Modelling tropical deforestation: A study of GCM land–surface parametrizations

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SUMMARY

Tropical deforestation, by changing land surfaces, may have important consequences for the climate system. Predicting even the local, immediate effects of replacing tropical broadleaf forest with impoverished grassland has been difficult, because the land–surface parametrization schemes used previously in climate models have been inadequate. The forest canopy is particularly important for the surface-energy budget in tropical regions, and models neglecting the occurrence of such a canopy may give an unrealistic partitioning between various surface-energy fluxes. Inclusion of a land–surface scheme with a vegetation canopy into a version of the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM) with a diurnal as well as a seasonal cycle permits an exploratory study of the possible effects of tropical deforestation. In a 13-month integration that assumes that all of the Amazon tropical forest in South America is replaced by impoverished grassland, surface hydrological and temperature effects dominate the response. Reduced mixing and less interception and evaporation from the canopy cause runoff to increase and surface temperatures to rise by 3–5 K. The period of driest soil is increased in the model from one month to several, but the possibility that this change is random cannot be excluded. Increased temperatures and drier soil could have a detrimental impact on survival of the remaining forest and on attempts at cultivation in deforested areas. The land–surface model, driven in a stand-alone mode by prescribed atmospheric conditions and with an imposed seasonal cycle of rainfall, mimics the seasonal cycle of soil moisture and runoff found in the CCM. Hence, it is used to estimate the relative contribution of the various changes imposed to simulate deforestation in the CCM with respect to the model’s response at the surface. The change in surface roughness interacting with the canopy hydrology is evidently a major factor in determining the surface response to deforestation. However, the response to change in roughness is less pronounced for simpler models.

1. INTRODUCTION

Tropical forests are being perturbed in a number of different ways (e.g. Myers 1980a, b; Jordan 1982). Traditional shifting cultivation is being overtaken by wholesale removal of vast tracts, often prompted by highway development and economic incentives for large cattle ranches and small-scale farming colonization (Fearnside 1982, 1987). The question of the effects of removal of a large area of tropical forests on regional and global climate has been considered in a number of contexts (Newell 1971; Potter et al. 1975; Lettau et al. 1979). Salati and Vose (1984) and Salati (1987) show that over the Amazon Basin about half of the rainfall is returned to the atmosphere by evapotranspiration. The balances of energy and water at the surface are known from general circulation model (GCM) simulations to be important for the climate of tropical continental areas (Charney et al. 1977; Walker and Rowntree 1977; Shukla and Mintz 1982; Mintz 1984). Model patterns of precipitation and surface pressure have been shown to depend on prescriptions of evapotranspiration and albedo. Sud and Smith (1985) have additionally emphasized the effects of surface roughness.

Although these past studies have suggested the importance of various surface processes, they do not provide a quantitative understanding of the effects of tropical deforestation. To study these effects, we need both a suitable GCM and an adequate

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description of surface processes, including evapotranspiration, for the tropical forests and for whatever might replace them, for example, degraded grasslands. It is doubtful whether past GCM studies of tropical deforestation have satisfied the latter criterion.

Henderson-Sellers and Gornitz (1984) considered the impact of removing the Amazon forest in the Goddard Institute for Space Studies (GISS) GCM. A total area of $5 \times 10^6$ km$^2$ of tropical moist forest was assumed to be replaced with grassland, by changing surface-roughness length, moisture-holding capacity of the two model soil layers, and surface albedo. The response was that rainfall decreased by 0.5-0.7 mm d$^{-1}$ and evapotranspiration and total cloud cover also decreased. Surface albedo increased from 0.11 to 0.17 and surface temperature change was negligible. The planetary albedo was also essentially unchanged as decreased cloud cover compensated for the change in surface albedo, and there was no important response in the Hadley or Walker circulation regimes.

Wilson (1984) considered the sensitivity of the Meteorological Office (UKMO) GCM to conversion of tropical forest to grassland throughout the tropics. The response was initially dominated by surface albedo modification, but as the nine-month integration progressed, the partitioning of the latent and sensible heat flux terms shifted. The temperature response was variable with no systematic change in any of the deforested regions.

Past simulation experiments suggest some common themes: that changes in the surface hydrology are at least as important as changes in the surface albedo; that changes in surface roughness may also be significant; that the results are sensitive to the parametrizations inherent in the model and to the input land–surface information; and that while the local disturbance may be large, it might be difficult to establish climatic change in any area distant from the region of perturbation.

The present study is intended to help clarify the role of the vegetation canopy in determining the climatic response to tropical deforestation. Our results especially emphasize the importance of the interaction between surface roughness and hydrology. Because of the complexity and difficulty of realistically modelling surface processes, we restrict our analysis largely to the local surface response. We believe it necessary to understand this question first, before analysis of atmospheric effects becomes meaningful.

2. LAND–SURFACE SCHEMES USED TO REPRESENT TROPICAL FORESTS: A RANGE OF COMPLEXITY AND APPLICABILITY

In early GCMs, the land surface was simply a reflector of solar radiation and an emitter of infrared radiation. The first hydrological parametrization scheme, that of Manabe (1969), has been termed the ‘bucket’ model. In this highly simplified scheme, the ‘soil’ had a ‘field capacity’ of 15 cm. The bucket filled with water when precipitation exceeded evaporation, and after the bucket became full the overflow was runoff. Evapotranspiration occurred in these models at its potential rate when the soil was at or close to saturation. When soil moisture dropped below some critical value, the evapotranspiration was proportional to potential evapotranspiration, the proportionality factor being set equal to the ratio of the current soil moisture to the critical soil moisture.

Recently, Hunt (1985) reviewed single- and multiple-layer soil parametrization schemes for climate modelling, considering particularly their applicability to drought-prone semi-arid regions. He compared the schemes of Manabe (1969) with that of Hansen et al. (1983) for mid-latitudes and with that of Deardorff (1978), assuming bare soil and a diurnal mean sun. He found the Deardorff model to have a fast initial response to drying conditions followed by a slow release of deeper water, from which he inferred it to be superior to the other two models.
The empirical basis for the bucket parametrization is \textit{diurnally averaged} data. It is therefore most useful in GCMs that use diurnally averaged solar heating, but it may be inappropriate for climate models that include a diurnal cycle of solar radiation at the surface. Once the diurnal cycle is included, it may be necessary to include realistic descriptions of the sensible and latent heat fluxes. To this end, Dickinson (1984) developed soil-water and canopy energy-balance parametrizations. The treatment of soil water is inferred from a much more detailed soil model and includes the diffusion limitation of evaporation that often occurs around midday.

The treatment of the canopy energy and moisture balance includes: (i) interception of precipitation by vegetation and subsequent evaporative loss and leaf drip; (ii) moisture uptake by plant roots, distributed between the upper and full soil columns; and (iii) stomatal resistance to transpiration. This land–surface scheme, referred to as the ‘biosphere–atmosphere transfer scheme’ (BATS), in common with the ‘simple biospheric model’ (SiB) of Sellers et al. (1986), can represent a very wide range of vegetation–soil coupled systems by selection of the appropriate land-cover and soil-description class. The BATS includes a complete range of vegetation types (Dickinson et al. 1986; Wilson et al. 1987), in addition to the soil parametrizations.

BATS and SiB generalize the aerodynamic transfer formulations usually used in GCMs to allow for the presence of multiple surfaces, i.e. canopy and soil, that separately are in energy balance. Aerodynamic drag coefficients are calculated for the canopy according to mixed-layer theory, as a function of the reference height at which atmospheric variables are available, roughness length, and stability. For describing the transfer of heat and moisture to the multiple surfaces, additional resistances are calculated on the basis of the structure and biophysiology of the canopies.

To identify possible limitations of a bucket soil hydrology model in the context of tropical deforestation, we compare simulations using BATS in a stand-alone mode with simulations using the scheme of Hansen et al. (1983), which was used by Henderson-Sellers and Gornitz (1984) in their recent study of tropical deforestation. We use for the latter the version applied to a tropical forest, that is, having two layers with water-holding capacities of 200 mm and 450 mm for the upper and lower layers, respectively, and coupled with a one-day time scale if the upper layer is wetter than the lower, but otherwise coupled instantaneously, as a single bucket. Our comparison of BATS with that of Hansen et al. is somewhat analogous to the study by Hunt, since BATS represents further development of the Deardorff (1978) soil and canopy model. A similar comparison with the Manabe (1969) model gave results comparable to those given by Hunt and to those reported here for the Hansen et al. model. The bare soil model of Deardorff, as applied by Hunt, is inappropriate for a tropical forest.

For this stand-alone comparison of the surface models, we prescribe surface winds as $3 \text{ m s}^{-1}$, temperature as $300 \text{ K}$ plus a sinusoidal diurnal cycle with the range $6 \text{ K}$, and water-mixing ratios as constant at the saturated value for 00 h local time; all prescribed parameters are applied at 10 m above the zero displacement height of the canopy. Incident diurnal solar radiation during dry conditions has a noontime peak of $890 \text{ W m}^{-2}$ and daytime sinusoidal variation is halved during rainfall. Downward long-wave radiation is prescribed as a function of air temperature and water vapour according to the Brunt formula.

(a) \textit{Drying-out scenario}

In Fig. 1, we compare the drying out over a 150-day period of the stand-alone BATS versus the two-layer bucket model of Hansen et al. (1983). The latter is referred to as the Goddard Institute for Space Studies (GISS) model, although that institution is
currently developing a canopy model similar to the one used here. All cases are started with saturated soil. Values shown are three-day means. Both models were run assuming a tropical forest roughness length of 2 m and a short vegetation roughness length of 0.05 m. For neutral conditions, the aerodynamic drag coefficient is obtained from $k^2/\ln^2(z/z_o)$ where $k$, the von Kármán constant, is put equal to 0.4, $z$ is the 10 m reference height, and $z_o$ the roughness length; hence, the forest and short vegetation have drag coefficients at 10 m of about 0.06 and 0.006, respectively. For the sake of comparison, the same transfer coefficients and absorbed solar radiation calculated for BATS were also applied to the GISS model. Other parameters used with the BATS model are the soil texture 10 and colour class 3, as defined in Table 1, and the vegetation cover class 6 (evergreen broadleaf tree), as defined in Table 2.

As seen in Fig. 1(a), the GISS model with tropical forest roughness rapidly dries out for the first 50 days, with latent heat fluxes well in excess of the net radiation and corresponding (not shown) negative sensible heat fluxes of up to several hundred W m$^{-2}$. The BATS model, both rough and smooth, also dries out somewhat rapidly for about the first 25 days, but average sensible fluxes remain positive, around 20–30 W m$^{-2}$ for the moist conditions, increasing to 100–150 W m$^{-2}$ for the dry conditions when the latent fluxes have dropped to a residual of about 40–50 W m$^{-2}$ for the remainder of the period,
### TABLE 1. Soil parameters used for South America

<table>
<thead>
<tr>
<th>Texture class (from sand (1) to clay (12))</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
</tr>
</thead>
<tbody>
<tr>
<td>a) Porosity (volume of voids to volume of soil)</td>
<td>0.42</td>
<td>0.45</td>
<td>0.48</td>
<td>0.51</td>
<td>0.54</td>
<td>0.57</td>
<td>0.60</td>
<td>0.63</td>
<td>0.66</td>
</tr>
<tr>
<td>b) Maximum soil suction (m)</td>
<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
</tr>
<tr>
<td>c) Saturated hydraulic conductivity (m s(^{-1}))</td>
<td>(0.13\times10^{-4})</td>
<td>(0.89\times10^{-5})</td>
<td>(0.63\times10^{-5})</td>
<td>(0.45\times10^{-5})</td>
<td>(0.32\times10^{-5})</td>
<td>(0.22\times10^{-5})</td>
<td>(0.16\times10^{-5})</td>
<td>(0.11\times10^{-5})</td>
<td>(0.8\times10^{-6})</td>
</tr>
<tr>
<td>d) Ratio of saturated thermal conductivity to that of loam</td>
<td>1.2</td>
<td>1.1</td>
<td>1.0</td>
<td>0.95</td>
<td>0.90</td>
<td>0.85</td>
<td>0.80</td>
<td>0.75</td>
<td>0.70</td>
</tr>
<tr>
<td>e) Exponent 'B', Clapp and Hornberger (1978)</td>
<td>5.0</td>
<td>5.5</td>
<td>6.0</td>
<td>6.8</td>
<td>7.6</td>
<td>8.4</td>
<td>9.2</td>
<td>10.0</td>
<td>10.8</td>
</tr>
<tr>
<td>f) Moisture content (relative to saturation) at which transpiration ceases</td>
<td>0.266</td>
<td>0.300</td>
<td>0.332</td>
<td>0.378</td>
<td>0.419</td>
<td>0.455</td>
<td>0.487</td>
<td>0.516</td>
<td>0.542</td>
</tr>
</tbody>
</table>

### II Functions of colour parameter

<table>
<thead>
<tr>
<th>Colour (from lightest (1) to darkest (8))</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
</tr>
</thead>
<tbody>
<tr>
<td>a) Dry soil albedo (&lt;0.7\ \mu m)</td>
<td>0.23</td>
<td>0.22</td>
<td>0.20</td>
<td>0.18</td>
<td>0.16</td>
<td>0.14</td>
<td>0.12</td>
<td>0.10</td>
</tr>
<tr>
<td>(\geqslant0.7\ \mu m)</td>
<td>0.46</td>
<td>0.44</td>
<td>0.40</td>
<td>0.36</td>
<td>0.32</td>
<td>0.28</td>
<td>0.24</td>
<td>0.20</td>
</tr>
<tr>
<td>b) Saturated soil albedo (&lt;0.7\ \mu m)</td>
<td>0.12</td>
<td>0.11</td>
<td>0.10</td>
<td>0.09</td>
<td>0.08</td>
<td>0.07</td>
<td>0.06</td>
<td>0.05</td>
</tr>
<tr>
<td>(\geqslant0.7\ \mu m)</td>
<td>0.24</td>
<td>0.22</td>
<td>0.20</td>
<td>0.18</td>
<td>0.16</td>
<td>0.14</td>
<td>0.12</td>
<td>0.10</td>
</tr>
</tbody>
</table>
constrained by the rate of diffusion of moisture through the soil. The smooth GISS case continues to lose its water over the 150-day period, with its evaporative fluxes dropping to values lower than those of BATS after 150 days. In comparing the results from the low roughness GISS model in Fig. 1 with the examples considered by Hunt (1985), who assumed a drag coefficient even lower than that of our low roughness case, we see that the time scale of drying out is largely controlled by the assumed water-holding capacity of the GISS "bucket".

An additional point revealed by Fig. 1(a) is that the surface roughness (or drag coefficient) is also important in determining the time scale of drying out for bucket models. Presumably, some aspects of the dependence on drag coefficient are somewhat exaggerated by the assumption, for all cases, of the same atmospheric conditions above the canopy, since various atmospheric feedbacks would weaken the dependence of the rate of drying out on the drag coefficient (e.g. Jarvis and McNaughton 1986). In particular, a larger drag coefficient would weaken surface winds and hence the transfer rates. The rate at which the planetary boundary layer can eliminate its accumulated heat and moisture may become limiting. Also, fixing surface air temperature precludes possible decreases in this temperature for increased drag coefficient that could decrease soil and canopy temperature, and hence reduce the evapotranspiration. Nonetheless, for the assumed conditions, the BATS model appears to give more realistic results for sensible and latent heat fluxes than does the bucket model, as Hunt also concluded for the Deardorff bare soil model.

The rise in soil temperature that accompanies the soil moisture losses is indicated in Fig. 1(b). With wet conditions, the soil in the BATS low roughness case is warmer by 2 K than that of the high roughness case and this difference increases to 5 K or more as the drying progresses. The soil temperature in the GISS model is less affected by the differences in roughness, but after 75 days of drying, the temperature of the low roughness case begins to increase relative to that of the forest-roughness case. As a result of the larger upward long-wave emission occurring with increased temperature, the net radiation of the low roughness BATS case is less than that of the high roughness case, and this difference becomes greater than 30 W m\(^{-2}\) after the initial ten days. The thermal emission, like temperature, remains close for the two GISS cases over the first 100 days of the integration.

Even for the same large forest roughness, the soil temperature for the BATS model exceeds that of the GISS model by 1–2 K. Some of this difference during the first 60 days of drying out is ascribed to the greater evaporative cooling of the GISS model. A difference of 1.3 K persists, however, even after the GISS model has lost its water. This difference depends on the diurnal cycle of heating and on the presence of the canopy model in BATS, and it disappears when diurnal mean conditions are applied. During the day, the slowly transpiring canopy becomes warmer than the soil of either model because of lack of thermal inertia, but it also drops to lower temperatures at night, and so has a mean temperature about 0.4 K less than that of the soil. However, during the conditions of daytime mixing, the warm canopy transfers its heat to the soil more readily than it removes heat from the soil during stable nighttime conditions.

The canopy in BATS, as the primary surface of energy exchange, should have about the same diurnal averaged temperature as the soil in the GISS model when both models have dry soil and the same drag coefficient and solar heating are applied. The small difference in the tropical forest cases of ~0.9 K warmer for the BATS canopy than the GISS soil temperature, results primarily from the additional leaf boundary-layer resistance applied in the BATS model and a smaller drag coefficient applied over the 0.1 of the BATS model area that is assumed unvegetated.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Land cover/vegetation type</th>
<th>1</th>
<th>2</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>11</th>
<th>17</th>
<th>18</th>
<th>19</th>
</tr>
</thead>
<tbody>
<tr>
<td>a) Maximum fractional vegetation cover</td>
<td></td>
<td>0.85</td>
<td>0.80</td>
<td>0.80</td>
<td>0.90</td>
<td>0.80</td>
<td>0.10</td>
<td>0.80</td>
<td>0.80</td>
<td>0.80</td>
</tr>
<tr>
<td>b) Difference between maximum fractional vegetation cover and cover at 269K</td>
<td></td>
<td>0.6</td>
<td>0.1</td>
<td>0.3</td>
<td>0.5</td>
<td>0.3</td>
<td>0.3</td>
<td>0.3</td>
<td>0.2</td>
<td>0.3</td>
</tr>
<tr>
<td>c) Roughness length (m) of vegetation</td>
<td></td>
<td>0.06</td>
<td>0.02</td>
<td>0.8</td>
<td>2.0</td>
<td>0.1</td>
<td>0.1</td>
<td>0.1</td>
<td>0.8</td>
<td>0.05</td>
</tr>
<tr>
<td>d) Depth of the total soil layer (m)</td>
<td></td>
<td>1.0</td>
<td>1.0</td>
<td>2.0</td>
<td>1.5</td>
<td>1.0</td>
<td>1.0</td>
<td>1.0</td>
<td>2.0</td>
<td>1.0</td>
</tr>
<tr>
<td>e) Depth of upper soil layer (m)</td>
<td></td>
<td>0.1</td>
<td>0.1</td>
<td>0.1</td>
<td>0.1</td>
<td>0.1</td>
<td>0.1</td>
<td>0.1</td>
<td>0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>f) Rooting ratio (upper to total soil layers)</td>
<td></td>
<td>3</td>
<td>8</td>
<td>10</td>
<td>12</td>
<td>8</td>
<td>8</td>
<td>5</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>g) Vegetation albedo for wavelengths &lt;0.7 (\mu)m</td>
<td></td>
<td>0.10</td>
<td>0.10</td>
<td>0.08</td>
<td>0.04</td>
<td>0.08</td>
<td>0.17</td>
<td>0.08</td>
<td>0.06</td>
<td>0.08</td>
</tr>
<tr>
<td>h) Vegetation albedo for wavelengths &gt;0.7 (\mu)m</td>
<td></td>
<td>0.30</td>
<td>0.30</td>
<td>0.28</td>
<td>0.20</td>
<td>0.30</td>
<td>0.34</td>
<td>0.28</td>
<td>0.24</td>
<td>0.30</td>
</tr>
<tr>
<td>i) Minimum stomatal resistance (s m(^{-2}))</td>
<td></td>
<td>150</td>
<td>250</td>
<td>250</td>
<td>250</td>
<td>250</td>
<td>250</td>
<td>250</td>
<td>250</td>
<td>250</td>
</tr>
<tr>
<td>j) Maximum leaf area index (LAI)</td>
<td></td>
<td>6</td>
<td>2</td>
<td>6</td>
<td>6</td>
<td>6</td>
<td>6</td>
<td>6</td>
<td>6</td>
<td>6</td>
</tr>
<tr>
<td>k) Minimum LAI</td>
<td></td>
<td>0.5</td>
<td>0.5</td>
<td>1.0</td>
<td>5.0</td>
<td>0.5</td>
<td>0.5</td>
<td>1.0</td>
<td>3.0</td>
<td>0.5</td>
</tr>
<tr>
<td>l) Stem (and dead matter) area index</td>
<td></td>
<td>0.5</td>
<td>4.0</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
</tr>
<tr>
<td>m) Inverse square root of leaf dimension (m (^{-1/2}))</td>
<td></td>
<td>10</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
</tr>
<tr>
<td>n) Light sensitivity factor (m W(^{-1}))</td>
<td></td>
<td>0.01</td>
<td>0.01</td>
<td>0.03</td>
<td>0.03</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>0.03</td>
<td>0.01</td>
</tr>
</tbody>
</table>

1 See definitions in Table 2(b).
2 Vegetation type 19 has been defined especially for this study and is not part of the NCAR CCM land-type data set.
TABLE 2(b). VEGETATION/LAND COVER TYPES USED IN THE CCM

| 2. Short grass               | 12. Ice-cap/glacier    |
| 3. Evergreen needleleaf tree | 13. Bog or marsh       |
| 5. Deciduous broadleaf tree  | 15. Ocean             |
| 7. Tall grass                | 17. Deciduous shrub    |
| 9. Tundra                   |                        |
| 10. Irrigated crop          | *19. Impoverished scrub-grassland |

* Land type 19 has been introduced especially for this study and is not part of the CCM land data set.

(b) Steady-state scenario

The BATS and the Hansen et al. models were run to steady state, assuming that showers occurred every third day, beginning at 00, 06, 12, and 18 hours, local time, and lasting for 0·5 h with instantaneous rainfall rates of 0·003 mm s\(^{-1}\), corresponding to 5·4 mm per shower and \(=220\) mm month\(^{-1}\), that is, relatively dry conditions for a tropical forest. The diurnal cycle of energy fluxes for these two cases during the second dry day following the rainy day is shown in Fig. 2. The most notable distinction is that the BATS model gives significantly larger sensible fluxes and smaller latent fluxes than does the Hansen et al. model. The peak afternoon dry day differences in both fluxes are about 150 W m\(^{-2}\), and the mean difference in latent heat flux averaged over wet and dry days is about 22 W m\(^{-2}\) or about 0·4 mm d\(^{-1}\) of evapotranspiration. The fluxes from BATS appear more realistic than those from the Hansen et al. model, in comparison with those observed by Shuttleworth et al. (1984). The relatively low dry day evapotranspiration of the BATS model that results from its canopy resistance to water flux is compensated, in part, by increased interception loss during the rainy day. The mean difference in evapo- transpiration is compensated by less runoff in the GISS model.

During the rainy day (not shown), both models give similar flux patterns except that during rain the latent fluxes of the BATS models, as a result of interception loss, exceed somewhat the net radiative heating, whereas the Hansen et al. model actually reduces its latent fluxes in response to the assumed reduction in solar heating. This large evaporative flux in the BATS model is compensated by negative fluxes of sensible heat. Shuttleworth et al. (1985) show that total evaporation from a fully-wetted canopy in Amazonia often exceeds 'potential evaporation', the additional energy being obtained from air drawn into the wetted area from adjacent dry canopy forest.

We believe that BATS (which includes separate parametrizations of the foliage energy budget, the interception and subsequent evaporation of precipitation, and two soil layers with moisture and energy transfer) offers a more responsive and diurnally correct model of energy and moisture fluxes in a tropical forest. Although more complex than conventional surface schemes, when efficiently coded it should require but a few percent of the total computation cost in a GCM. The BATS land–surface parametrization is used in the following study of Amazonian deforestation.

3. RESULTS FROM A GCM TROPICAL AMAZONIAN DEFORESTATION EXPERIMENT

The land–surface parametrization scheme of Dickinson (1984) has been incorporated into a version of the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM) (Williamson 1983) with rhomboidal-15 spectral truncation and a
4.5° latitude × 7.5° longitude spatial mesh. This version also includes a nonlinear parametrization of atmospheric vertical diffusion (Williamson et al. 1987) and diurnal as well as annual cycles in solar radiation. To this version of the CCM has been added a geographical distribution of vegetation type, soil colour, and texture (Dickinson et al. 1986). The sensitivity of the coupled vegetation–soil, land–surface parametrization scheme to variations in vegetation and soil properties has been studied by Henderson-Sellers et al. (1986) and Wilson et al. (1987). Tables 1 and 2 list the soil properties and vegetation types used in the model for South America, and Figs. 3(a), (b) and (c) show their spatial distributions. In the deforestation experiment described here, all of the grid cells classified as vegetation type 6 (evergreen broadleaf tree) in South America have been modified to new vegetation type 19 (Fig. 3(d), cf. Fig. 3(a)) that is supposed to possess the characteristics of impoverished scrub–grassland typical of deforested regions in Amazonia (e.g. Fearnside and Rankin 1982; Dickinson 1987).

At each of the modified grid locations, the soil characteristics were also changed. The soil texture was made finer by two texture classes up to the maximum of 12, and the soil colour lighter by two colour classes. The gross effects of this assumed model of deforestation are thus to increase surface albedo—with the increased soil albedo, increased vegetation albedo, and reduced fractional vegetation cover—and to reduce the total available soil water and the vegetation-roughness length. In addition, the model's
surface-runoff formulation was modified for runoff to decrease less rapidly with drying of the surface layer.

Before considering the impact of the modelled deforestation, we review the climate simulation of the control (i.e. unperturbed) climate model.

(a) Unperturbed model climate of South America (Brazil)

Observationally, the seasonal movement of the equatorial trough (intertropical convergence zone) dominates the climate of northern Brazil, although changing pressure fields at higher (southern) latitudes are also important. A heat low becomes well developed in the centre of the continent in summer (January), which slightly weakens the easterly winds in the Amazonian region from the winter (July) pattern. Advedcted air loses moisture by precipitation as prevailing easterlies approach the Andes. In the summer (January), a southward air flow from the Amazon region also occurs which has been likened to a monsoon circulation. The surface pressure patterns generated by the model are in qualitative agreement with observational data (Fig. 4). The model vertical
velocity above the Amazon Basin responds to the seasonality, the ascending limb of a Walker circulation being well simulated in January, at which time there is ascent in all but the lowest model layer. In July, the vertical velocity is downward, strong near the surface but fairly weak aloft (cf. Fig. 8 of Henderson-Sellers and Gornitz (1984)).

The modelled precipitation patterns resemble the observed patterns that are shown in Fig. 5. Although the model annual totals over the continent are high (~2280 mm yr $^{-1}$ over the whole of South America), the seasonality is comparable to that of observational data. We have divided the northern part of South America into four regions (Fig. 6). Of these, region 2 includes most of the southern, central, and upper part of the Amazon Basin, whereas region 1 contains the northern basin forest, and region 3 includes the semi-arid north-east of Brazil. Region 4, southern Brazil, is not discussed further here.

Region 1 has an annual total precipitation of 3230 mm, and (somewhat unrealistically) region 2 receives a larger total rainfall of 3345 mm. The success of the seasonal simulation of precipitation can also be seen in Fig. 7(a) which compares the seasonal precipitation of region 2 with that of region 1. The average rainfall in these regions is fairly representative of the annual cycles in precipitation characteristic of the southern and northern portions of the Amazon Basin (Fig. 7(b)). The model also gives qualitatively the correct seasonal pattern for region 3, but otherwise does not capture particularly well the rainfall climatology of north-east Brazil. The total precipitation in that region is too high, with the mean annual precipitation ~1825 mm and rainfall of 225 and 45 mm in January and
Figure 5. Distribution of precipitation in the Amazon region (mm): January and July totals from the control simulation (model) and observations from Salati et al. (1978).

Figure 6. Four regions in South America used for climatological comparison of model results. The hatching indicates the region 2 which is emphasized in our analysis.
July, respectively. This annual value is considerably higher than the observed mean annual rainfall of <800 mm per year over this area (Trewartha 1961; Ratisbona 1976). The seasonal range in air temperature for region 2 (Fig. 8) is at least double the observed range of 2–3 K (Ratisbona 1976).

(b) Deforested climate

The major changes imposed by the modelled deforestation are decreases in the aerodynamic roughness length for vegetation, in the total soil depth, in the vegetation cover, and in the sensitivity of the vegetation stomata to visible radiation, and increases in the surface runoff, in the albedos of vegetation and dry soil, and in the relative density of vegetation roots in the upper soil layer. Decreasing the vegetation roughness and cover reduces the turbulent exchange, and hence potentially reduces energy transfer between the surface and the atmosphere for fixed surface-to-atmosphere gradients, and reduces the fraction of rainfall intercepted by vegetation.

The CCM deforestation experiment was initialized using 30 December data from the second year of a control CCM integration. The deforestation experiment has so far been run for about a year (end of December for 13 months to beginning of February). The results presented here are for the seasonal cycle. Since adjustment at the surface to the changed conditions is rapid, we include the complete integration period from the first January in our comparisons. However, for the first month or two of the deforestation, the atmosphere would still be approaching its new equilibrium.
Figure 8. (a) January and July distribution of $T_a$, surface-air temperature (K), and $T_s$, soil-surface temperature (K). (b) Monthly variation in $T_a$ and $T_s$ for region 2 (central, south Amazon).

Figure 9(a) shows the seasonal variations of rainfall for the control and deforested cases and their difference. The differences in rainfall appear noisy, with any systematic mean change likely to be less than 20% of the mean rainfall and not ascertainable without a longer integration. Likewise, whether any shifts in the seasonal pattern of rainfall have occurred cannot be answered here. The regional patterns of change for the first January and July (Fig. 9(b)) show some similarity, but again establishing their significance is not possible here. The daily mean temperatures of surface air and soil increase by 1 to 4 K (Fig. 10). The soil temperature increase exceeds that of the air by at least 1 K. A similar but weakened warming pattern is found in the lowest several model layers, e.g. at an altitude of ~500 m, the air temperature change is about half of that at the surface. The surface warming occurs largely in the daytime, as evidenced by a much larger increase in daily maximum than in daily minimum temperatures.
The surface-temperature change in the CCM is positive throughout, but it has two major peaks, one occurring in June/July and another larger one in October (Fig. 10(a)). The June/July temperature increase occurs during the dry season and is related to a considerable decrease in evapotranspiration in June (Fig. 11). The evaporation difference curve (Fig. 11) also dips in October, although less than in June. The only month showing an evaporation increase is September, presumably a result of a large increase in precipitation in that month (Fig. 9(a)). Total soil moisture is less (Fig. 12(a)) for the deforestation case because of the assumed decrease from 1-5 m to 1 m in the total soil column. However, relative saturation has increased somewhat and the upper soil moisture has increased slightly in all months except June and October. There is an overall slight decrease from one January to the next in the upper-layer soil moisture of the deforested case as well as a decrease in the total soil moisture (Fig. 12(b)), possibly related to transient adjustments over the first January of the soil water.
Figure 10. (a) As Fig. 9(a) except for surface-air temperature, $T_a$, soil-surface temperature $T_s$, and lower soil temperature $T_{\text{soil}}$ (K). (b) January and July temperature-difference maps (deforested – control) for $T_s$ and $T_a$ (K).
Figure 11. (a) As Fig. 9(a) except for total evaporation from the surface (mm month$^{-1}$). (b) Evaporation-difference (deforested - control) maps for June (mm month$^{-1}$).

Figure 12. (a) As Fig. 9(a) except for total and upper-layer soil moistures (cm). (b) Maps of differences (deforested - control) in total and upper-layer soil moistures for the second January (cm).
Figure 13. (a) As Fig. 9(a) except for interception (mm d\(^{-1}\)), total runoff (mm d\(^{-1}\)) and surface runoff. (b) Maps of differences in interception and total runoff for the second January (mm).

Interception decreased in all months except August (Fig. 13(a)) when very high rainfall in the deforested case caused a net increase. Both surface and total runoff show increases overall, in part due to the imposed surface-runoff increase, but mostly related to the increased wetness of the soil that results from the weakened evapotranspiration.

The second year of integration shows the major changes continuing. The maps of interception and total runoff differences for the second January (Fig. 13(b)) show decreased interception and increased runoff in the south and east. Figure 14 shows that differences between control and deforestation for the second January are very similar to those for the first January (Figs. 9 and 10), with increased surface-air and soil-surface temperatures and decreased total evaporation.
4. **Interpretation of CCM Results in Terms of BATS Model**

The differences between control and deforested cases are illustrated using the integrations of the stand-alone BATS model, as defined in section 2. The tropical forest parametrization was identical to that described in section 2, and the deforestation changes were the same as those described in section 3 as imposed on the CCM. Rainfall was prescribed as in section 2 but also at 0.5, 1.5 and 2.0 times the standard rate, and the model was again run to steady state. The deforested case was run with the same air temperature as the control and also with a 2K increase to evaluate the feedback of the surface-air warming that is noted in section 3. For comparison, we also ran a case in which only the roughness change of tropical deforestation is applied and one having this change plus the change of vegetation albedo.

The mean response of BATS to the prescribed conditions is shown in Fig. 15. As shown in Fig. 15(a), the temperature of canopy foliage warms with deforestation by 4–5 K during the day for fixed air temperature, and by an additional 2 K if the air is warmed by that amount. Changing only the roughness length still gives about a 3–4 K warming, but the increased albedo of the deforested vegetation reduces this warming by about 0.4–1 K. The mean surface-soil temperature shown in (b) increases by a comparable amount, although a somewhat smaller fraction of the warming is accounted for by the roughness change, and the effect of canopy albedo change is less noticeable. Response of the temperatures to deforestation appears to be almost independent of rainfall rates, although the canopy daytime temperatures become remarkably higher during dry conditions.
Evaporative and sensible fluxes are shown in (c) and (d), respectively. The tropical forest shows a much wider range of these fluxes for tall vegetation than for short vegetation over the rainfall scenarios considered. Fluxes of sensible heat and evapotranspiration are less sensitive to rainfall in the deforested case than in the forest case and approach constant levels for a rainfall rate greater than 200 mm month\(^{-1}\) as the soil water storage reaches near-constant values. Consequently, almost all further rainfall increases go into runoff. At large rainfall rates, evapotranspiration is reduced by deforestation by up to a factor of two with the change largely controlled by the change in roughness with some contribution from the albedo change. At low rainfall rates, the change in evapotranspiration is dominated by changes other than that of roughness and albedo. Increasing albedo reduces both the sensible and latent fluxes, whereas deforestation changes other than roughness reduce latent but increase sensible fluxes over the whole range of rainfall rates. The increase of air temperature to 302 K increases latent fluxes and decreases
sensible fluxes, but only by a small amount. As seen in (e), at large rainfall rates, the roughness change has the largest effect on changing runoff, but the other factors, including the increase of canopy albedo, also further increase the runoff, except that the air warming reduces it slightly. The change of soil water, as shown in (f), is dominated by the change in soil texture and by available storage assumed for deforestation. However, the change in roughness alone is seen to significantly increase the soil water, consistent with the reduced evapotranspiration.

The response of BATS to rainfall (Fig. 15) can be compared with similar variations seen in the CCM simulations. The assumed range of about 100 to 400 mm was selected to roughly match the 60 to 500 mm monthly rainfall shown in Fig. 9. The correspondence of the variations with rainfall of Fig. 15 with the evapotranspiration variations in Fig. 11, those of soil water in Fig. 12, and of runoff in Fig. 13 is remarkable. However, some differences are seen. Time-lag effects are noticeable in the CCM soil-moisture response to low rainfall amounts. Evapotranspiration is modified less with deforestation in the CCM simulation than in the stand-alone simulation, especially during the rainy season, as expected because of the lack of atmospheric feedbacks in the latter. Differences in the statistics of rainfall intensity and frequency between control and deforested cases in the CCM simulation would also affect changes in evapotranspiration and runoff.

In summary, of the various changes made in the model to simulate the effect of deforestation, the stand-alone simulation suggests that for steady state the reduction in the vegetation-roughness length has the largest overall effect, especially with large rainfall rates, but that the other deforestation changes are also significant. The reduction of the water-holding capacity of the soil appears to be important during transient dry periods.

5. DISCUSSION

Tropical deforestation leads to a wide range of environmental changes (Dickinson 1987). The biosphere–atmosphere transfer scheme first described by Dickinson (1984) has been incorporated into a version of the NCAR CCM that already includes the diurnal cycle of radiation and stability-dependent diffusion in the boundary layer. A simulation of the change from forest to deforested conditions has been attempted by including changes in the vegetation, a prescribed change in the soil texture and colour, and an increase in the rate of surface runoff as a function of the relative saturation of the upper soil layer.

The results from a 13-month CCM simulation are generally explicable in terms of the land–surface model used. Interception decreases and so does total evapotranspiration. Less removal of energy by latent heat results in increased temperatures and increased runoff. Surface-air temperature increases by ~1–3 K and soil-surface temperature by 2–5 K. The warming with reduced surface roughness evidently depends largely on the canopy energy balance, including interception, interacting with the diurnal cycle of solar radiation. For prescribed atmospheric conditions (uncoupled to the CCM) and the steady-state scenario considered in section 2, the GISS model soil temperatures increase by less than 1 K whereas, with BATS, soil temperatures increase by more than 3 K. Both the GCM studies of Henderson-Sellers and Gornitz (1984), using the GISS model, and that of Wilson (1984), using another simple model, found no systematic overall change in temperature for tropical deforestation. It appears that a model of the land surface neglecting the canopy hydrology cannot successfully represent the diurnal cycle of surface fluxes over a tropical forest.

With the short (13-month) GCM experiment described here, we can make no direct inferences about the likely non-local climatic effects of tropical deforestation. Indeed,
even with a much longer integration, in which the statistical significance of model changes could be established, we could not make such inferences with any confidence unless the validity of the treatment of land–surface processes for the forested and deforested regions had been adequately established. This study has focused on the physical mechanisms of land–surface schemes and has shown that the BATS model, with its treatment of the forest canopy, gives a rather different, and presumably more correct, response to changes in surface roughness than do models without a canopy parametrization.

The inferred warming of the surface is model-dependent, so our confidence in its occurrence and our ability to predict it will increase as we increase our confidence in or improve the components of the BATS land–surface model. However, the warming does agree with observational studies of the local impact of deforestation. For example, Ghuman and Lal (1987) have carried out a study in a tropical forest in Nigeria where they have deforested some study plots and examined the consequences for local microclimate. They found that the soil temperature was 3 K higher and air temperature 1.7 K higher on the average, and that this temperature increase at the surface was almost entirely a result of greater daytime maximum temperatures.

Only the local climate has been discussed here, but we expect that the changes from deforestation would also be important for the regional-scale atmospheric circulation, given the major changes in surface fluxes of sensible and latent heat that are found in the stand-alone and CCM simulations. There is some evidence (Fig. 12) that with deforestation the length of the season with dry soil would increase. A lengthened season of dry soil is consistent with the greater runoff and lower water-holding capacity of the deforested system. However, the wider minimum in soil water in Fig. 12 is also related to the decreased rainfall in May-June and increased rainfall in July-August (Fig. 9), which could be a result of random changes in the monthly mean rainfall. The warmer surface and possibly a longer dry season, as described here, would be likely to have a detrimental impact on survival of remaining forest and on attempts at agriculture after the removal of the tropical moist forest. Increased runoff could enhance flooding.

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REFERENCES


Dickinson, R. E.  

Dickinson, R. E., Henderson-Sellers, A., Kennedy, P. J. and Wilson, M. F.  

Fearnside, P. M.  
1982 Deforestation in the Brazilian Amazon: How fast is it occurring? Interciencia, 7, 82–88

Fearnside, P. M. and Rankin, J. M.  

Ghuman, B. S. and Lal, R.  


Henderson-Sellers, A. and Gornitz, V.  
1984 Possible climatic impacts of land cover transformations, with particular emphasis on tropical deforestation. Climatic Change, 6, 231–258

Henderson-Sellers, A., Wilson, M. F. and Dickinson, R. E.  
1986 'Zero-dimensional sensitivity studies with the NCAR CCM land surface parameterization scheme'. In 'Proc. international satellite land surface climatology project (ISLSCP) conf.', Rome, Italy, 2–6 December 1985. European Space Agency

Hunt, B. G.  

Jarvis, P. G. and McNaughton, K. G.  

Jordan, C. F.  
1982 Amazon rain forests. Am. Sci., 70, 394–401

Lettau, H., Lettau, K. and Molion, L. C. B.  

Manabe, S.  
1969 Climate and ocean circulation. I: The atmospheric circulation and hydrology of the earth's surface. Ibid., 97, 739–774

Mintz, Y.  

Myers, N.  

1980b 'Conversion of tropical moist forests'. National Academy of Sciences, Washington, DC

Newell, R. E.  

Potter, G. L., Elsasser, H. J. W., MacCracken, M. C. and Luther, F. M.  

Ratisbona, L. R.  

Salati, E.  

Salati, E., Marques, J. and Molion, L. C. B.  
1978 Origen e distribuidas chuvas na Amazonia. Interciencia, 3, 200–205
Salati, E. and Vose, P. B.

Sellers, P. J., Mintz, Y., Sud, Y. C. and Dalcher, A.

Shukla, J. and Mintz, Y.


Sud, Y. C. and Smith, W. E.

Trewartha, G. T.

Walker, J. and Rowntree, P. R.

Williamson, D. L.


Wilson, M. F.

Wilson, M. F., Henderson-Sellers, A., Dickinson, R. E. and Kennedy, P. J.


A simple biosphere model (SiB) for use within general circulation models. *J. Atmos. Sci.*, 43, 505-531


Observation of radiation exchanges above and below Amazonian forest. *Q. J. R. Meteorol. Soc.*, 110, 1163-1169


*The Earth’s problem climates*. University of Wisconsin Press


‘Description of the NCAR community climate model (CCMOB)’. National Center for Atmospheric Research, Boulder, CO. Tech. Note TN-210+STR, NTIS No. PB83231068

‘Description of NCAR community climate model (CCM1)’. National Center for Atmospheric Research, Boulder, CO. Tech. Note TN-285+STR
