TROPICAL DEFORESTATION: ALBEDO AND THE SURFACE-ENERGY BALANCE

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Abstract. Recent micrometeorological measurements for Amazonian rainforest are reviewed, emphasising those aspects of the radiation and heat balance which are likely to change with deforestation. The possible consequences of such deforestation are considered by examining the sensitivity of the surface energy balance to changes in those parameters which would be most drastically altered.

1. Introduction

Tropical deforestation on the scale at which it is currently progressing (Lal, 1987; Myers, 1988) represents a drastic change in the surface characteristics of a large and, from the climatological viewpoint, important area of the globe. The growing realisation of the seriousness of this problem is bringing process hydrologists and dynamical meteorologists together to apply the methods and measurements of the former to the forecasting models of the latter. These models of the atmosphere's global circulation (GCMs) are at present the only feasible way of incorporating the feedback between the surface and the atmosphere into forecasts of climate change. In using a GCM to forecast the effects of a vegetation change it is obviously important that the orginal vegetation should be adequately represented in a control experiment. Unfortunately there is a dearth of information on the surface characteristics of vegetation in the tropics. This is particularly the case for tropical forests, which is not surprising in view of the extreme problems which they present in the design and carrying out of field experiments. However, between September 1983 and September 1985 a joint Anglo-Brazilian micrometeorological experiment took place in a reservation of tropical rainforest near Manaus in central Amazonia. This experiment was a collaboration between the Institute of Hydrology of the UK Natural Environment Research Council, and the Brazilian National Institutes of Amazonian Research (INPA) and Space Research (INPE). In this paper we summarise some of the data obtained in this experiment, and speculate – and at this stage it can be no more than that – on some of the possible consequences that they suggest for the climatic impact of tropical deforestation.

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2. Albedo

2.1. *The Definition of Albedo*

At the top of the atmosphere the radiation from the sun is equivalent to that from a full radiator at 6000 K. The spectrum of the radiation arriving at the earth's surface is modified through absorption by atmospheric gases, particularly water vapour. It is also changed by scattering from the individual molecules of the atmosphere in clear sky conditions and from water droplets in clouds during overcast conditions. Most of the solar radiation is contained in the wavelength band 0.3–3 μm, with a maximum in the visible at 0.48 μm. Any particular spectrum will depend on the ratio of the diffuse to the direct beam radiation.

On reaching the ground most of the radiation will normally be absorbed, but some is reflected. Individual leaves will each reflect, transmit or absorb incident radiation to a different degree depending on the angle of incidence and the wavelength of the beam, as well as on the optical properties of the leaves themselves. For a complex, vegetated surface, radiation will undergo multiple reflection and transmission, such that, in general, the deeper the canopy the more radiation is trapped. For a given wavelength, λ, the radiation reflected from any particular surface will be $\rho(\lambda)S(\lambda)$, where $\rho(\lambda)$ is the reflectivity and $S(\lambda)$ the incident radiation at wavelength λ. The albedo, a, is defined as the ratio of the integrated total of reflected solar radiation over all wavelengths, to the integral of the incoming solar radiation over the same wave band (see Monteith, 1973), viz.

$$a = \int \rho(\lambda)S(\lambda)\mathrm{d}\lambda / \int S(\lambda)\mathrm{d}\lambda. \tag{1}$$

Clearly the albedo of any particular surface will be a complex function of the incoming radiation and the physical properties of the vegetation. However, Sellers (1985) has shown that, to a fair approximation, the albedo of vegetation can be modelled in terms of

(1) the scattering coefficients for leaves and soil;
(2) the leaf area index;
(3) the leaf angle distribution; and
(4) the angle of incident radiation.

By assuming previously determined empirical descriptions of leaf and canopy properties, Sellers was able to model both the magnitude and diurnal variation of the albedo of a wheat crop. Comparison with more complex models has also been successful (e.g. Kimes *et al.*, 1987).

2.2. *Measurements of Albedo*

Measured albedo is highest for dry soils and deserts; for example, Stanhill (1966) observed a value of 0.37 over desert vegetation in Israel. Agricultural crops are typically in the range of 0.18–0.25 (see Brutsaert, 1982, or Oke, 1987). Forests

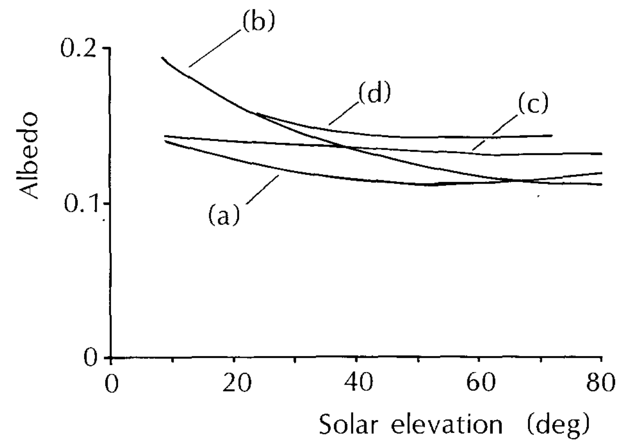

Fig. 1. The variation of the albedo of tropical forest with solar angle, as measured by (a) Shuttleworth *et al.* (1984b) and Pinker *et al.* (1980), (b) for clear sky and (c) for overcast conditions, and (d) as predicted by Sellers *et al.* (1989).

trap more radiation and therefore have lower albedos than short vegetation: typical values are in the range of 0.10–0.16. There are only a few measurements of the albedo of tropical forest. Oguntoyinbo (1970) found a value of 0.13 +/− 0.01 for rain forest in the coastal region of Nigeria. Pinker *et al.* (1980) and Pinker (1982) studied the albedo of a tropical forest in the monsoon climate of Thailand. They found a mean albedo of 0.13, but varying through the day from about 0.16 in the early morning and late afternoon to 0.11 at midday. This change was more marked on clear days, when values as high as 0.18–0.19 were recorded, but was largely suppressed on overcast days, when the albedo was almost constant at about 0.13.

A further set of measurements was made by Shuttleworth *et al.* (1984b) over the tropical rainforest at the INPA Reserva Ducke research site, near Manaus in central Amazonia. They found a mean value of 0.12. Henderson-Sellers *et al.* (1987) analysed a series of 1200 all-sky photographs taken over a period of three months at the same site. The average cloud cover was 84%, with no diurnal trend in mean cloud amount. The cloud was found to be typically high cirrus in the early morning and late afternoon, with cumuliform cloud being dominant at other times during the day. Totally clear or overcast days are thus rare in Amazonia, and Shuttleworth *et al.* did not therefore distinguish between clear sky and overcast conditions. However they did observe a slightly higher value of 0.14 at low solar angles. The results of Pinker *et al.* and Shuttleworth *et al.* are compared in Figure 1.

Using the modelling technique of Sellers (1985), Sellers *et al.* (1989) estimated an albedo of 0.14 for the Amazonian forest site. The difference, 2% of incident solar radiation, was attributed to the failure of the model to account for 'clumping' of the foliage (Baldocchi *et al.*, 1985). Essentially the model assumes horizontal homogeneity within the canopy, which is clearly not the case for a heterogeneous natural forest. These model estimates are also shown in Figure 1.

2.3. *The Likely Change in Albedo Following Deforestation*

Although the amount of albedo data available for tropical forest is small, it is consistent and agrees quite well with that expected from theoretical models. It therefore seems reasonable to suppose that the albedo of tropical forest will on average be between 0.11 and 0.13. Following deforestation there will almost always be an increase in albedo, although to what value depends, of course, on the replacement vegetation. Plantation forest may have a similar or only slightly higher albedo. Monteny *et al.* (1985) report albedo varying between 0.13 and 0.16 for a rubber plantation in the Ivory Coast. Plantations of perennial shrubs such as coffee might also be expected to have an albedo less than 0.20, as would savannah-type grassland (Oguntoyinbo, 1970). However, if, as is often the case, forest is replaced by short grass, grazed by cattle, then albedos of up to 0.25 or greater can be expected. It is therefore likely that tropical deforestation will always result in a decrease in the energy absorbed by the surface equivalent, to some 2–13% of the solar radiation before the change. A change of about 8% is probably a typical value (see also Henderson-Sellers and Gornitz, 1984).

3. The Available Energy

The surface energy balance can be expressed as

$$(1 - a)R_S + \varepsilon(R_L - \sigma T_s^4) = LE + H + G + S + P, \qquad (2)$$

where R_S and R_L are the incident solar and longwave radiation, respectively, ε and σ are the emissivity and the Stefan-Boltzman constant, respectively, and T_s is the absolute temperature of the surface. On the right-hand side of Equation (2), L is the latent heat of vaporisation, E is the evaporation and H the sensible heat flux leaving the surface, G is heat flux into the soil, and S is the change in heat stored in the biomass and the air below the level at which the radiation is measured. P is the energy used for photosynthesis. Photosynthesis is generally assumed to be only a few per cent of solar radiation, with an upper limit of 5 W m^{-2} (e.g. Stewart, 1988). Differences in photosynthesis between different vegetation types are therefore likely to give negligible changes in the total energy available at the surface.

The available energy for partition between evaporation and sensible heat flux, A, is thus

$$A = (1 - a)R_S + \varepsilon(R_L - \sigma T_s^4) - G - S. \qquad (3)$$

3.1. *Soil Heat Flux and Storage*

The soil heat flux below dense tropical forest is low. Shuttleworth *et al.* (1984b) measured the below-canopy solar radiation at the Reserva Ducke site mentioned previously and found it to be, on average, only 1.2% of the above canopy value. This was equivalent to an average flux of 4 W m^{-2} throughout the daylight hours. Soil heat flux was reported as being of the same size, with an equivalent flux of 4 W m^{-2} entering the soil during daylight and leaving the soil during the night.

The large thermal capacity of the biomass and the large depth of air around it mean that the storage term for tropical forests cannot be neglected. Moore and Fisch (1986) took measurements of humidity and air and biomass temperature for the Reserva Ducke site and estimated that the energy going into storage could exceed +/− 80 W m^{-2}. There were approximately equal contributions to the storage in the air, the biomass, and in the energy stored as latent heat through increased humidity in the air. Typical average values of S over the daylight hours were 30–40 W m^{-2}. The biomass temperature lags behind the air temperature, so that even when air temperatures were falling during the afternoon there was still a net flux of energy into storage. Combining the storage and soil heat flux terms together therefore means an equivalent daylight average energy flux of some 40 W m^{-2} may not be available for evaporation during daylight. The photo-response of stomata causes them to close at night, and there is consequently little transpiration during darkness. In Amazonia there is little seasonal variation in soil temperature, so the energy stored during the day in the biomass, air and soil is consequently lost at night by radiative cooling to space.

In contrast to forest the small biomass of short vegetation ensures a small energy storage term, but the soil heat flux, although very variable, will almost always be appreciable. For sparse crops with much of the soil exposed, soil heat fluxes in excess of 100 W m^{-2} are possible, but even when canopy cover is complete, soil heat fluxes will be greater for short vegetation than for forest. For example, Oliver *et al.* (1987) found values approaching 60 W m^{-2} under dense grass, but peak values of less than 10 W m^{-2} under an adjacent dense forest. It seems likely that the sum of the soil heat flux and storage terms may thus be of similar magnitude for forest as for short vegetation, although the contributions from the two constituents will be very different.

3.2. *Evaporation and Sensible Heat*

Using the electrical analogue to turbulent energy transport the sensible heat flux can be expressed as

$$H = -\rho C_p(T - T_s)/r_a, \tag{4}$$

where ρ is air density, C_p is the specific heat of air at constant pressure, T is air temperature, and r_a is the aerodynamic resistance. The evaporation can be expressed as

$$E = -\rho(q - q_{sat}(T_s))/(r_a + r_s), \tag{5}$$

where q is the specific humidity, $q_{sat}(T_s)$ is the saturated humidity at the surface temperature, and r_s is the extra 'surface' resistance encountered by water vapour in diffusing through the plants' stomata. Equations (4) and (5) can be combined with the surface energy balance equation to eliminate the surface temperature, giving the Monteith-Penman equation

$$LE = [\Delta A + \rho C_p(q - q_{sat}(T))/r_a]/[\Delta + C_p(1 + r_s/r_a)/L], \tag{6}$$

where Δ is the slope of the saturated specific humidity curve at the mean of the air and surface temperatures.

For a given set of meteorological conditions the ratio of evaporation to sensible heat flux will depend on r_a and r_s. The aerodynamic resistance, r_a, depends primarily on the roughness of the surface, windspeed and the atmospheric stability. For the Reserva Ducke site Shuttleworth (1988) deduced a relationship

$$r_a = 34.2/u \text{ s m}^{-1}, \tag{7}$$

where u is the windspeed in m s^{-1}. This implies typical values of some 10 s m^{-1}. For short vegetation values of order 100 s m^{-1} are more typical.

The stomata of plants respond to a variety of controls, and it is possible to model surface resistance in terms of these variables. For example the Jarvis-Stewart model (Stewart, 1988) describes surface conductance (the reciprocal of surface resistance) in terms of a maximum conductance, which is modified by normalised functions of leaf area, solar radiation, humidity deficit, temperature, and soil moisture deficit. Models of this general type are highly suitable for incorporation into the land surface description of GCMs, since they quantify the feedback between the transpiration and the atmospheric variables which control it. Dolman *et al.* (1991) have derived the Jarvis-Stewart parameters for the Reserva Ducke data, and Sellers *et al.* (1989) report the first stage of applying a similar model for use in a GCM. Their model was also calibrated against the Reserva Ducke data. Alternatively a simple empirical description can be used which describes the variation of surface conductance as a function of time of day (Shuttleworth *et al.*, 1984a; Shuttleworth, 1988). In this case the environmental controls on stomatal conductance are implicit in the variation with time of day. In a tropical climate with little seasonal or day-to-day variation, and little change in soil moisture, such a model is likely to work well (Dolman *et al.*, 1991).

Following deforestation, despite the albedo change giving reduced available energy, short grass, with complete canopy cover and an adequate supply of water, is likely to have a lower surface resistance and therefore transpire at a higher rate than the forest (see Shuttleworth *et al.*, 1984a). Although rainforests grow in areas of generally frequent and copious rainfall there are few areas which do not have a period of the year with a reduced and less frequent rainfall, and concurrent higher evaporative demand. During these periods the shallower rooting depth of short vegetation, together with the effects of soil degradation, which often accompany deforestation, will increase the likelihood of water stress decreasing the transpiration of the short vegetation. These periods will also be associated with the greatest difference in albedo as the short vegetation can change from a full green cover of relatively low albedo to a high albedo, senescent stubble condition.

When a plant canopy is wet during or following rainfall the surface resistance becomes zero and evaporation is controlled solely by the aerodynamic resistance and the meteorological variables. The roughness of the forest surface gives a low aerodynamic resistance, efficient turbulent transfer away from the surface, and hence high evaporation rates for intercepted rainfall. This is important because it results in a rapid re-evaporation of a significant proportion of the rain, during or soon after the rain storm, which is then available to feed the convective system with further water vapour and energy in the form of latent heat.

Rainfall interception can be measured as the difference between gross rainfall, measured above the canopy, and net rainfall, measured as throughfall and stemflow on the forest floor. Unfortunately in tropical forest this apparently simple measurement problem is fraught with difficulties (see Jackson, 1971; Lloyd and Marques, 1988; and Shuttleworth, 1989). These result from the high rainfall rates, the small difference between the gross and net rainfall and the sampling problems associated with measuring throughfall under a very variable canopy. Many of the numerous studies of tropical interception loss have produced results where the measurement and sampling errors are so large as to make the results of questionable use. At Reserva Ducke Lloyd *et al.* (1988) measured an interception loss of just 9 +/− 4% of the incident rainfall.

Interception loss can be modelled from a knowledge of the canopy structure parameters, the forest roughness and the meteorological variables above the canopy. A family of models exist based on that pioneered by A. J. Rutter (Rutter *et al.*, 1971) which calculate a running water balance of the canopy, using Equation (6) to estimate the evaporation. The most important of the forest structure parameters is the canopy capacity. As defined by Rutter, this is the amount of water left on the canopy at the end of a storm, in conditions of zero evaporation, and when all drip from the canopy has ceased. It is thus the minimum water necessary to saturate the canopy. Values are typically about 1 mm. Lloyd *et al.* (1988) found a value of 0.74 mm for the Reserva Ducke site. This store of water will be evaporated rapidly at the end of a storm.

For short vegetation the higher aerodynamic resistance reduces the evaporation of intercepted water during storms. The canopy capacity of short vegetation does not differ greatly from that of forests (see Rutter, 1975), but it will of course take longer to dry out after rainfall has ceased. Transpiration will therefore be suppressed for a greater time after the ends of storms, and the overall water loss during rainy periods further reduced.

4. Discussion

Charney (1975) proposed a mechanism by which an increasing albedo resulted in the atmosphere above an arid area becoming a net radiative sink, with increased transfer of radiation back to space at the top of the atmosphere. This leads to enhanced subsidence of the atmosphere with consequent reductions in clouds and rainfall. Subsequent experiments on the sensitivity of GCMs to albedo have modelled this effect. Examples are the study by Cunnington and Rowntree (1986) and other studies reviewed by Rowntree (1988). However, as pointed out by Henderson-Sellers *et al.* (1988), these studies should be regarded as GCM sensitivity analyses rather than climate change forecasts. Only when all the appropriate parameters are allowed to change simultaneously can it be hoped to make any prediction of the overall consequences of a large-scale vegetation change. Studies of this type are rare: Henderson-Sellers and Gornitz (1984) and Dickinson and Henderson-Sellers (1988) have carried out GCM experiments on the effects of large-scale deforestation of the Amazon basin. These and other studies are discussed by Henderson-Sellers *et al.* (1988) and Shuttleworth and Dickinson (1989). The general conclusion has been that deforestation will reduce evaporation. Without any change in rainfall this should increase runoff from catchments, and for small clearings this should be expected. However, for large-scale clearing this may not be the case, since although the situation with regard to rainfall is still unclear, an estimated 50% of the rainfall in central Amazonia is recycled (see Salati, 1987), and it seems likely that a large-scale reduction in evaporation would result in reduced rainfall.

Although climate forecasts can only realistically be made with GCMs, it can be useful in understanding the processes involved to look at the sensitivity of the instantaneous energy balance as defined by Equation (6). Consider a typical midday in the humid tropics, with dry-canopy conditions and prescribed meteorologi-

TABLE I: Estimates of the available energy, A, the evaporation, LE, the sensible heat flux, H, and the surface temperature, T_s, made using Equations (3) and (6). During dry conditions solar radiation was taken as 600 W m^{-2}, longwave downward radiation as 400 W m^{-2}, air temperature as 28 °C, humidity deficit 8 g kg^{-1} and the storage/soil heat flux term as −50 W m^{-2}. Under wet conditions solar radiation was reduced to 100 W m^{-2}, air temperature was taken as 24 °C, humidity deficit as 2 g kg^{-1} and the storage/soil heat flux over the short vegetation as −8.3 W m^{-2}. The storage/soil heat flux term for the forest was not changed from the dry condition value

	Albedo	r_a (s m^{-1})	r_s	A –	LE (W m^{-2})	H –	T_s (°C)
Dry forest	0.12	10	80	413	297	115	29.0
Dry short vegetation	0.20	100	80	336	270	66	33.6
Wet forest	0.12	10	0	10	165	−155	22.7
Wet short vegetation	0.20	100	0	39	44	−5	23.6

cal variables of 600 W m^{-2} of solar radiation, 400 W m^{-2} of downward longwave radiation, air temperature of 28 °C and specific humidity deficit of 8 g kg^{-1}. For the forest let the albedo be 0.12, r_s be 80 s m^{-1} and r_a be 10 s m^{-1}. By estimating an initial value of the surface temperature Equation (6) can be used in a series of iterations to derive the values of available energy, transpiration, heat flux and surface temperature given in Table I. Emissivity was assumed to be 0.98 and the energy going into storage/soil heat flux was assumed to be 50 W m^{-2}. As discussed previously the two major changes likely to occur on deforestation are an increase in the albedo to 0.20 and an increase in the aerodynamic resistance to 100 s m^{-1}. The results of repeating the calculations with these values, but leaving the others unchanged, are also given in Table I. The available energy is reduced by 19%, the transpiration by 9% and the sensible heat flux by 43%. In practice the different vegetation and the feedback between the fluxes, the atmosphere and the vegetation will mean that the driving variables will change – so that these calculations should be regarded as indicative rather than definitive. However, they make the point that there is likely to be less energy going into the atmosphere as latent and sensible heat, and more energy lost to space by reflection and emission of radiation. The reduction in sensible heat flux may also have implications in terms of the convective energy available to develop the atmospheric boundary layer. André *et al.* (1989) have shown that, at least in one particular case, the difference in sensible heat fluxes rising from areas covered with forest and short vegetation resulted in the formation of a deeper boundary layer and consequently more cloud formation over the forest compared to the area of short vegetation.

Under wet conditions more typical values would be solar radiation of 100 W m^{-2}, temperature of 24 °C, and a humidity deficit of 2 g kg^{-1}. The storage/soil heat flux term would be reduced under the short vegetation in response to the lower radiation, but not for the forest, where the changing temperature might be expected to maintain the storage component. The results of reducing the soil heat flux under the short vegetation in proportion to the reduction in radiation, but changing the other variables and parameters as before, are also given in Table I. For the wet forest Equation (6) predicts available energy increases over the forest from 10 to 39 W m^{-2} over the short vegetation. This reflects the reduction in soil heat flux. The evaporation of intercepted water changes by a factor of four, from 165 over the forest to 44 W m^{-2} over the short vegetation, and the sensible heat flux from −155 W m^{-2} into the forest to −5 W m^{-2} into the short vegetation. A flux of 165 W m^{-2} evaporation is equivalent to 0.24 mm hr^{-1}, which can be compared with the value of 0.21 mm hr^{-1} estimated by Lloyd *et al.* (1988) for the average evaporation during rainfall for their two-year run of tropical forest data. It demonstrates the high rate of intercepted rainfall evaporation from forests, which is fed by advected energy from dry areas outside the storm. Lloyd *et al.* found that at the Reserva Ducke site the interception loss was approximately equally divided between evaporation during storms and evaporation from the canopy store after rainfall had ceased. Discounting any possible change in the canopy store, the evaporation of intercepted rainfall from short vegetation should therefore be expected to be reduced to some 60% of its forest value, that is to some 5–6% of rainfall. In central Amazonia this would amount to a reduction of some 100 mm a year. In addition the longer time taken for the canopy store to dry out after rain would suppress, and therefore further reduce, the transpiration loss.

5. Conclusion

The arguments advanced in the previous section are necessarily approximate and speculative, nevertheless they lead to the overall conclusion that deforestation will result in a reduction of the amount of radiative energy being absorbed by the earth's surface. This in turn will usually result in less evaporation, and for closed continental systems may well lead to a reduction in rainfall. The consequences near the edge of continents, and for islands, may well be very different (see Shuttleworth, 1989).

More quantitative forecasts of the effects of tropical deforestation on climate need to address the feedback between vegetation and the atmosphere. This requires not only improved formulation of the land surface within climate models, but also more basic data against which such formulations can be calibrated. The need for this research cannot be over-emphasised.

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