribution and intensity of rainfall, and the infiltration capacity. In this case the imposed reduction in infiltration capacity dominates over the reduction in rainfall, allowing higher surface runoff. However, the response in surface runoff is highly dependent upon the formulation employed. For example, in the earlier deforestation experiment by Lean and Warrilow (1989) surface runoff was seen to decrease, as higher reductions in rainfall counteracted the reduction in infiltration capacity. This highlights the difficult nature of the problem in hand, as it is not easy to verify the treatment of runoff or the values chosen for infiltration capacities against observations.

(e) Soil characteristics

Soil characteristics were kept constant owing to a lack of direct measurements from cleared forested areas; however, in order to compare with other similar experiments it is useful to speculate on the impact due to probable soil changes following deforestation. Recent deforestation scenarios have assumed that the soil texture will become finer (Nobre et al. 1991; Dickinson and Henderson-Sellers 1988). As the texture becomes finer two counterbalancing mechanisms operate (Warrilow et al. 1986); the field capacity increases and so allows more water to remain in the soil, whereas the saturated conductivity decreases and so promotes more surface runoff. Hence it is difficult to predict the impact on total runoff. Evaporation will be largely affected during the southern winter when the soil moisture falls below its critical value.

In the future, data collected from the joint Anglo-Brazilian Amazonian Climate Observation Study (ABRACOS) (Wright et al. 1992; Shuttleworth et al. 1991) will provide quantitative measurements of differences in land surface characteristics, including soil characteristics, for large adjacent areas of cleared and uncleared forest at three sites across the Amazon basin. These will allow adoption of agreed scenarios for the imposed changes to be run in different GCMs.

CONCLUSIONS

In this paper we have reported results of experiments which indicate that, in a GCM simulation, deforestation of Amazonia will change the climate over the deforested area and some adjacent regions. However, until we are able to compare such GCM results directly with actual measurements, for example for the areas presently deforested in Amazonia, we must be cautious in interpreting them as predictions of a future deforested climate. Also, because of the high levels of natural climate variability in tropical locations
it has not been possible to establish the implications for climate in other parts of the tropics using results for only three simulated years. However, the merit of this and other such experiments should not be underestimated, as they have provided substantial insight into the complex interactions that occur between the atmosphere and the surface following forest removal.

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