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THE ROLE OF CLOUDS IN THE SURFACE ENERGY BALANCE OVER THE AMAZON FOREST

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ABSTRACT

Deforestation in the Amazon region will initially impact the energy balance at the land surface through changes in land cover and surface hydrology. However, continuation of this human activity will eventually lead to atmospheric feedbacks, including changes in cloudiness which may play an important role in the final equilibrium of solar and terrestrial radiation at the surface. In this study, the different components of surface radiation over an undisturbed forest in the Amazon region are computed using data from the Amazon region micrometerological experiment (ARME). Several measures of cloudiness are defined: two estimated from the terrestrial radiation measurements, and one from the solar radiation measurements. The sensitivity of the surface fluxes of solar and terrestrial radiation to natural variability in cloudiness is investigated to infer the potential role of the cloudiness feedback in the surface energy balance. The results of this analysis indicate that a 1% decrease in cloudiness would increase net solar radiation by *ca*. 1.6 W/m^2. However, the overall magnitude of this feedback, due to total deforestation of the Amazon forest, is likely to be of the same order as the magnitude of the decrease in net solar radiation due to the observed increase in surface albedo following deforestation. Hence, the total change in net solar radiation is likely to have a negligible magnitude. In contrast to this conclusion, we find that terrestrial radiation is likely to be more strongly affected; reduced cloudiness will decrease net terrestrial radiation; a 1% decrease in cloudiness induces a reduction in net terrestrial radiation of *ca*. 0.7 W/m^2; this process augments the similar effects of the predicted warming and drying in the boundary layer. Due to the cloudiness feedback, the most significant effect of large-scale deforestation on the surface energy balance is likely to be in the modification of the terrestrial radiation field rather than the classical albedo effect on solar radiation fields. The net effect of clouds is to reduce net radiation; a 1% increase in cloudiness induces a reduction in net radiation of *ca*. 1 W/m^2. The implications of this negative feedback on large-scale land-atmosphere interactions over rainforests are discussed. © 1998 Royal Meteorological Society.

KEY WORDS: surface energy balance; cloud feedbacks; Amazon forest; deforestation

1. INTRODUCTION AND BACKGROUND

Deforestation modifies the different components of the energy balance at the land-atmosphere boundary through changes in the physical properties of the land surface such as albedo, roughness, and root zone depth. These in turn modify the radiative energy input to the surface, defined as net radiation, as well as the partitioning of that energy into sensible and latent heat. Net radiation over a region is composed of both terrestrial (long-wave) and solar (short-wave) radiation; the impact of deforestation on the energy balance of any region is manifested through changes in both of these components. One of the observed effects of deforestation is the increase in surface albedo from *ca*. 12–13% to almost 16–18% (Bastable *et al*., 1993; Culf *et al*., 1995). This change in albedo results in reflecting a greater fraction of incoming solar radiation, and hence favors a smaller amount of net solar radiation. Another observed impact of

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deforestation is the increase in temperature of the surface and the lower layers of the atmosphere, (Bastable *et al*., 1993). This effect results in higher upwards flux of terrestrial radiation and therefore less net terrestrial radiation. Hence, in general, deforestation reduces net radiation at the land-atmosphere boundary. Both field observations (Bastable *et al*., 1993; Culf *et al*., 1996) and numerical simulations (Nobre *et al*., 1991; Dickinson and Kennedy, 1992; Lean and Rowntree, 1993; Eltahir and Bras, 1994) agree on the changes in solar and terrestrial radiation components.

Other effects of deforestation, such as the changes in rainfall and atmospheric circulation, may depend on the scale of deforestation. In a recent study, Bastable *et al*. (1993) made continuous and simultaneous measurements of meteorological variables over undisturbed forest and cleared ranch land. The undisturbed site is Reserva Ducke near Manaus, Amazonas, Brazil. The cleared forest site is Fazenda Dimona, a cattle ranch located at *ca*. 100 kilometers north of Manaus. The observations, which included radiation fluxes, precipitation, and cloud cover, span a period of 60 continuous days from the middle of October 1990 to the middle of December 1990. The daily average of both net solar and net radiation was larger over the forest than over the clearing; the difference in net radiation was greater than the difference in net solar radiation. This suggests that the difference in net radiation between the forest and the cleared site is accounted for by changes in both solar and terrestrial radiation. However, these effects were observed over a small-scale clearing site; the size of the clearing at Fazenda Dimona is *ca*. 10 km^2, and is effectively an 'island' of deforestation in a sea of forest. The impact of such a small anomalous pocket on the atmospheric circulations is certainly negligible; this is confirmed by observations at the Fazenda Dimona Site. In particular, no changes were detected between the patterns of rainfall and cloudiness over the cleared and undisturbed sites. At such small scales, heterogeneity in land cover introduced by deforestation may even enhance triggering of moist convection and formation of clouds (Rabin *et al*., 1990; Chu *et al*., 1994; Cutrim *et al*., 1995). However, numerical studies of deforestation, with scales of *ca*. 10^5–10^6 km^2, suggest that large-scale deforestation may result in decreased evaporation and precipitation (Nobre *et al*., 1991; Dickinson and Kennedy, 1992; Lean and Rowntree, 1993, Eltahir and Bras, 1994), weakened atmospheric circulation and decreased convergence of water vapor into the region (Eltahir and Bras, 1993), and a reduction in cloudiness (Lean and Rowntree, 1993; Dickinson and Kennedy, 1992). Since cloud cover regulates the solar and terrestrial radiation components at the land-surface, the cloudiness feedback may play an important role in the surface energy balance.

Net solar radiation over any surface is determined primarily by two factors: the land-surface albedo (which determines how much radiation is reflected at the Earth's surface), and radiatively active constituents of the atmosphere, which reflect, absorb, and scatter solar radiation. In particular, clouds which are important constituents of the atmosphere reflect, scatter, and absorb solar radiation. Cloud cover, defined as the fraction of sky covered by clouds (Monteith, 1973), regulates the amount of solar radiation traveling from the top of the atmosphere to the Earth surface. Hence, previous studies have tried to link percentage cloud cover and cloud type to the fraction of solar insolation reaching the Earth's surface from the top of the atmosphere, defined as the solar insolation ratio (Cunniff, 1958). Such studies have often developed relations for specific cloud types and elevations; this requires extensive observations of the cloud cover present. Henderson-Sellers *et al*. (1987) presented observations of cloud cover for the period March 1985–May 1985 and analyzed both the diurnal cycle of cloud cover and its correlation to other meteorological variables. These observations were taken near Manaus in the Amazon region. In particular, cloud cover was compared with the ratio of measured surface solar radiation to calculated top-of-the-atmosphere insolation. A significant though variable relationship between cloud cover and the solar insolation ratio was observed.

In this paper net terrestrial radiation over any land surface is defined as the downwards terrestrial radiation minus the upwards terrestrial radiation. Upward terrestrial radiation is the amount of radiation emitted by the land surface and is therefore almost an exclusive function of the surface temperature as described by the Stefan-Boltzmann law. Much of this radiation is absorbed by the atmosphere before reaching outer space; the downward terrestrial radiation is the amount of radiation emitted back towards the surface by radiatively active gases and aerosols in the atmosphere. Since water vapor and carbon dioxide are the most significant of these gases, their concentration in the atmosphere strongly determines emissions of terrestrial radiation. Clouds absorb a significant amount of the upward terrestrial radiation flux, and emit radiation back towards the surface. Consequently, clouds are important in determining the downward terrestrial radiation and hence the net terrestrial radiation at the surface. Because both net terrestrial and net solar radiation are strongly dependent on the cloud field, clouds have an important role in the surface energy balance.

The focus of this paper is on the role of clouds in the surface energy balance which will be inferred by studying simultaneous observations on natural variability of cloudiness measures, as well as the surface radiation field. The example of large-scale deforestation has been chosen to illustrate how changes in cloudiness may impact surface radiation fields. However, it should be emphasized that the conclusions of this study are general in nature, and should help in enhancing our understanding of the impact of deforestation on the surface radiation fields, regardless of the size of the deforested area. Deforestation is regarded as a perturbation in the distribution of vegetation, hence by studying the response of the atmosphere to deforestation it is hoped to define the role of vegetation distribution in the tropical circulations and climate, (Eltahir, 1996).

The Amazon region micro-meteorological experiment (ARME) data set is used in this study and is described in section 2. Several measures of cloudiness are defined in section 3. These measures are estimated for the Amazon region and presented in section 4. The sensitivities of the different components of the surface radiation field to natural variability in the cloudiness measures are discussed in section 5. The role of the cloudiness feedback in the surface energy balance following large scale deforestation is discussed in section 6. The conclusions of the study are stated and discussed in section 7.

2. DATA

The ARME data set was gathered during the period October, 1983–August, 1985 over the preserve Reserva Foresta Ducke, located at 25 km from Manaus, Amazonas, Brazil. This is an undisturbed region representative of much of the Amazon forest, composed of a wide range of species. The main canopy height averages 30–35 m, with occasional trees extending up to *ca*. 40 m. Precipitation over the site is highly seasonal, with a maximum normally occurring in March and a minimum in August. Evaporation is uniform throughout the year and extremely high, exceeding the potential evaporation by 10% in the wet season and approaching 90% of the potential evaporation during dry periods (Shuttleworth, 1988). Several relevant studies on evaporation, land surface processes, and cloudiness have been conducted at the same site and are reported by Henderson-Sellers *et al*. (1987), Shuttleworth (1988) and Bastable *et al*. (1993).

The observations during ARME include incoming solar radiation, net all-bands radiation, temperature, wet-bulb depression, wind speed and direction, and precipitation. During the period from March to June of 1984, the radiometers recording net radiation cracked, and the data from that period has been replaced using a best-fit equation and observations of solar radiation. The following relationship is calibrated using the rest of the data;

$$R_n = \{0.782 - 0.028 \cos(2\pi[N_d - 31])/365)\} S_O - 16 + 7 \cos(2\pi[N_d - 31]/365) \quad (1)$$

where R_n is net radiation, N_d is day number of the year, and S_O is measured hourly incident solar radiation (Shuttleworth, 1988). This formula was used in determining net radiation for the specified period.

In this study, an hourly radiation balance was performed using ARME data. Net radiation and incident solar radiation are measured variables; albedo, A, is estimated as a function of solar altitude angle, α;

$$A = 15.09 - 0.136\alpha + 0.00123\alpha^2 \quad (2)$$

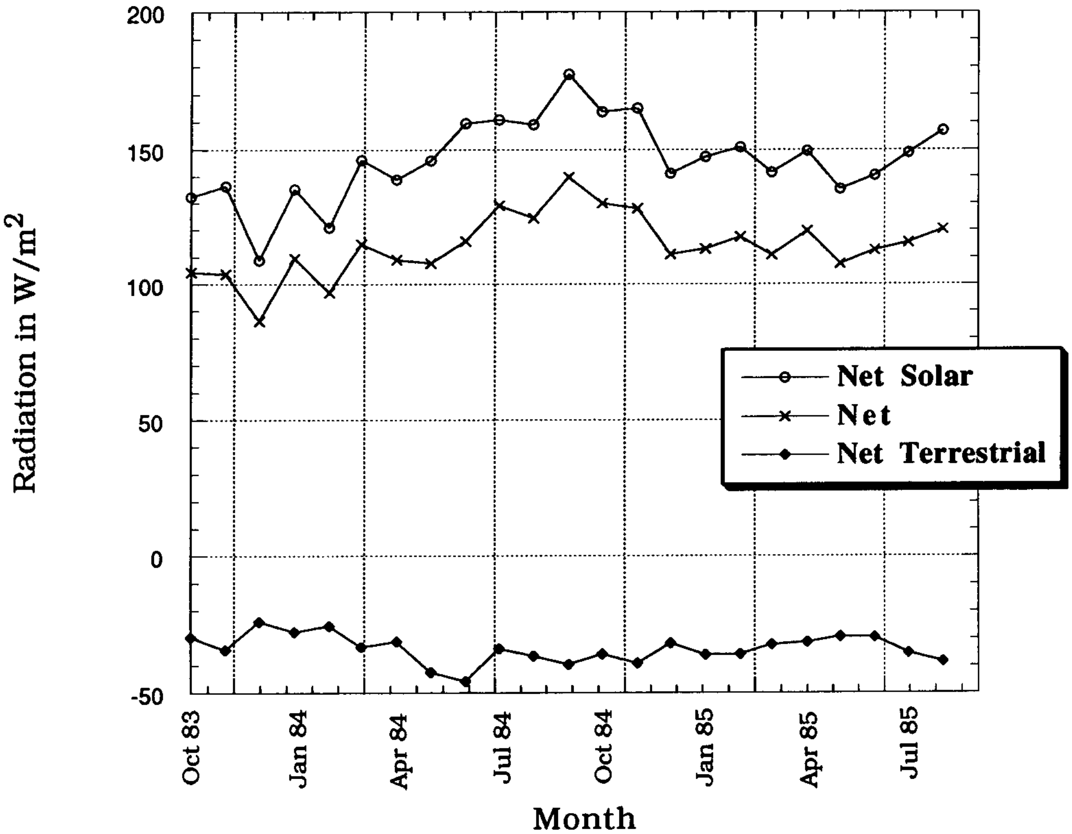


Figure 1. Monthly averages of radiation components over Amazon forest, 1983–1985

This relationship was calibrated by Shuttleworth *et al.* (1984) using observations of albedo. Hourly upward terrestrial radiation, L_u, was computed from the Stefan-Boltzmann law, using the recorded air temperature and assuming emissivity of the surface as unity. From these variables, downward terrestrial radiation is computed from the statement of surface radiation balance

$$R_n = (1 - A)S_O + L_d - L_u \tag{3}$$

or rearranging,

$$L_d = R_n - (1 - A)S_O + L_u \tag{4}$$

where R_n, A, S_O, and L_u are all either observed or estimated from observations. Monthly averages of net solar radiation, net terrestrial radiation, and net radiation are shown in Figure 1; a summary of the meteorological statistics is presented in Table I.

Table I. Meteorological statistics from the ARME study

Variable	Monthly average	S.D. of monthly averages	Units
Solar insolation	849.8	29.7	W/m^2
Solar radiation	165.3	19.7	W/m^2
Net solar radiation	144.1	17.2	W/m^2
Upwards terrestrial radiation	447.1	9.97	W/m^2
Downwards terrestrial radiation	413.7	8.34	W/m^2
Net terrestrial radiation	−33.4	5.47	W/m^2
Net radiation	112.6	13.3	W/m^2
Temperature	24.8	1.8	°C
Precipitation	217.3	108.2	mm/month

3. MEASURES OF CLOUDINESS

Solar radiation at the surface is ultimately limited by the solar radiation at the top of the atmosphere. The amount of the solar radiation flux at the top of the atmosphere depends on the solar constant and the solar altitude angle, which is a function of latitude, time of day, and day of the year as described by Eagleson (1970);

$$\sin \alpha = \sin \delta \sin \Phi + \cos \delta \cos \Phi \cos \tau \tag{5}$$

where α is the solar altitude angle, δ is the latitude, Φ is the solar declination angle (a function of the day of the year), and τ is the hour angle of the sun (a function of the time of day). The flux at the top of the atmosphere is then given by:

$$I^* = I \sin(\alpha) \tag{6}$$

where $I = 1353$ W/m^2 and is the solar constant (Liou, 1980).

Two processes are primarily responsible for regulating the flux of solar radiation traveling through the atmosphere towards the surface. The first of these is reflection and absorption by clouds. Cloud albedo is the fraction of the incoming solar radiation reflected by clouds. This important variable is a function of the type, density, and extent of cloud cover. The second process is absorption and scattering of the incoming solar radiation by other atmospheric aerosols and atmospheric gases. The atmospheric water vapor content plays an important role in this process, since water vapor selectively absorbs a significant fraction of incoming solar radiation at certain windows of the electromagnetic radiation spectrum. Although atmospheric humidity varies at daily to seasonal time scales, the magnitude of this variation over equatorial regions is small (Peixoto and Oort, 1993). Hence, we assume that absorption and scattering of solar radiation by atmospheric gases account for an insignificant fraction of the variability in incident solar radiation at the surface. This assumption is particularly valid if the variability in atmospheric water vapor is compared to the variability in cloud cover and precipitation over equatorial regions. There can be significant seasonal variation in atmospheric aerosol concentration over the Amazon as a result of burning of pasture during the dry season. However, we neglect those effects and assume that clouds are the major contributors to the variability of solar radiation at the surface. Other atmospheric constituents may be important in determining the average solar radiation received at the surface, but without contributing significantly to the variability of that radiation. Thus, one measure of cloudiness will be directly related to the solar insolation ratio (Cunniff, 1958; Henderson-Sellers *et al.*, 1987). This measure will be defined here by:

$$\beta = 1 - \eta = \frac{I^* - S_O}{I^*} \tag{7a}$$

where S_O is the measured incident solar radiation at the surface, I^* is defined by Equation (6), and η is the solar insolation ratio. In theory, β approaches a value of 0 for a totally clear sky (strictly speaking, in absence of the atmosphere), and approaches a value of 1 for a totally cloudy sky. However, since I^* is the solar flux at the top of the atmosphere, an assumed constant fraction of that flux will be eliminated through absorption and scattering by atmospheric gases. Since this reduction is included in the definition of solar insolation ratio, β is a measure of how much solar radiation is reduced by both clouds and the atmosphere. Assuming that under totally clear-sky conditions, incident solar radiation is equivalent to I^*, and that under totally cloudy-sky conditions incident solar radiation flux is zero, Equation (7b) can be expressed as:

$$\beta = \frac{S_O^{\text{clear}} - S_O}{S_O^{\text{clear}} - S_O^{\text{cloudy}}} \tag{7b}$$

which expresses β as a measure of the difference between clear sky solar radiation, S_O^{clear}, and observed solar radiation, S_O, normalized by the difference between S_O^{clear} and cloudy sky solar radiation, S_O^{cloudy}. This definition of β will be useful in comparing this measure to the other cloudiness measures that will be developed based on terrestrial radiation.

The terrestrial radiation field at the surface is also regulated by clouds. The upward flux of terrestrial radiation from the land surface is related to temperature, T_a, by the Stefan-Boltzmann law

$$L_u = \varepsilon\sigma T_a^4 \tag{8}$$

with the Earth's emissivity, ε, assumed equal to 1. As suggested by Bastable *et al.* (1993), using the air temperature in Equation (8) underestimates L_u; setting ε to 1 (artificially larger than the true value by a few percent) may offset that error. It is difficult to quantify the error introduced by these assumptions. The downward flux of terrestrial radiation, L_d, is proportional to cloudiness as demonstrated by Monteith (1973), and verified, roughly, for the Amazon region by Shuttleworth *et al.* (1984). The two measures, C_1 and C_2, were used by Monteith (1973) and Shuttleworth *et al.* (1984) for describing high and low clouds, respectively. They assume a linear and a parabolic relationship between cloudiness and the downward terrestrial radiation flux

$$L_d = (1 - C_1)\varepsilon_a\sigma T_a^4 + C_1(\sigma T_a^4 - 9) \tag{9}$$

$$L_d = \varepsilon a(1 + 0.2C_2^2)\sigma T_a^4 \tag{10}$$

The apparent emissivity of clear sky, ε_a, is described by

$$\varepsilon_a = 0.65 + 0.007(T_a - 273) \tag{11}$$

and has an average value of *ca.* 0.8 for the average temperature of 24.8°C over the measurement period of ARME. Although the empirical equations (9)–(11) were first proposed by Monteith (1973) for typical conditions in midlatitudes, Shuttleworth *et al.* (1984) demonstrated the applicability of these equations to tropical conditions. Hence, it is the authors' belief that the two variables C_1 and C_2 provide approximate but reasonable measures of cloudiness over the Amazon forest.

In the following, the cloudiness measures, which are based on the terrestrial radiation field, will be compared to the measure that is based on solar radiation. Let us focus on Equation (9). This equation implies that downwards flux of terrestrial radiation from a totally clear sky ($C_1 = 0$) is given by

$$L_d^{\text{clear}} = \varepsilon_a\sigma T_a^4 \tag{12}$$

and for a totally cloudy sky ($C_1 = 1$), this flux approaches

$$L_d^{\text{cloudy}} = \sigma T_a^4 - 9 \tag{13}$$

where 9 is an empirical constant (Monteith, 1973). This constant accounts for the fact that a totally cloudy sky may still emit less radiation than that emitted by a black body at the same temperature. For comparison with Equation (7b), Equation (9) was re-arranged, resulting in

$$C_1 = \frac{\varepsilon_a\sigma T_a^4 - L_d}{\varepsilon_a\sigma T_a^4 - (\sigma T_a^4 - 9)} = \frac{L_d^{\text{clear}} - L_d}{L_d^{\text{clear}} - L_d^{\text{cloudy}}} \tag{14}$$

which defines C_1 as the difference between observed downwards flux of terrestrial radiation and the corresponding flux for totally clear sky conditions, normalized by the difference between downwards flux of terrestrial radiation for a totally cloudy sky and that for a totally clear sky. Hence C_1 is a normalized measure of downwards flux of terrestrial radiation just as β is a normalized measure of incident solar radiation.

4. ANALYSIS OF CLOUDINESS OVER THE AMAZON

This section describes analysis of cloudiness over the Amazon forest using the measures that have been described in the previous section. The daily average of L_d is used to compute C_1 and C_2 using Equations (9) and (10). Solar insolation at the top of the atmosphere is computed from Equations (5) and (6) using a 6-min time step. The daily solar insolation is used with the daily averaged measured

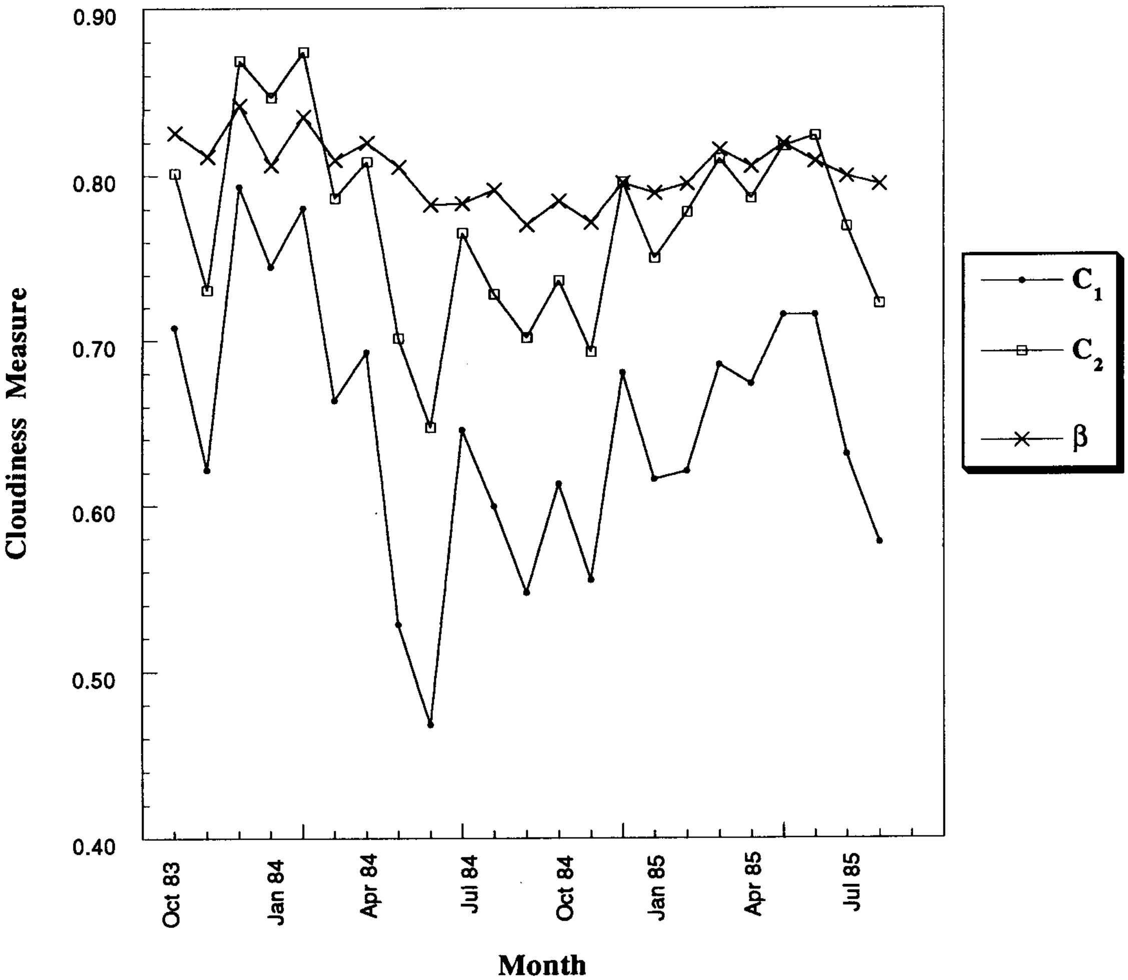


Figure 2. Monthly averages of cloudiness measures C_1, C_2, and β over Amazon forest, 1983–1985

incident solar radiation at the surface to estimate β using equation Equation (7a). Figure 2 shows the monthly variability of the three cloudiness measures.

The monthly variability of the cloudiness measures are compared to the climatology of cloud cover over the Amazon. Ratisbona (1976) reported a yearly average value of cloud cover for the Amazon of 0.66; of the three measures, C_1 compares best with a value of 0.65, compared with 0.78 for C_2 and 0.81 for β. At shorter time scales, Henderson-Sellers *et al.* (1987) observed an average cloud cover of 0.84 for the wet period, March 1985–May 1985. The average values of C_2 of 0.80 and β of 0.81 for that same period agree with observed values for cloud cover. However, the seasonal variability of C_1 best mimics the seasonal variability of cloud cover. Bastable *et al.* (1993) reported a cloud cover of 0.48 during a dry period. For the dry periods of June–August for each of the 2 years of ARME measurements, C_1 averages 0.61, whereas C_2 and β have averages of 0.74 and 0.79, respectively. In this case, C_1 compares better with observations in both absolute value and relative magnitude. Overall, the generally higher variation in C_1 (range of 0.33 vs. 0.23 for C_2 and 0.09 for β), the better the agreement between the yearly averages of C_1 and observed cloud cover, and the fact that relative seasonal variations of C_1 most closely agree with observed variations of cloud cover for all periods indicate that, among the three cloudiness measures, C_1 most closely mimics the climatology of cloud cover.

In the following, the question as to why the three cloudiness measures are significantly different from observed cloud cover is addressed. The latter is defined as the fraction of the sky covered by clouds. Here too, the observations of Henderson-Sellers *et al.* (1987) are referred to. As that study points out, there are several explanations for the discrepancy between the solar insolation ratio and observed cloud cover. One limitation of cloud cover as a cloudiness measure is its non-unique mapping of the atmospheric conditions; this is due to the low correlation between cloud cover and cloud density. Solar and terrestrial radiation are both affected by much more than just fractional cloud cover. Cloud depth, height, type, and density all greatly affect the radiation field in the atmosphere; in addition, other properties of the atmosphere such as water vapor content also affect the radiation fields. Hence, η and β should not be taken as indicators of pure cloud cover, but instead may be considered lumped measures of how atmospheric conditions affect the amount of incoming solar radiation. C_1 and C_2 can likewise be taken as gauges of the effect of atmospheric conditions on terrestrial radiation rather than indicators of percentage cloud cover. 'Atmospheric conditions' include, but are not limited to, cloud cover, height, depth, and density, as well as atmospheric water vapor content. It is then appropriate to redefine C_1, C_2, and β as measures of effective cloudiness, which is defined here as the total effect of the atmosphere on the radiative energy components at the land surface.

5. SENSITIVITY OF SURFACE RADIATION FIELDS TO NATURAL VARIABILITY IN CLOUDINESS MEASURES

The role of clouds in the surface energy balance is examined in this section by considering how solar and terrestrial radiation are each affected by cloudiness. This requires quantification of the sensitivities of the two components of net radiation (net solar radiation and net terrestrial radiation) to natural variability in the three cloudiness measures. The term sensitivity is used here to mean the derivative of the radiation component with respect to the cloudiness measure. Estimates of the sensitivities of net radiation, net solar radiation, and net terrestrial radiation to effective cloudiness are obtained through linear regression analysis. Plots of net terrestrial radiation, net solar radiation, and net radiation versus, β, C_1 and C_2 are presented in Figures 3–5. The magnitudes of the sensitivities are markedly different for β in comparison to C_1 and C_2. Whereas sensitivity of net radiation to a 1% change in C_1 is *ca.* 1 W/m^2, the corresponding sensitivity to a 1% change in β is almost 6 W/m^2 (the cloudiness measures have units of percent, hence a 1% change is used in this paper to indicate an absolute change of magnitude 0.01, no normalization is implied). This difference is due to the relatively low variability of β compared with C_1 and C_2. Since β is a relatively stable measure of cloudiness, a small change in β could be interpreted as a significant change in cloudiness.

Before performing the sensitivity analysis, it is important to note that C_1, C_2, and β are explicit functions of the radiation components that are used to estimate them (see Equations (7a), (9) and (10)). Hence, estimation of the sensitivity of solar radiation to β and the sensitivity of terrestrial radiation to C_1 or C_2 may not provide any new information. In order to provide meaningful estimates of the sensitivity of a given surface radiation component to cloudiness, that estimate has to be independent of the definition of the cloudiness measure. Hence, the sensitivity of terrestrial radiation to cloudiness will be estimated using an effective cloudiness measure which is estimated from solar radiation measurements; and the sensitivity of solar radiation to cloudiness will be estimated using an effective cloudiness measure which is estimated from terrestrial radiation measurements. Thus β will be used to estimate the sensitivity of terrestrial radiation to cloudiness, and either C_1 or C_2 can be used to estimate the sensitivity of solar radiation to cloudiness. Since C_1 better maps the variability of cloud cover than C_2, it is used in the sensitivity analysis for solar radiation. The sensitivity of net radiation to cloudiness is the sum of the corresponding sensitivities of net solar radiation and net terrestrial radiation.

In order to make an accurate assessment of the changes in the surface energy balance due to a given change in cloudiness, these sensitivities must be expressed using the same units of cloudiness: units of C_1 do not directly translate into units of β. This is due in part to the higher variability of C_1 in comparison to β (a 1% change in C_1 corresponds to a much smaller change in effective cloudiness than a 1% change in β). Plots of C_1 and C_2 versus β are shown in Figure 6. Estimation of the slope from the plot of β versus C_1 reveals that a 3.3% change in C_1 is equivalent to a 1% change in β. Dividing the sensitivity of terrestrial radiation to a 1% change in β by 3.3 will then convert it to the same units as the sensitivity of solar radiation to C_1. The final results for all sensitivities are tabulated in Table II. A 1% decrease in cloudiness results in an increase of net solar radiation of *ca.* 1.6 W/m^2, and a decrease in net terrestrial radiation of *ca.* 0.7 W/m^2. Hence, in conclusion a 1% decrease in cloudiness should result in an increase of net radiation of *ca.* 1 W/m^2.

6. DEFORESTATION AND THE CLOUDINESS FEEDBACK

Changes in the surface radiation balance, following large scale deforestation, can be estimated by quantifying changes in the properties of the land surface and the atmosphere. The major factors affecting

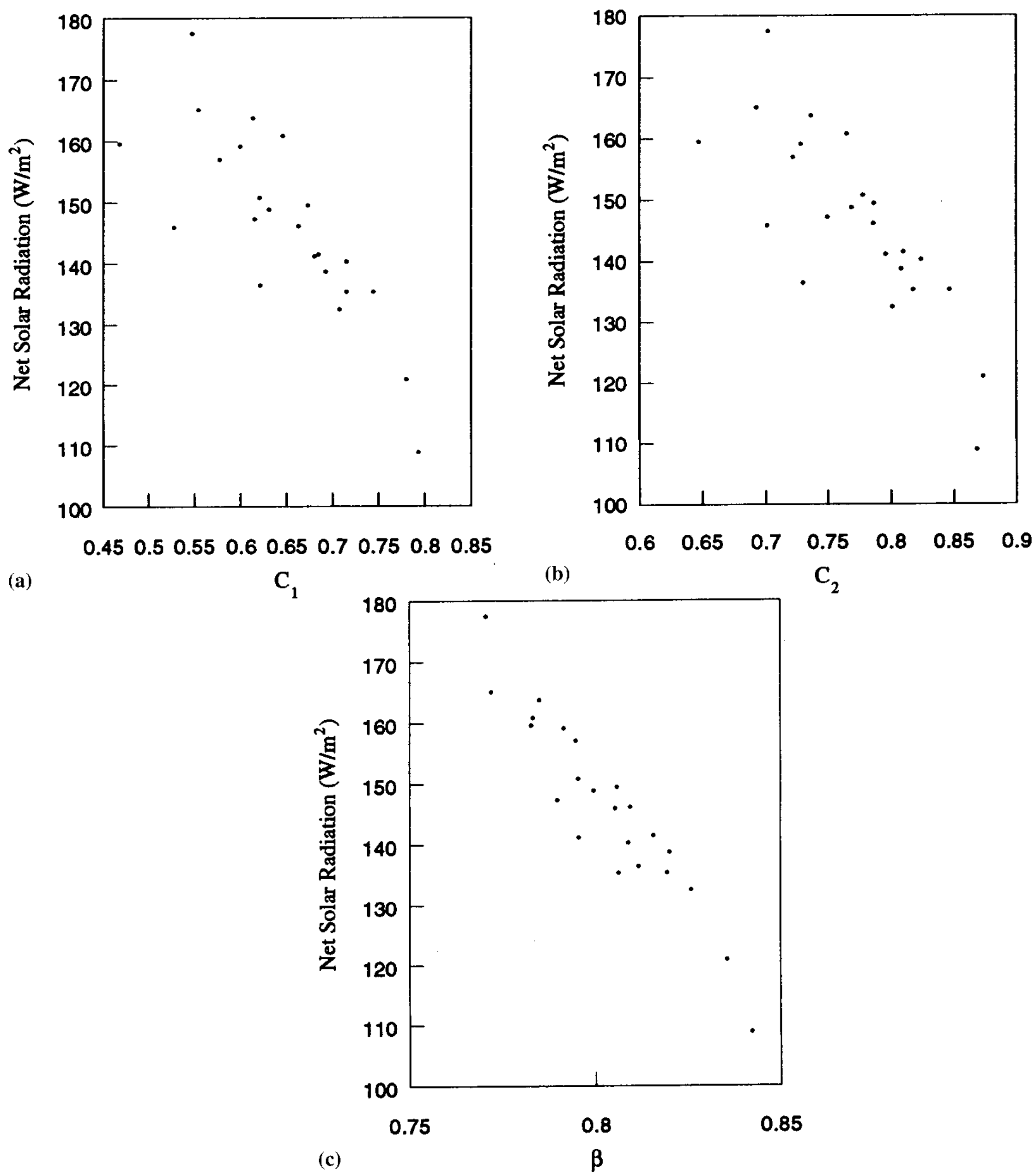


Figure 3. Monthly average of a cloudiness measure, (a) C_1; (b) C_2; and (c) β, plotted against monthly averaged net solar radiation

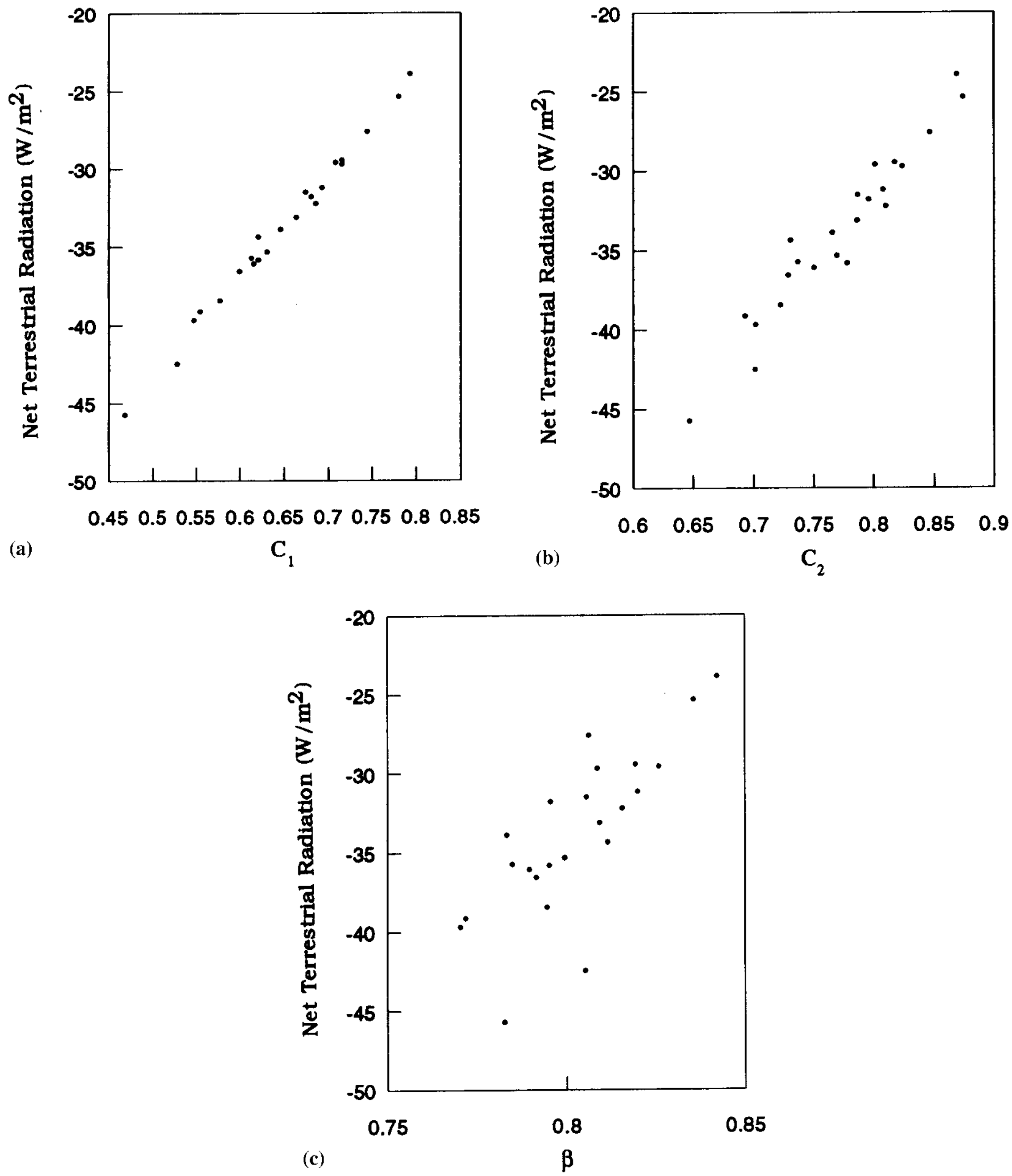


Figure 4. Monthly average of a cloudiness measure, (a) C_1; (b) C_2; and (c) β plotted against monthly averaged net terrestrial radiation

the radiation balance are effective cloudiness and land surface albedo for solar radiation, and effective cloudiness and surface temperature for terrestrial radiation. Figures 7–9 were developed to describe the changes in the surface radiation components that correspond to a wide range of potential changes in the relevant land surface and atmospheric properties. The sensitivity of net solar radiation is computed using:

$$\delta N_S = \frac{\partial N_S}{\partial A}\,\delta A + \frac{\partial N_S}{\partial C}\,\delta C \tag{15}$$

where N_S is net solar radiation, A is albedo, and C is cloudiness. $\partial N_s/\partial C_1$ is used to estimate the sensitivity of net solar radiation to cloudiness. Likewise, the sensitivity of net terrestrial radiation is estimated by

$$\delta N_t = \varepsilon\sigma(T_{a_f}^4 - T_{a_o}^4) + \frac{\partial N_t}{\partial C}\delta C \tag{16}$$

where N_t is net terrestrial radiation, ε is emissivity, and σ is the Stefan-Boltzmann constant, T_{a_f} is the surface temperature after land surface change, T_{a_o} is the original land surface temperature (taken to be 24.8°C in agreement with the observations of Bastable *et al.* (1993). $\partial N_t/\partial\beta$ is used to estimate the sensitivity of net terrestrial radiation to cloudiness after converting it to per units of C_1. The change in atmospheric water vapor following large scale deforestation is an important process that has a significant impact on net terrestrial radiation at the surface. This water vapor feedback has not been considered explicitly in this analysis. However, it is likely to have the same sign as the cloudiness feedback.

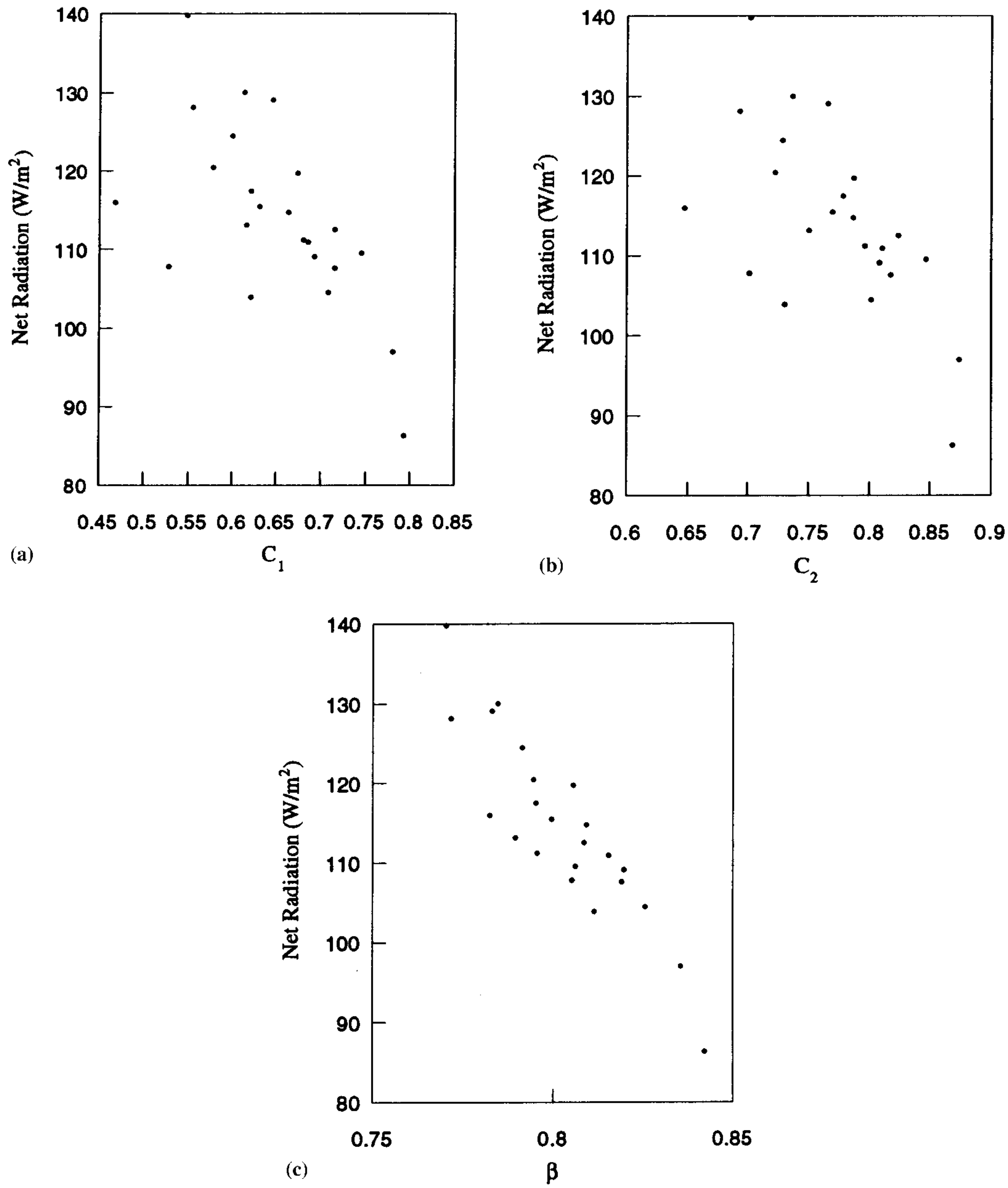
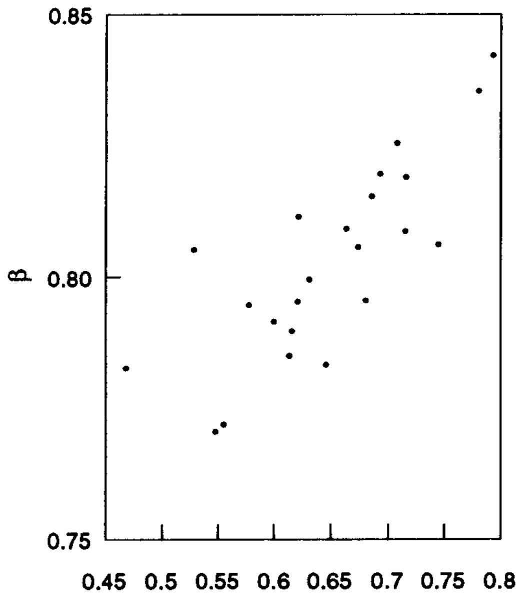
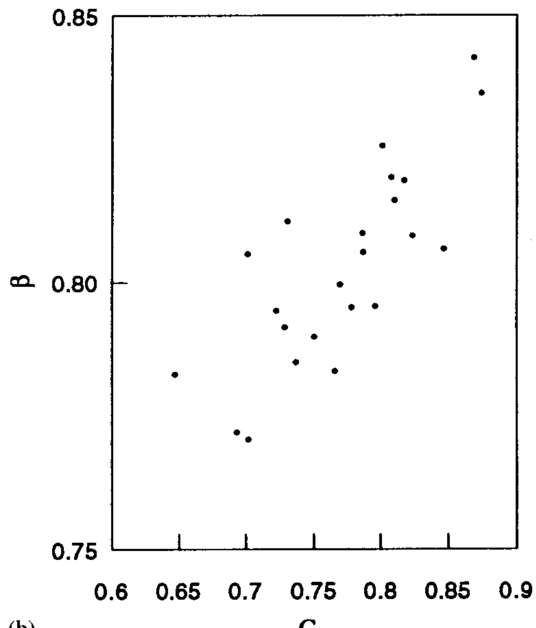


Figure 5. Monthly average of a cloudiness measure, (a) C_1; (b) C_2; and (c) β, plotted against monthly averaged net radiation



(a)



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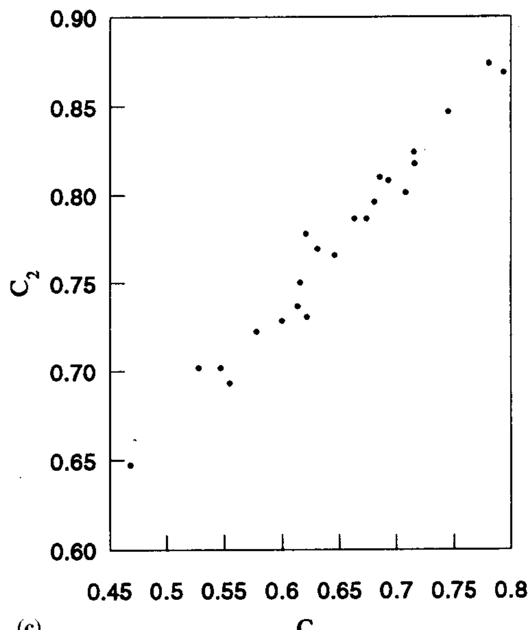

(c)

Figure 6. Comparisons of the monthly averages of the cloudiness measures

For specified changes in the surface albedo, surface temperature, and effective cloudiness, the resulting changes in net solar radiation, net terrestrial radiation, and net radiation are described in Figures 7–9, respectively.

Observed and simulated changes of the relevant meteorological variables following deforestation generally fall in the range of an increase in albedo of *ca.* 4–5% (Bastable *et al.*, 1993; Culf *et al.*, 1995, 1996), an increase in surface temperature of *ca.* 1–3°C, a decrease in precipitation of *ca.* 10–30% (Nobre *et al.*, 1991; Dickinson and Kennedy, 1992; Lean and Rowntree, 1993), and a decrease in cloudiness of *ca.* 4–7% (Dickinson and Kennedy, 1992; Lean and Rowntree, 1993). Examination of Figures 7–9 show that for an increase in surface albedo of *ca.* 5%, net solar radiation is expected to decrease by *ca.* 8 W/m^2;

Table II. Sensitivity of net radiation, net terrestrial radiation, and net solar radiation to cloudiness

$\partial N_S/\partial C$	-1.58	W/m^2/ 1% cloudiness
$\partial N_t/\partial C$	$+0.67$	W/m^2/ 1% cloudiness
$\partial R_n/\partial C$	-0.91	W/m^2/ 1% cloudiness

N_S, net solar radiation.
N_t, net terrestrial radiation.
R_n, net radiation.
C, cloudiness, converted to C_1 units.

a significant change. But when this change is accompanied by a 5% decrease in effective cloudiness, the net effect is almost 0 W/m^2, which demonstrates the significance of the cloudiness feedback. For terrestrial radiation, a temperature increase of 2°C results in a decrease of net terrestrial radiation of *ca*. 12 W/m^2; with a 5% decrease in effective cloudiness superimposed, this effect is amplified to a decrease of *ca*. 15 W/m^2. Without considering the cloudiness feedback, net radiation would decrease by 20 W/m^2; when the effect of a 5% change in effective cloudiness is considered, the decrease in net radiation is reduced to *ca*. 15 W/m^2, a difference of 5 W/m^2. The total change in net radiation due to deforestation represents *ca*. 13% of the observed net radiation (see Table I).

A comparison of GCM simulations results provides further evidence for the importance of the cloudiness feedback in the surface energy balance of the Amazon forest. As shown in Table III, the simulations of Dickinson and Kennedy (1992) and Lean and Rowntree (1993), which account for the cloudiness feedback, show a small decrease in net solar radiation, a relatively large decrease in net terrestrial radiation, and a moderate decrease in net radiation. On the other hand, the earlier study of Nobre *et al*. (1991), which did not account for the cloudiness feedback, gives a much higher estimate for the decrease in net solar radiation but a significantly smaller decrease in net terrestrial radiation. By comparing the results of these simulations, it is concluded that the cloudiness feedback is an important process in the surface radiation balance over the Amazon forest. The results of these simulations are consistent with the empirical analysis of this paper.

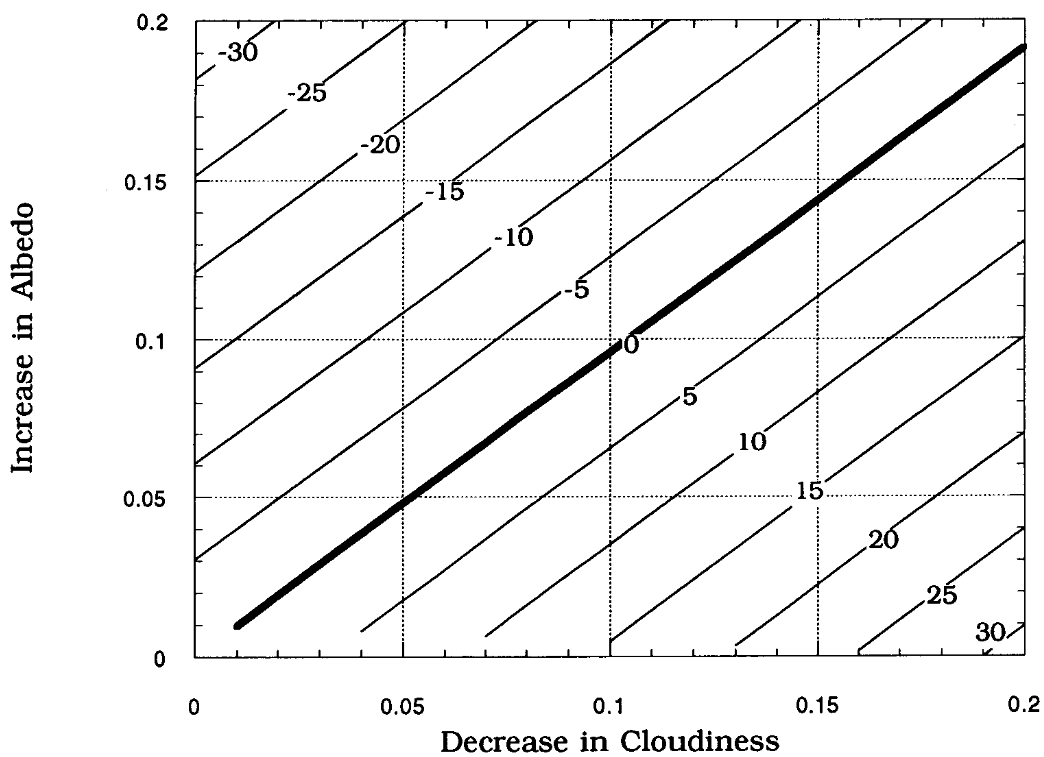

Figure 7. Changes in net solar radiation for varying degrees of cloudiness and albedo change

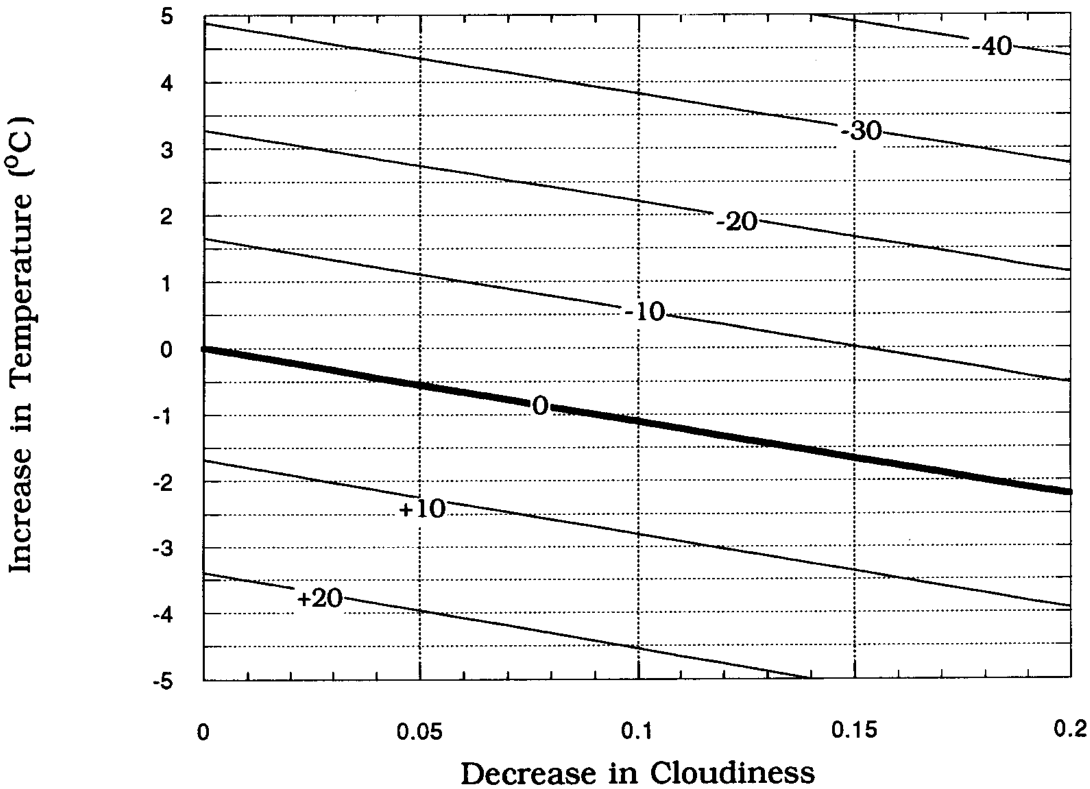

Figure 8. Changes in net terrestrial radiation for varying degrees of cloudiness and temperature change

7. DISCUSSION AND CONCLUSIONS

The classical definition of cloudiness as fractional cloud cover does not describe sufficiently the overall effect of clouds on the surface energy balance. The height, depth, and density of clouds all affect the radiation fields; hence the concept of effective cloudiness has been introduced and defined as the conglomerate effect of the atmosphere on the surface radiation field. This is a broad definition which can be refined into specific aspects of cloudiness with further study; such endeavors would require extensive study of the details of cloud observations. However, the concept of effective cloudiness better describes the role of clouds in modifying the surface energy balance than merely fractional cloud cover.

The sensitivity of the surface fluxes of solar and terrestrial radiation to natural variability in cloudiness is investigated to infer the potential role of the cloudiness feedback in the surface energy balance. The results of this analysis indicate that net solar radiation is likely to decrease due to any increase in cloudiness; a 1% increase in cloudiness induces a decrease in net solar radiation of *ca.* 1.6 W/m^2. On the other hand, net terrestrial radiation is likely to increase due to any increase in cloudiness; a 1% increase in cloudiness induces an increase in net terrestrial radiation of *ca.* 0.7 W/m^2. Hence, the net effect of clouds is to reduce net radiation; a 1% increase in cloudiness induces a reduction in net radiation of *ca.* 1 W/m^2.

The cloudiness feedback following deforestation has a different sign and magnitude for solar and terrestrial radiation. Field observations indicate that deforestation increases the surface albedo significantly; this on its own would reduce the net solar radiation, especially for small clearing areas over which effective cloudiness remains unchanged or may even increase. However, as deforestation spreads over larger areas, effective cloudiness is expected to decrease, allowing more solar radiation to reach the land surface and acting as a negative feedback. Using observations of the changes in surface albedo and the

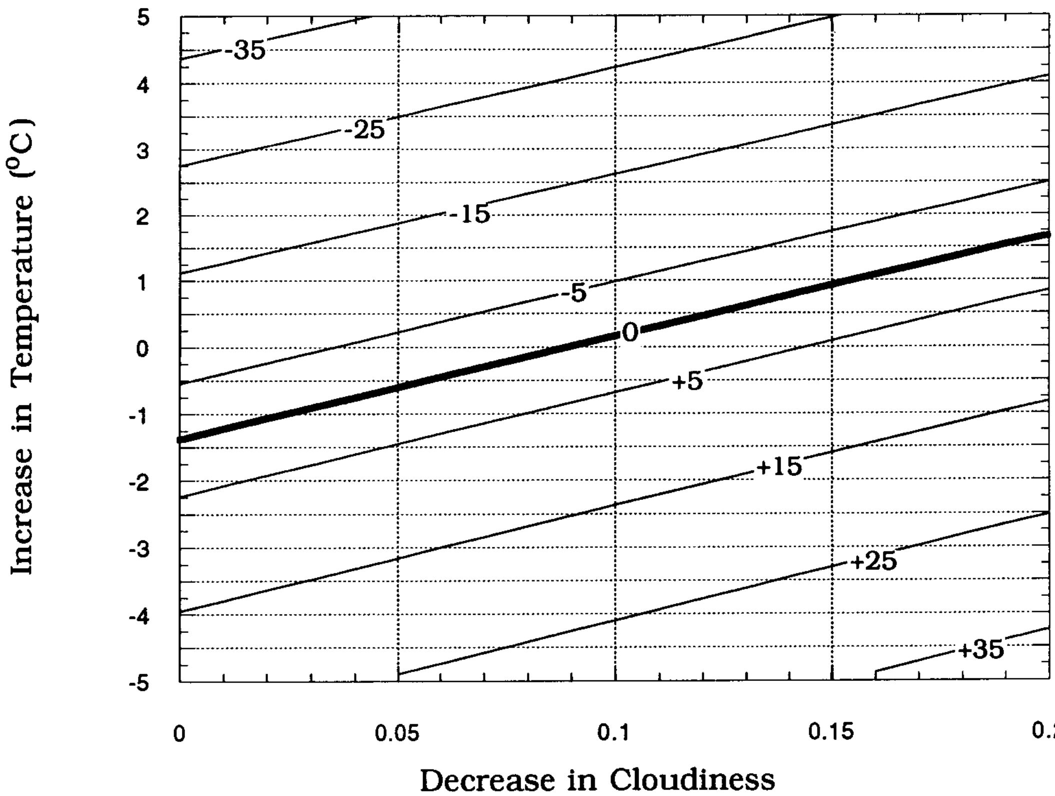

Figure 9. Changes in net radiation for varying degrees of cloudiness and temperature change. Change in albedo is 5%

current GCM predictions of cloudiness changes following deforestation, it is here estimated that the magnitude of the decrease in net solar radiation due to the observed increase in surface albedo, and the magnitude of the increase in net solar radiation due the cloudiness feedback, are almost equal. Thus, the impacts of the two processes would tend to cancel each other and the overall changes in net solar radiation are likely to be small.

However, the corresponding change in net terrestrial radiation is likely to be enhanced due to the cloudiness feedback. The change in cloudiness following large scale deforestation introduces a positive feedback that enhances the initial reduction in net terrestrial radiation due to the increase in surface temperature and the reduction in atmospheric water vapor. The magnitude of the negative cloudiness feedback on net solar radiation is likely to be larger than the magnitude of the positive cloudiness feedback on net terrestrial radiation. As a result, the overall effect of the change in cloudiness on net radiation is likely to introduce a negative feedback, though somewhat weakened by the positive feedback

Table III. Changes in net radiation, net solar radiation, and net terrestrial radiation based on numerical simulations

Study	Scale of deforestation (km^2)	Change in net surface radiation (W/m^2)	Change in net solar radiation (W/m^2)	Change in net terrestrial radiation (W/m^2)
Numerical simulations				
Nobre *et al.* (1991)	10^6	−21	−18	−8
Dickinson and Kennedy (1992)	10^6	−18	−3	−15
Lean and Rowntree (1993)	10^6	−19	−4	−15

of cloudiness on terrestrial radiation. GCM simulations which ignore the cloudiness feedback may have a small error in evaluating the impact on net radiation (see Table III). However, the individual components of net radiation, especially net solar radiation, simulated by such models, may be in considerable error.

The conclusion that clouds exert a negative feedback on net radiation over tropical rainforests has implications that go beyond the response of the tropical atmosphere to large scale deforestation. In particular, this negative feedback is likely to play an important role in large-scale land–atmosphere interactions over rainforests. One direct implication concerns the large scale atmospheric circulations that induce vertical motion and enhance cloud formation over rainforests and continental regions. The negative cloudiness feedback that is associated with these circulations is likely to reduce net radiation. The results of the ARME experiment, (Shuttleworth, 1988), suggest that net radiation may be the limiting factor for evaporation from the rainforest. Further, Eltahir (1996) argues that the magnitude of net radiation determines the total flux of heat from the surface, and as a result, net radiation partly controls the moist static energy in the boundary layer. Recent theories of large scale atmospheric circulations in the tropics (Emanuel *et al*., 1994; Eltahir, 1996) suggest that circulations over rainforests are driven by the gradients of boundary layer moist static energy. Hence, by putting all these pieces together it is here hypothesized that clouds introduce a negative feedback that impacts not only the surface radiation field but goes beyond that to potentially regulate evaporation, moist static energy, and the large-scale atmospheric circulations over tropical land regions. Some of the future research intended by the same authors will focus on further development and testing of this hypothesis.

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