POSSIBLE CLIMATIC IMPACTS OF LAND COVER TRANSFORMATIONS, WITH PARTICULAR EMPHASIS ON TROPICAL DEFORESTATION

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Abstract. The climatic impact of albedo changes associated with land-surface alterations has been examined. The total surface global albedo change resulting from major land-cover transformations (i.e. deforestation, desertification, irrigation, dam-building, urbanization) has been recalculated, modifying the estimates of Sagan *et al.*, (1979). Tropical deforestation (11.1 million ha yr^{-1}, or 0.6% yr^{-1}, Lanly, 1982) ranks as a major cause of albedo change, although uncertainties in the areal extent of desertification could conceivably render this latter process of similar significance. The maximum total global albedo change over the last 30 yr for the various processes lies between 0.000 33 and 0.000 64, corresponding to a global temperature decrease of between 0.06 K and 0.09 K (scaled from the 1-D radiative convective model of Hansen *et al.*, 1981), which falls well below the interannual and longer period variability.

An upper bound to the impact of tropical deforestation was obtained by concentrating all vegetation change into a single region. The magnitude of this modification is equivalent to 35–50 yr of global deforestation at the current rate, but centered on the Brazilian Amazon. The climatic consequences of such tropical deforestation were simulated, using the GISS GCM (Hansen *et al.*, 1983). In the simulation, a total area of 4.94×10^6 km^2 of tropical moist forest was removed and replaced by a grass/crop cover. Although surface albedo increased from 0.11 to 0.19, the effect upon surface temperature was negligible. However, other climate parameters were altered. Rainfall decreased by 0.5 -0.7 mm day^{-1} and both evapotranspiration and total cloud cover were reduced. The absence of a temperature decrease in spite of the increased surface albedo arises because the reduction in evapotranspiration has offset the effects of radiative cooling. The decrease in cloud cover also counteracts the increase in surface albedo. These locally significant changes had no major impact on regional (Hadley or Walker cells) or the global circulation patterns.

We conclude that the albedo changes induced by current levels of tropical deforestation appear to have a negligibly small effect on the global climate.

1. Introduction

Land cover transformations (both natural and man-made) may affect the climate in a number of ways. The most readily observable change following a land cover alteration is the change in the surface (and often the planetary) albedo, which has been investigated as a potential cause of climatic variation (Charney, 1975; Charney *et al.*, 1977; Sagan *et al.*, 1979; Potter *et al.*, 1975, 1981). We attempt here to reassess some of the reported findings.

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Tropical deforestation has been recognized as one of the two most significant current land cover changes, the other being desertification (Bolin, 1977; Woodwell *et al.*, 1978; Hampicke, 1980; Otterman, 1974, 1977a). In this paper, we investigate those possible climatic effects of tropical deforestation, which result *only* from surface albedo changes – i.e. we do not attempt to include the effects upon the CO_2 budget. We try to estimate the *maximum* impact likely to occur as a result of tropical deforestation by maximizing the changes important to the climate. Our investigation is divided into two parts:

(1) An assessment of the probable alteration in the global climate, calculated using simple one-dimensional climate models. We undertake this part of our examination not necessarily because we believe the models to be accurate, but because they have served as the basis for other studies and therefore, if our conclusions differ, it is important to re-examine the approach they used. In this part of our analysis, we include the effects of surface albedo changes caused by desertification, because these form part of the earlier surveys with which our results disagree.

(2) We examine the probable regional and global climatic effects of tropical deforestation alone, by using a new general circulation climate model (GCM) (Hansen *et al.*, 1983). Here we isolate the total impact of deforestation in one area (tropical Amazonia) and again seek to maximize the impact. In this case, the additional effects of desertification are not considered. Deforestation is simulated by altering the tropical forest to grass/savanna cover in the climate model.

2. Tropical Deforestation

It is widely recognized that destruction of the tropical rain forests is environmentally undesirable. Biologists have called for preservation of substantial areas of this complex biome, because of its unique species diversity, potential source of valuable commercial products including lumber, rubber, pharmaceuticals, etc. and because of a lack of viability of alternate forms of land-use (Gomez-Pompa *et al.*, 1972; Richards, 1973; Eckholm, 1979; Myers, 1980a). Tropical rain forests are biologically diverse, multi-layered, predominantly evergreen forests, with little to no seasonality, heavy rainfall ($\approx$200–300 mm month^{-1}) and relatively constant average temperatures (around 25 °C). The dark, dense, moist vegetation gives these forests a lower visible (385–760 nm) surface albedo than most natural or man-made terrains. In addition to low visible albedo, tropical rain forests also exhibit low infrared emissivity. That is, while most of the incident shortwave energy is absorbed, the emission of longwave or thermal energy is inhibited, yielding a positive net radiation balance. This energy is utilized in photosynthesis and in evapotranspiration of water. The turbulence caused by the stand itself accelerates the exchange of sensible heat so efficiently, in spite of the positive radiation balance, that forest canopies are cool compared with cultivated or open land in the summertime. The fixation of solar energy into wood is, of course, an important component of the carbon cycle and hence influences the level of atmospheric CO_2. While tropical deforestation is widely believed to contribute to the observed increase in atmospheric CO_2 (Bolin, 1977; Woodwell *et al.*, 1978; Hampicke, 1980), the extent and timing of the CO_2 release depends

on the proportion of cleared forests which are immediately burned to those converted to permanent structures. Not all burned wood is released as CO_2 – a substantial fraction remains on the ground as charcoal (Seiler and Crutzen, 1980). On the other hand, 'permanent' wooden structures are eventually torn down and burned, thus returning CO_2 to the atmosphere.

Removal of the dense surface vegetation in tropical forests will cause massive increases in runoff and soil erosion leading to increased flooding hazards and river sediment transport (Delwaulle, 1973; Eckholm, 1979). Deforestation is particularly detrimental in the tropical environment, since most of the nutrients are concentrated in the above-ground biomass and therefore removal of the vegetation leads to rapid decrease in soil fertility (Richards, 1973; Jordan, 1982).

The claim is often advanced that the removal of tropical rain forests will substantially alter the climate, either by adding CO_2 to the atmosphere, thereby enhancing the greenhouse effect, or by increasing the global surface albedo (Bolin, 1977; Woodwell *et al.*, 1978; Sagan *et al.*, 1979; Hampicke, 1980; Potter *et al.*, 1981; Shukla and Mintz, 1982). Such a claim merits careful investigation. In this paper, we examine the climatic impact of albedo changes associated primarily with tropical deforestation. Other major land transformations which may affect albedo and hence climate, are also briefly reviewed. An upper limit to possible climatic effects of tropical deforestation is obtained by assuming large-scale clearing of the Brazilian Amazon. The magnitude of this surface modification is equivalent to 35–50 yr of deforestation at the current global rate, but concentrated in the Brazilian Amazon.

3. Changes in Tropical Forest Area and Associated Albedo Changes

Before the possible impact of tropical forest removal on surface albedo can be assessed, it is necessary to establish the extent of deforestation. Global forest inventories which span a period of only 60 yr do not reveal a reduction in the area of the world's forests. Zon and Sparhawk (1923) provide one of the earliest global forest surveys. The FAO World Forest Inventories (1948, 1953, 1958, 1963) attempted to compile and standardize global forestry data, yet initially coverage was incomplete for many developing countries and a uniform definition of forest was lacking (different countries may classify stands of quite different tree density as 'forest'). Revisions in definitions from one survey to the next limit the temporal comparability of the WFI.

The reported global total forest area has remained close to 4 billion ha over the last 30 yr, partly because of incomplete coverage in older surveys, resulting in underestimates and partly because tropical removal has been compensated to some extent by temperate reforestation (Armentano and Ralston, 1980; European Timber Trends and Prospects 1950–2000). The large uncertainties in the forestry data conceal any trend over time although a decrease in area is widely accepted.

Recent estimates for total tropical forest areas range between 935 million to 1970 million ha (Table I). However, the totals fall into two discrete groups, averaging around 1.1 and 1.9 billion ha respectively, which correspond to closed forests and total forest

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TABLE I: Tropical Forests, 1000 ha.

Region	Myers, 1980b	Summer, 1976		Persson, 1974*		FAO Prod. YB. 1980* (data for 1979)	Lanly, 1982	
	Tropical moist forest	Moist forest	Total forest	Closed	Total forest		Closed forests	All forest land
Asia and Pacific	271 396	254 000	417 000	294 005	405 983	357 054	305 510	336 458
Latin America	641 580	506 000	964 000	576 822	734 065	943 057	678 655	895 652
Africa	151 390	175 000	334 000	195 947	754 960	669 213	216 634	703 079
Total	1 064 366	935 000	1 715 000	1 066 774	1 895 008	1 969 324	1 200 799	1 935 189

Mean (total forest) $1879 \pm 113 \times 10^6$ ha.
Mean (closed forest) $1067 \pm 133 \times 10^6$ ha.
* Data for same 16 countries (Asia), 23 countries (Latin America) and 37 countries (Africa) as in Lanly, (1982).

land, including open woodland, and often, savanna (Persson, 1974; Sommer, 1976; Myers, 1980a, b; Lanly, 1982).

Persson (1974) defined 'total forest and other wooded land' as areas covered by trees and shrubs, and also including open woodland and savanna (trees cover 5–20% area), but excluding succulents, scrub, grassland, and city parks, orchards, or non-forestry tree crops (e.g. rubber), wooded pasture or rangeland. In addition, Persson defined 'closed forest' as land with trees whose crowns cover over 20% of the area, including large windbreaks or shelter belts, and temporary openings caused by cutting, burning or shifting cultivation, assuming that the land will be ultimately reforested. It excludes areas under 0.5 ha or permanently cleared land. Persson's 'total forest and wooded areas' are comparable to 'forest land' in the 1979 and 1980 FAO Production Yearbooks.

Tropical forests are being cleared in a number of different ways. The traditional shifting cultivation is being overtaken by whole-scale removal of vast tracts to make way for highways, cattle ranches and farming colonization. Estimates for losses of tropical forest land due to shifting cultivation and new permanent agriculture range between 10–20 million ha yr^{-1} (Myers, 1980a, b) and 3.5–8.3 million ha yr^{-1} (Seiler and Crutzen, 1980). Deforestation due to forest clearing for cattle grazing has been estimated at 1.35–2.0 million ha yr^{-1} (Myers, 1980a) and 4.8–6.0 million ha yr^{-1} (Seiler and Crutzen, 1980). Logging could account for 1.77–2.9 million ha yr^{-1}, and the demand for fuelwood could consume up to 2.5 million ha yr^{-1} (Myers, 1980a). Taken together, these various processes could remove up to 9–15 million ha yr^{-1} (Seiler and Crutzen, 1980), or even 24.5 million ha yr^{-1} of tropical moist forest (Myers, 1980a). Lanly (1982) finds a deforestation rate of around 7.3 million ha yr^{-1} of all closed forests and 3.8 million ha yr^{-1} of open forest, or a total of 11.1 million ha yr^{-1} (0.6% yr^{-1}), falling within the range of Seiler and Crutzen (1980). Myers' higher figure applies to the total area of degraded forests, and therefore covers a larger area than Lanly, who considers deforestation in a more restricted sense of a complete (and permanent) removal of the tree cover. Degradation refers only to partial destruction of forests or selective tree removals. Degraded forests are more difficult to assess than completely cleared forests, since not only are there uncertainties in the percentage of area that will revert back to secondary forest, but also in the percentage of the remaining tree cover. The albedo contrast associated with degradation is probably much less than that of complete clearing. Therefore we will examine the case of complete clearing, using Lanly's figures, which while they may under-estimate the total areal extent of forest disturbance, will nevertheless maximize albedo contrast.

Although the overall rate of clearing of all tropical forests is around 0.6% yr^{-1}, this rate varies widely from country to country, and even between regions within a country (Figure 1). Furthermore, the history of deforestation is variable. In many developing countries, deforestation increases rapidly as new roads penetrate previously inaccessible areas and loggers and farmers move in. In heavily exploited areas, like the Ivory Coast and Thailand, present high rates of clearing will probably deplete the forest resource within the next 20–30 yr, but presently remote or unproductive areas like the interior of the Amazon or the Congo (Zaire) basins can expect to undergo much higher rates of defore-

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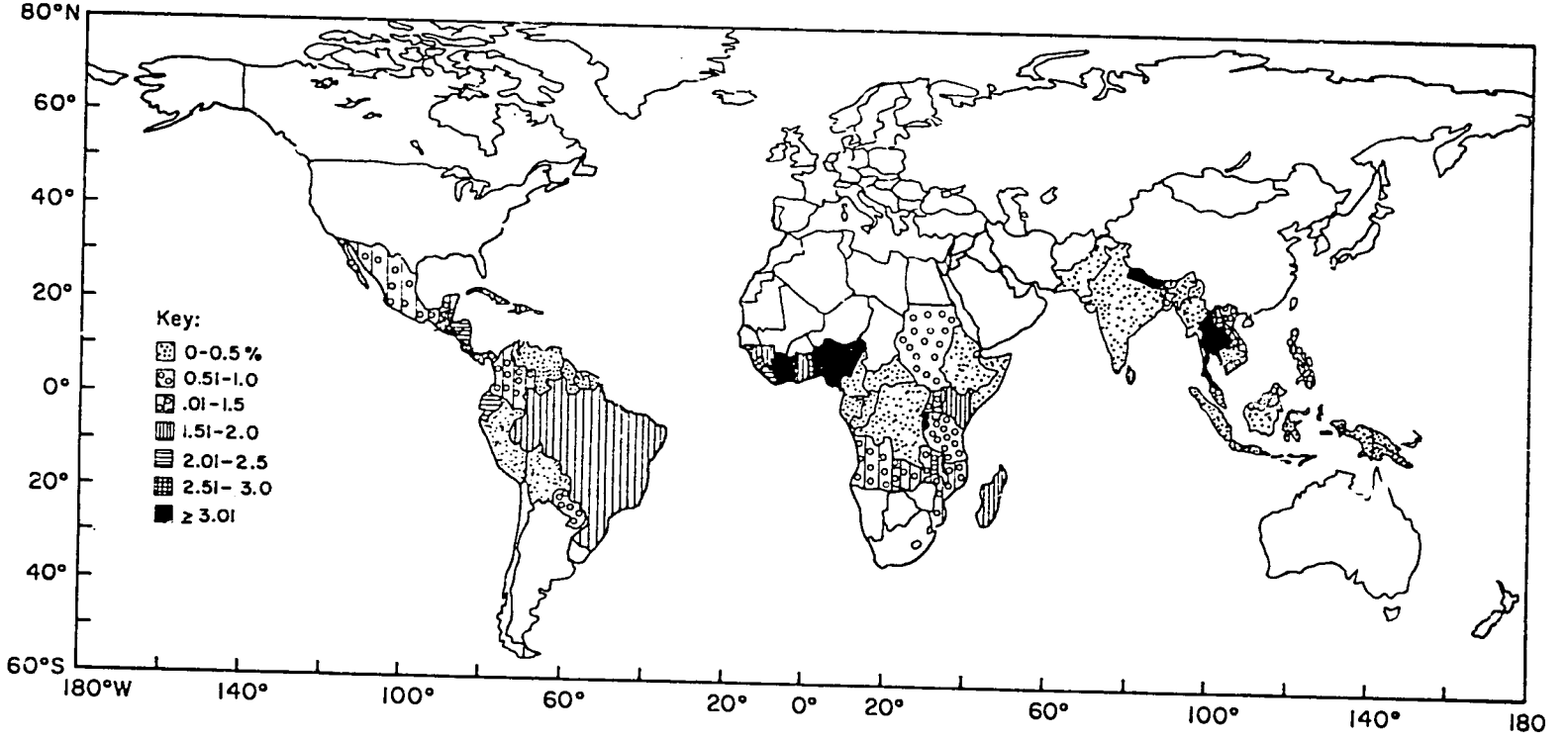

Fig. 1. Deforestation rates: % annual decrease in closed tropical forests: after Lanly and Rao, 1981; Lanly and Clement, 1981, Lanly, 1982.

station in the future. The very non-uniform spatial and temporal pattern of deforestation and lack of reliable data make it difficult to estimate changes in area and in rates. The estimates of Lanly (1982) and Seiler and Crutzen (1980) cover a span of 10–15 yr. These rates are probably higher than those of 20–30 yr ago. In the absence of reliable deforestation data for 30 yr ago, one could tentatively assume a roughly linear increase, so that given a deforestation rate of around 11 million ha yr^{-1}, approximately 330 million ha could have been eliminated over a 30 yr period. Clearing of mangrove swamps and other periodically inundated woody vegetation is included in this estimate.

Deforestation of temperate forests in North America and Europe has slowed down, if not halted altogether, and therefore change in their areal extent can be assumed to be negligible for this calculation (Armentano and Ralston, 1980; European Timber Trends and Prospects, 1950–2000).

Clearance of tropical rain forest increases the surface albedo from around 0.13 to around 0.16 for adjacent clearings and derived savanna (Oguntoyinbo, 1970; Pinker *et al.*, 1980). Cultivated crops in tropical environments have a mean albedo of around 0.17 (Oguntoyinbo, 1970). The sudanian and sahelian savannas have mean albedoes between 0.19–0.21, but these grow in semi-arid zones, north of the tropical forest zone, in Africa. Thus the average albedo change due to deforestation is probably considerably less than that assumed by Sagan *et al.* (1979) and Potter *et al.* (1975, 1981) and may not exceed 0.04 (Dickinson, 1980). The associated regional or global climate response may therefore have been overestimated. Here we consider two extreme cases of surface albedo change following deforestation: (1) maximizing the change in surface albedo, r_s (Table II) by taking a lower limit on the tropical forest albedo of 0.11 (based on Table 6 in Hansen *et al.*, 1983) and an upper limit to the new albedo represented by savanna grassland and (2) minimizing the change in r_s (Table II) by using albedo values from Pinker *et al.* (1980) of 0.13 changing to 0.16.

Table II has been revised from Sagan *et al.* (1979), using our estimates of areal modifications in vegetation cover and associated albedo changes (where not specifically referenced, taken from Hansen *et al.*, 1983).

The global albedo change, for each cause i is $\Delta R_i = A_i\, I_i [F_i\, \Delta r_t + (1 - F_i)\, \Delta r_s]$ (Sagan *et al.*, 1979) where $r_t = r_c + (1 - r_c)^2\, r_s/(1 - r_s\, r_c)$, and $r_c = 0.48$ (cloud albedo), F = cloud cover fraction, I = an insolation factor which includes the effects of latitudinal and seasonal variation in insolation, A = area changed, Δr_s = change in surface albedo. The calculated change in global planetary albedo is critically dependent upon both the *rate of vegetation change and the range* of albedo change. The effects of tropical deforestation could be very significant although other types of land transformation may also have a major impact on the earth's albedo. These are briefly discussed in the following paragraphs.

Irrigation for 21 arid countries has increased by 37 million ha over 30 years (Framji and Mahajan, 1969; FAO Production Yearbooks). Taking the average albedo of sandy desert soil as 0.35 and that of cultivated fields as 0.18 (mean value, from Budyko, 1958; Kondratyev, 1972; Kung *et al.*, 1964, and Oguntoyinbo, 1970), the corresponding albedo difference is –0.17 (Table II).

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TABLE II: Global albedo changes due to anthropogenic modifications.

Process	Land type change	Δr_s, surface albedo change (integrated)	A, change in area, over 30 yr rel. to earth's surface area	F, cloud cover fraction	I, insolation factor	ΔR, fractional change in global albedo over 30 yr
Deforestation of tropical forests	(1) Forest → savanna (Max.) grassland	(1) 0.11 → 0.20 = 0.09	0.006 47	0.5	1.2	(1) 4.59×10^{-4}
	(2) Forest → clearings (Min.) (Pinker *et al.*, 1980)	(2) 0.13 → 0.16 = 0.03				(2) 1.53×10^{-4}
Deforestation of temperate forests	Forest → field pasture	0.13 → 0.18 = 0.05	small	0.5	0.8	small
Dam-building	Field → water	0.18 → 0.04 = −0.14	0.000 59	0.25	1	-0.68×10^{-4}
Salinization	Field → saline field	0.18 → 0.24 = 0.06	0.000 074	0.25	1	0.04×10^{-4}
Irrigation of arid land	Desert soil → field	0.35 → 0.18 = −0.17	0.000 72	0	1	-1.23×10^{-4}
Urbanization	(Field, pasture → urban	0.18 → 0.13 = −0.05	0.000 59	0.5	1	-0.19×10^{-4}
	(Forest → urban	0.13 → 0.13 ≅ 0.0	0.000 59	0.5	1	
Desertification	(i) Scrub/shrub → desert soil	0.23 → 0.35 = 0.12	(a) 0.003 18	0	1	(a) 3.81×10^{-4}
	or		(b) 0.001 59	0	1	(b) 1.91×10^{-4}
	(ii) Protected exclosure → desert soil	0.39 → 0.44 = 0.05	(c) 0.003 18	0	1	(c) 1.59×10^{-4}
			(d) 0.001 59	0	1	(d) 0.79×10^{-4}

Total resulting planetary albedo change

	Case 1 Maximum tropical deforestation albedo change = 0.09	*Case 2* Minimum tropical deforestation albedo change = 0.03
For Desertification case (i) and (ii) above	(i) (a) 6.35×10^{-4}; (b) 4.45×10^{-4}	(i) (a) 3.29×10^{-4}; (b) 1.39×10^{-4}
	(ii) (c) 4.13×10^{-4}; (d) 3.33×10^{-4}	(ii) (c) 1.07×10^{-4}; (d) 0.27×10^{-4}

Note: Cases (a) and (c) use the UN estimate of the area of desertification whilst cases (b) and (d) use half this value.

Construction of dams has flooded 10.9 million ha, in the 41 largest artificial lakes of greater than 1000 km^2 area, since 1950 (Fels and Keller, 1973). However, Fels and Keller (1973) estimate that the total area of man-made lakes may exceed 30 million ha. (More recent compilations of major world dams give reservoir capacities but not surface area). Assuming that the reservoirs inundate mostly agricultural land, the associated albedo difference is about −0.14 (Table II).

Salinization of irrigated fields destroys up to 125 000 ha yr^{-1} (UN, 1977), or 3.75 million ha in 30 yr. Although crusts of pure halite (NaCl) can have an albedo as high as fresh snow (~0.60), an increase in salinity of only a few percent could ruin an irrigated crop. Thus the effect on albedo may be due more to the absence of vegetation (i.e. bare field) than to the presence of salt. Summer surface albedos for saline soil range between 0.17 and 0.29 (mean = 0.24; Kung *et al.*, 1964). The albedo change from cultivated field to saline soil is 0.18 to 0.24 → +0.06.

We have adopted the rate of urbanization (2×10^6 ha yr^{-1}) used by Sagan *et al.* (1979). The mean albedo for urban areas is close to 0.15 (Oke, 1974). However, more recent measurements suggest somewhat lower values (e.g. 0.13, Dabberdt and Davis, 1978). Thus, the albedo change from temperate deciduous and coniferous forests to cities is close to zero (Table II). (The areal conversion of tropical forests to cities is considered negligible). The conversion of pastures and fields to built-up areas decreases the albedo by 0.05. If half of the urbanization occurs at the expense of agricultural land, in temperate areas, this involves an area of 30 million ha over 30 yr.

The areal extent of permanent desertification is uncertain. The UN (1977) reports a global desertification rate close to 5.4 million ha yr^{-1}, representing a loss of 125 000 ha of irrigated land to salinization and/or waterlogging, and deterioration of 1.7 million ha of rainfed cropland, and 3.6 million ha of rangeland. Dregne (1977) estimates that 1.38 billion ha (28.4% arid land area) are severely affected by degradation of vegetation, sand drifting, salinization and spreading of undesirable forbs and shrubs. However, desertification may be reversible under favorable ecological conditions (Le Houerou, 1977), when local weather conditions ameliorate or when anthropogenic pressures are reduced (Otterman, 1977b; 1981).

In general, desertification increases surface albedo, except where exposed soil may be darker than the removed vegetation cover, or where replacement of nutritious grassland by less desirable shrubs may not affect albedo (Dregne, private communication 1983). A protected exclosure in the Sinai desert had an albedo of 0.39; the surrounding overgrazed terrain showed an albedo of 0.44 (Otterman, 1977b) This difference (0.44 to 0.39) may represent a lower limit of the albedo change associated with desertification, since the exclosure may not yet have reached its equilibrium value. An upper bound on albedo contrasts exists between the overgrazed Sinai (0.45) and the darkest part of the Negev (0.28), at the junction of the Gaza strip and the Negev-Sinai boundary (Otterman, 1981).

An extreme case of desertification is the conversion of semi-arid shrub or scrub into sandy desert soil, for which the albedo could increase by 0.12 (0.23 → 0.35), or even less likely up to 0.15, for transformation of grassland or savanna to desert. However, most desertification involves degradation of the vegetation cover, rather than burial

by drifting sand. If the protected Sinai exclosure (see above) is more representative of the albedo change associated with desertification, then an increase of +0.05 can be expected due to this cause. The albedo changes in Table II have been calculated based on both these albedo changes: (i) shrub/scrub (0.23) → sandy soil (0.35), +0.12 and (ii) protected exclosure (0.39) → sandier non-vegetated soil (0.44), +0.05. We have used the UN areal estimate of desertification (162 million ha in 30 yr), which probably represents an upper limit, in cases (a) and (c), and 1/2 that value, which we believe is more realistic since with improved rainfall and reduced grazing pressure, much of the desertified land may recover, in cases (b) and (d).

From Table II it may be seen that the total global albedo change over the last 30 yr for the various processes, based on best estimates, lies between 0.000 3 and 0.000 64 with the maximum values (case (1)) between 0.000 33 and 0.000 64, roughly half of Sagan *et al.*'s (1979) estimate of 0.001. The major differences from Sagan *et al.* (1979) are in the reduced albedo changes associated with tropical deforestation, more conservative estimates of permanently desertified land, and a lower albedo change due to this process, a significantly lower estimate of the effect of salinization, and inclusion of arid-land irrigation and dam-building, both of which tend to reduce albedo. From Table II it is clear that desertification could be comparable to tropical deforestation in the magnitude of the resultant albedo change, except under the more restrictive assumptions of desert areal change and albedo contrast. The assumptions regarding the extent of desertification are uncertain by a factor of 2 or 3 and the range of albedo alterations for both processes is very difficult to determine. Nevertheless, tropical deforestation is still one of the major causes of surface and planetary albedo change. In particular, for our case (1) in which the impact of tropical deforestation was maximized by using a very large albedo change, the total albedo change is only just over half that predicted by Sagan *et al.* (1979). Even so it may still be important to investigate its impact on climate. This will be the subject of the next section.

4. Estimating the Possible Climatic Impact of Tropical Deforestation by Use of Climate Models

The art of climate modelling depends upon the success of parameterizations employed in the construction of the climate model. It is necessary to parameterize any processes not modelled explicitly. In the simplest case, the null parameterization ignores the process in question. The next simplest method is climatological specification which can be annual, seasonal or by month. Beyond these methods, parameterizations generally attempt to portray physically realistic relationships. Even so, there is considerable dependence upon tuning parameters together, according to their present-day (i.e. observed) relationships. Because of the large amount of computer time involved in numerical modelling, any increase in either the number of parameters explicitly modelled or the physical realism with which they are treated *must* be traded off against spatial and/or temporal resolution. The interrelationships between the primary processes included in climate models and the possible level of resolution are discussed at length by Shine and Henderson-Sellers (1983) and need not be repeated here.

We estimate the possible climatic impact of tropical deforestation in two ways. Firstly, we re-examine the claims made by Sagan *et al.* (1979) by using results from two different 1-D radiative-convective (1-D RC) climate models. Secondly we report upon a climatic simulation of tropical deforestation using the GISS general circulation climate model (Hansen *et al.*, 1983). We selected this model *because* its coarse spatial resolution permits inclusion of a number of parameters, such as ocean transport and sea ice, considered vital for medium-term (i.e. longer than 1–2 yr) simulation of the climate system.

4.1. *Simple Climate Models*

Sagan *et al.* (1979) use results from the 1-D RC of Manabe and Wetherald (1967) to estimate a 2 K temperature decrease caused by a planetary albedo change of 0.01. They further suggest that anthropogenic changes over the last 25 yr have led to a global temperature decrease of around 0.2 K. Our maximized calculated planetary albedo increase of between 0.000 33 and 0.000 64 will give rise to a much smaller temperature decrease of the order of 0.06 K to 0.13 K. This temperature alteration is probably too small to be detected above the interannual and longer period variability (Hansen *et al.*, 1981). The calculation of the global albedo from the likely surface albedo changes also depends upon the appropriateness of the assumptions made about and values assigned to all of the other climatic variables. For example cloud changes are not included in the calculation of Sagan *et al.* (1979).

Hansen *et al.* (1981) in their 1-D RC climate model find a temperature decrease of 1.3 K for a surface albedo change of +0.015. Within the range zero to +0.015 in surface albedo a linear relationship between temperature and surface albedo can be used.* Sagan *et al.* (1979) used *planetary* rather than *surface* albedo alterations in their calculations. The surface albedo changes associated with the land cover changes listed in Table II can be calculated by multiplying the area, A, by the surface albedo change Δr_s. These are listed in Table III. If we choose to consider the *maximum* likely effect of tropical deforestation upon the surface albedo, the range of albedo increase is from 0.001 to 0.0007. These albedo changes are likely to give rise to surface temperature decreases of the order of 0.06 K to 0.09 K. These temperature changes, which except in the latter case, (a), are forced almost entirely by the alterations in tropical forest areas, are very small. We suggest that within the error ranges of global 1-D RC climate models, the climatic impact of deforestation on global mean temperatures over the last 30 yr is close to zero.

4.2. *A General Circulation Model Study*

The estimate of the possible impact of tropical deforestation, using simple climate models, suggests that the claims of substantial global climatic modification caused by albedo changes may have been premature. However these calculations considered the alteration

* This linear interpolation of the surface albedo effect has been shown to be valid in other work (e.g. Rossow *et al.*, 1982).

TABLE III: Surface albedo change over the last 30 yr (product of columns Δr_s and A of Table II)

Process		Surface albedo change		
Deforestation of tropical forests			(1)	5.82×10^{-4} (maximum)
			(2)	1.94×10^{-4} (minimum)
Deforestation of temperate forests				small
Dam-building				-0.83×10^{-4}
Salinization				0.04×10^{-4}
Irrigation of arid land				-1.22×10^{-4}
Urbanization				-0.30×10^{-4}
Desertification			(a)	3.82×10^{-4}
			(b)	1.91×10^{-4}
			(c)	1.59×10^{-4}
			(d)	0.795×10^{-4}
	Maximum (1)	Minimum (2)		
Total	(a) 9.77×10^{-4}	(a) 5.89×10^{-4}		
	(b) 7.86×10^{-4}	(b) 3.98×10^{-4}		
	(c) 7.54×10^{-4}	(c) 3.66×10^{-4}		
	(d) 6.75×10^{-4}	(d) 2.87×10^{-4}		

Note: See Table II and text for explanation of cases (a), (b), (c), and (d).

caused by forest removal as a globally averaged phenomenon, which is unsatisfactory since deforestation is highly localized and therefore the effects of such local surface modification could be magnified and possibly transmitted by the action of the general circulation of the atmosphere (Namias, 1979; Chervin *et al.*, 1980).

The nature of local feedback effects which could amplify the climatic impact of tropical deforestation are complex. Initially the increased albedo is likely to be offset by the reduced ability to lose energy through evapotranspiration and surface temperatures could increase (Dickinson, 1980). Initially increased surface temperatures could stimulate increases in convection and therefore, if there is sufficient water vapour available (say, transported from an upwind source area), cloud formation and possibly precipitation will increase. [This would be in contrast to the suggestion of Charney (1975) in which removing vegetation in arid areas leads to a decrease in absorbed energy and hence an overall descent of air over the modified region]. Hydrologists have not yet been able to make detailed studies of the local vs. regional movement of water vapour, and therefore estimating the environmental impact of forest removal and agricultural irrigation is difficult. Modelling studies by Lettau *et al.* (1979) suggest that a considerable proportion of the precipitation over the Amazonian forest results from regional evaporation rather than from advected moisture. The decrease in evapotranspiration following deforestation could be ~600–800 mm yr^{-1}. This must lead to increased local runoff if precipitation rates remain constant. The interactions between the perturbed energy and water cycles as a result of deforestation are likely to be very complex.

The interrelationships between radiation fluxes, vertical velocities, the hydrosphere and surface can be investigated using a general circulation climate model. It is likely

that while the local-scale change will be a function primarily of the nature of the local environment, even at this scale perturbations in precipitation may depend upon the regional climatic regime. Regional and global scale effects will be a result of feedback from the local change and will, once again, be a function of the geographical location of the perturbed area.

Here we report on the results of a simulation of deforestation using the GISS GCM (Hansen *et al.*, 1983). In this study, the possible impact of tropical deforestation was maximized by concentrating all the alterations of the surface vegetation into one locality: a large-scale deforestation of the Brazilian Amazon region (Figure 2). The magnitude of the modification is equivalent to 35–50 yr of deforestation at the current global rate

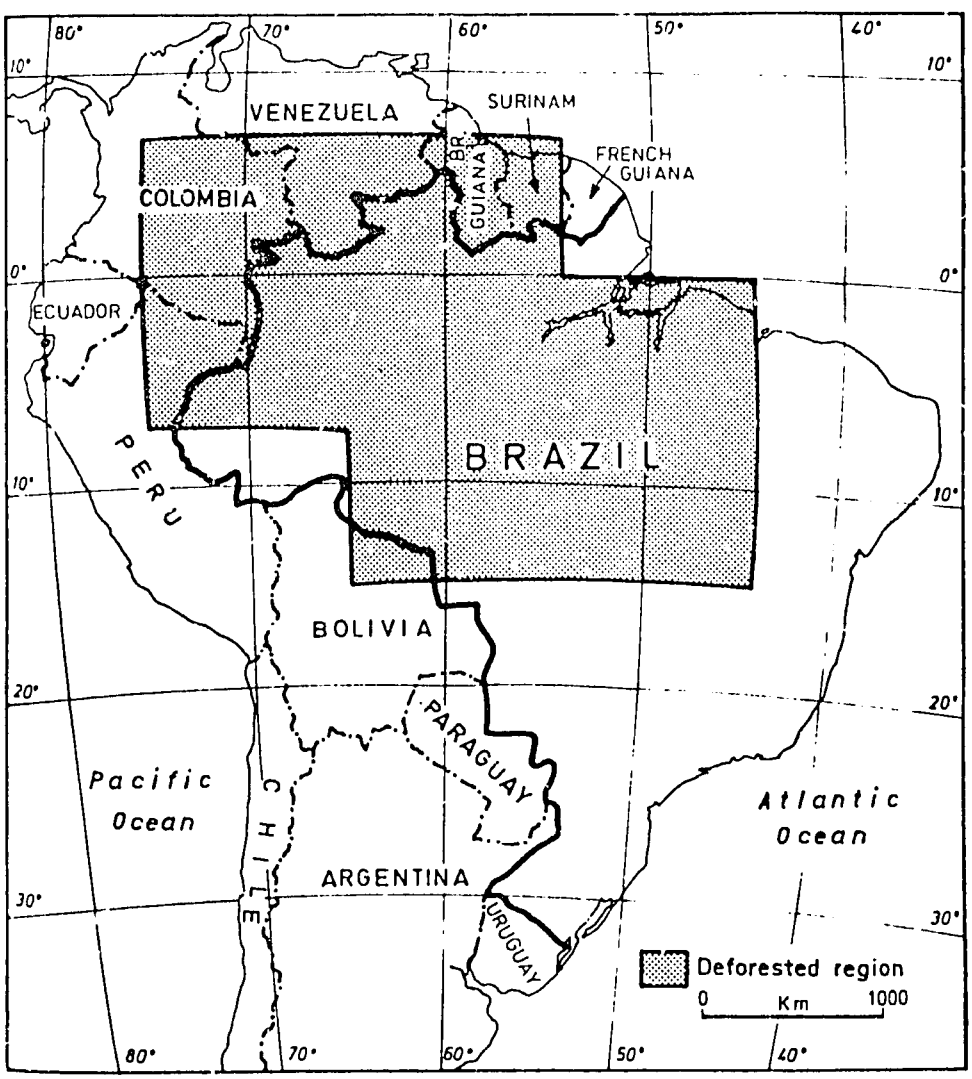

Fig. 2. Amazon region with the GISS GCM grid boxes in which simulated 'deforestation' occurred superposed.

concentrated in the Brazilian Amazon. This is therefore the locational antithesis of the global estimates considered in Section 4.1.

A pair of climatic simulations has been undertaken using the GISS general circulation climate model. This model is designed specifically to permit long-period climatic integrations. It includes computation of many parameters which are specified in other GCMs e.g. sea-ice, cloud cover and type. The version of the GISS GCM used here differs from Model II described by Hansen *et al.* (1983) only in that ocean surface temperatures are computed. Ocean transports are fixed so that the resulting sea surface temperatures are identical to those in Model II. Inclusion of more physically realistic oceanic parameters requires simulations of 5–10 yr before equilibrium is achieved. The model is computationally very efficient (Hansen *et al.*, 1983), which has the *advantage* of permitting investigation of possible teleconnections between the disturbed region and other areas of the globe which could take a number of years before becoming obvious. Naturally there is a compensating *disadvantage* – in this case the spatial resolution selected (8° x 10°) is coarse. We believe that the advantage of including more of the components of the climate system outweighs the lower resolution. The effect of the 'deforestation' is ascertained by comparison of the control and deforested climate integrations. Many shortcomings inherent in the control simulation can be ignored as it is the *relative* alteration that is reported here. Despite this it is worthwhile to compare results from the control simulation with observational data.

The climate of northern Brazil is dominated by the seasonal movement of the equatorial trough but is also affected by the changing pressure fields at higher latitudes (Figure 3). The Atlantic anticyclone is more important than the Pacific high pressure system since the Andes 'protect' Brazil from changes occurring to the west. A heat low becomes well developed in the centre of the continent in the summer which results in slightly weaker easterly winds over the Amazon region compared to the winter pattern (Figure 4) although winds are generally weak and variable in all seasons. In response to the formation of the continental heat low in the summer, air flows southward from the Amazonian region. This phenomenon has been compared to a monsoon circulation pattern. Advected air loses moisture by precipitation as the prevailing easterlies approach the continental divide. Much of the water within the Amazon basin is recycled through the atmosphere, possibly more than once (Molion, 1975 and Lettau *et al.*, 1979). Figures 3–5 compare climatological data and the results of simulations of the global circulation pattern by the GISS GCM (Hansen *et al.*, 1983). Although modelled precipitation is significantly lower than observed (Figure 5), many aspects of the local climatology and hydrology including the diurnal cycle of temperature and rainfall, are fairly well simulated.

Deforestation in the Amazon has recently been observed to be accelerating. In 1979, a joint survey by the Brazilian National Institute for Space Research and the Brazilian Institute for Forestry Development using LANDSAT imagery revealed that ~1.5% of the total Brazilian Amazon forest had been cleared between 1975 and 1978 (Tardin *et al.*, 1979). This represented a 170% increase in deforestation over the 3 yr period from 1973–1975. Around 260 000 km^2 of forest has been cleared since the 1950s. If this estimate is correct (Myers, 1980b), then over 5% of the original tropical moist forest

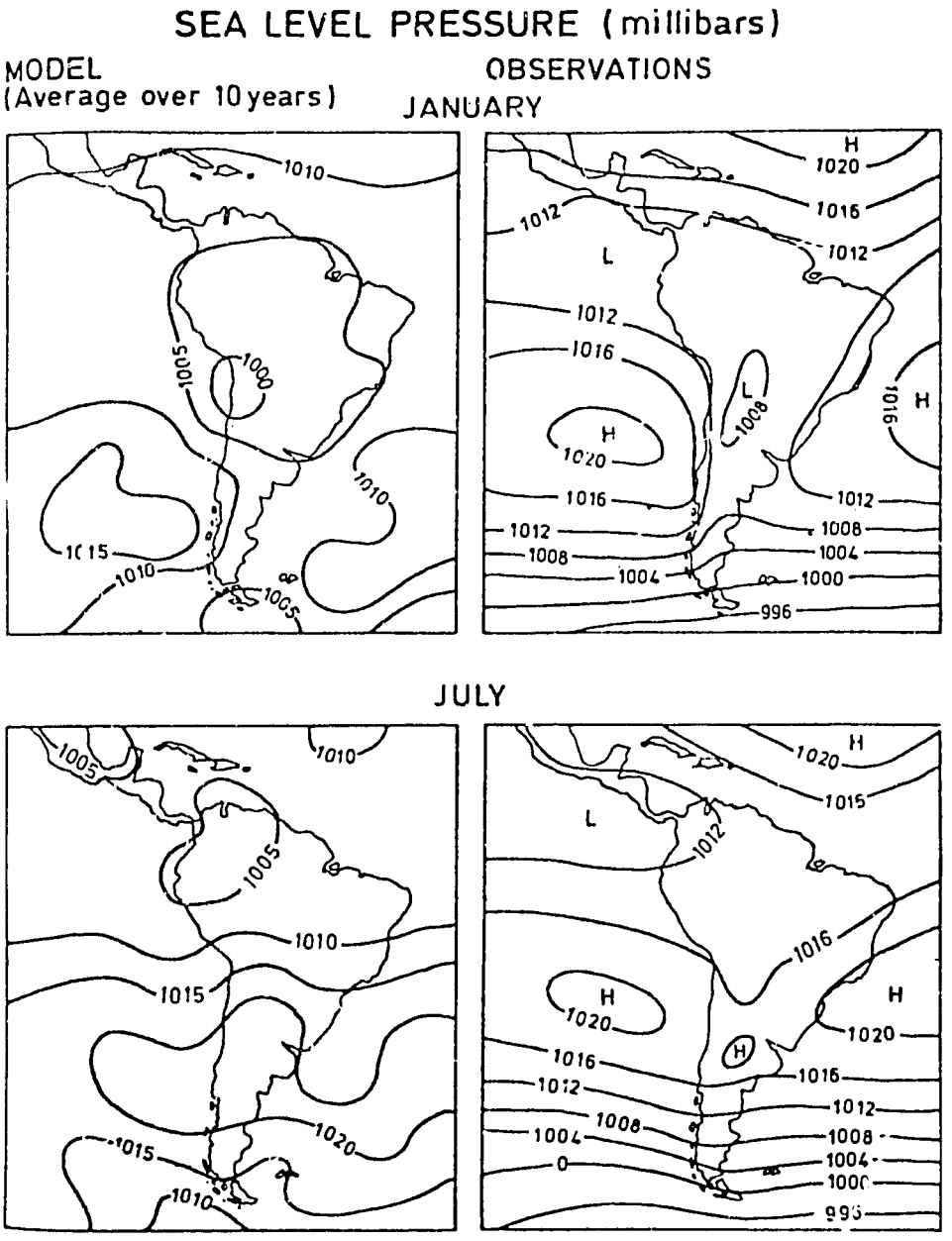

Fig. 3. Sea level pressure for January and July: the control run (left) and observations (after Schwerdtfeger, 1976) (right).

has already been removed. It is important to note that it has not been established that climax vegetation (a vegetation assemblage which has evolved to a stage of equilibrium under given climatic and soil conditions) ever existed over the whole of the area which bioclimatologically could support it (Sommer, 1976).

Prior to the completion of the LANDSAT survey (Tardin *et al.*, 1979), an estimated

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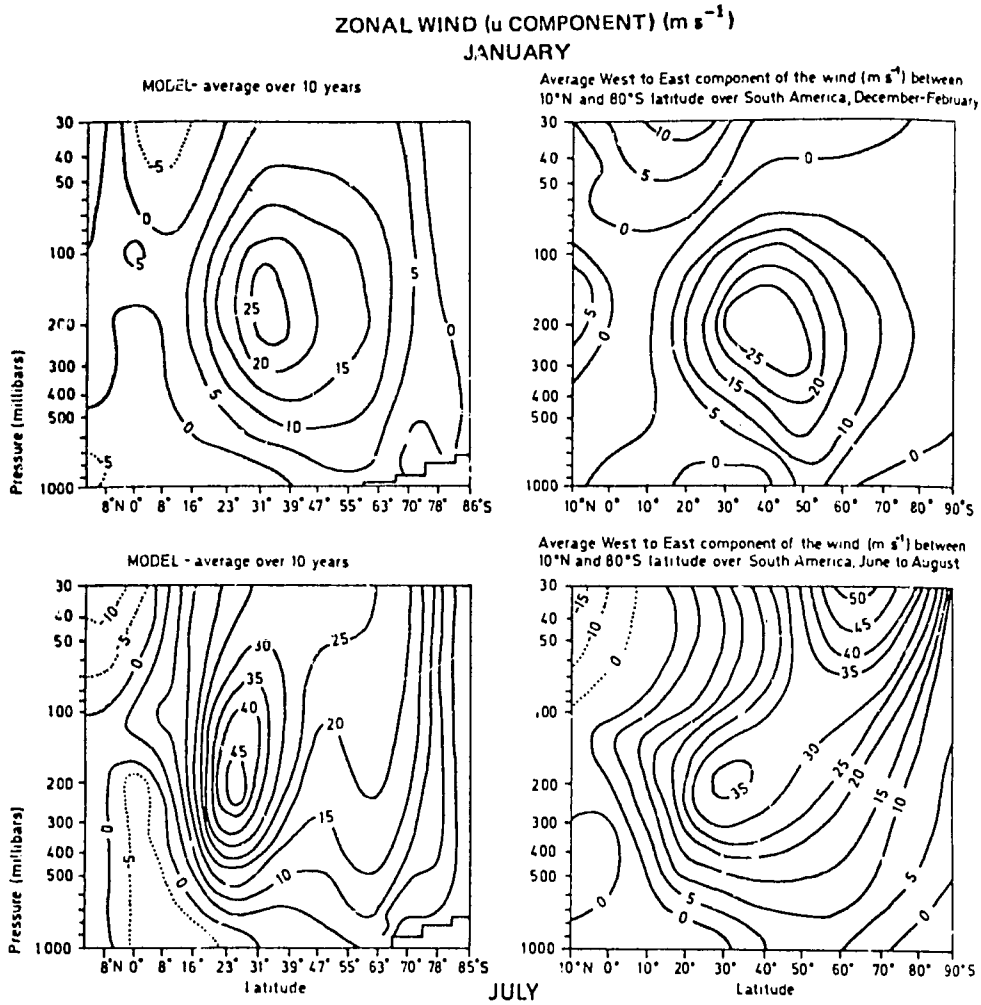

Fig. 4. Zonal wind (m s^{-1}) for January and July: control run (left) averaged over all longitudes and observations (after Schwerdtfeger, 1976) (right) averaged over longitudes appropriate to continental South America.

~115 x 10^3 km^2 of forest had been cleared during the 11 yr period 1966–1977, i.e. an annual average rate of ~10^4 km^2 (Myers, 1980b). This was comparable with the official 'expected' rate calculated by addition of the various known clearance plans (e.g. cattle ranches, small-holdings, other agriculture, highways and timber). This total suggests an average rate of approximately 12 800 km^2 per year. The annual removal rate seems to be increasing from ~9 500 km^2 yr^{-1} for the 1973–1975 period to ~16 000 km^2 yr^{-1} for 1976–1978.

The deforestation input to the GCM simulation is much greater than any likely near future deforestation of Amazonia. It was intended here to maximize any climate perturbation resulting from a 30-yr total tropical deforestation for comparison with the global calculations which represent a minimum likely response from simpler climate models

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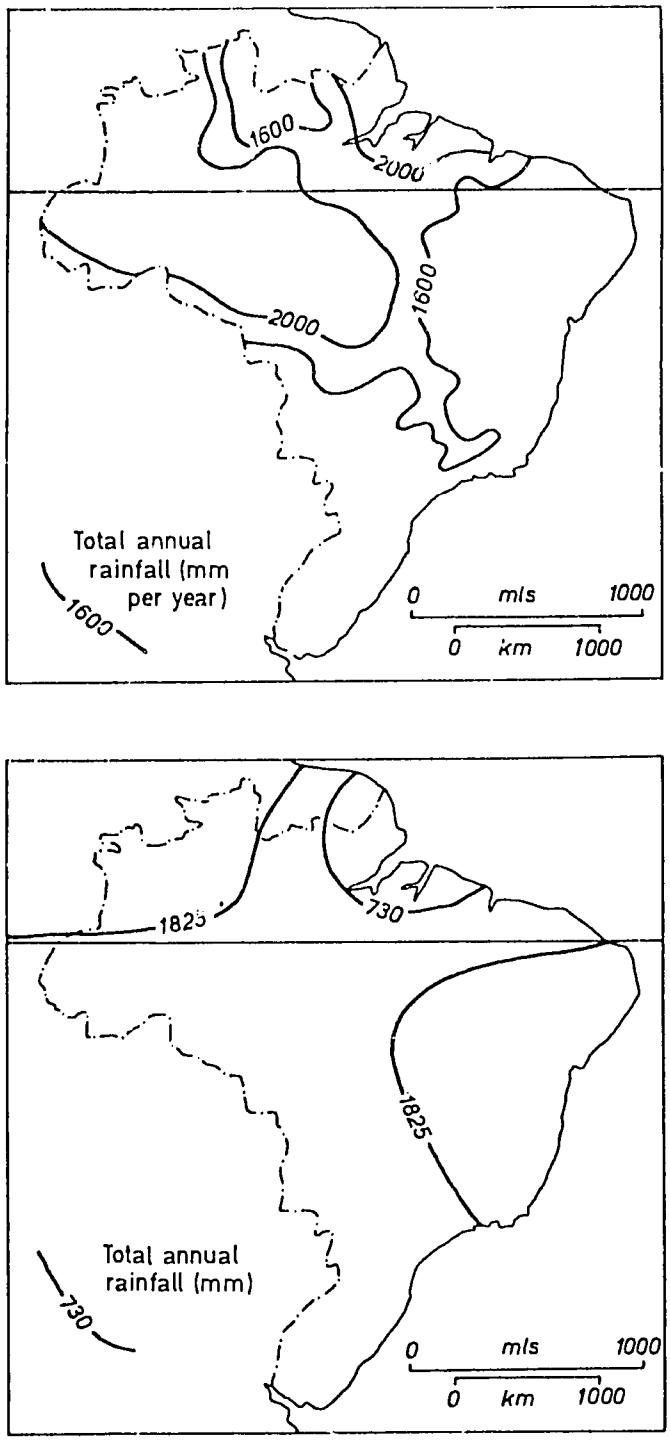

Fig. 5. Total annual precipitation: upper figure observations (after Schwerdtfeger, 1976), lower figure control run.

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(Section 4.1). Therefore the global tropical forest removal was concentrated on the Amazon region.

Figures 2 and 6 show the region of South America for which a simulated vegetation change was made. A total area of 4.94×10^6 km^2 of tropical moist forest was removed and replaced by grass/crop cover. The 'deforestation' occurred in 7 of the $8° \times 10°$ grid squares of the medium resolution model lying between 7.8° N and 15.65° S and 45°–75° W. All the areas classified as tropical forest were altered to grass/crop vegetation type but

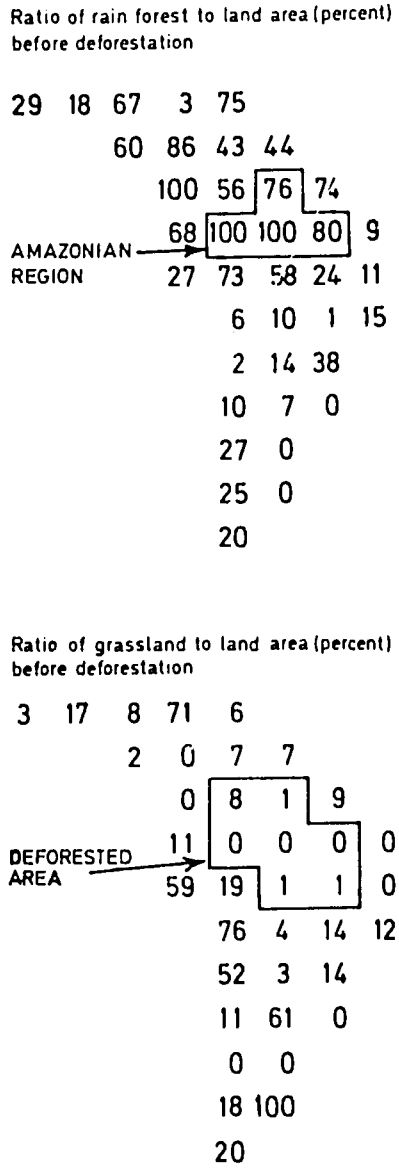

Fig. 6. Grid elements of the GISS GCM (medium resolution) showing % rain forest and grass cover in the control case and indicating the regions of 'deforestation' and of averaging for statistics used in Figure 7.

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other vegetation types already existing in the squares remained unaltered. The surface albedo was increased as a result of the vegetation modification. This effect is particularly noticeable in the near infrared spectral region where grass/crop cover is known to exhibit high albedos (values range from 0.1 at 0.5 μm to 0.35 at 1.0 μm). The roughness length is also significantly affected by the vegetation change due to the small scale of topography in this region. An anticipated climatic response to the changes shown in Figure 2 would be a lowering of surface temperatures together with a reduction in turbulent fluxes from the surface to the atmosphere, similar to the results of Potter *et al.* (1975, 1981). However the field capacities, which are related to the vegetation in the GCM are also perturbed by the simulated deforestation. The consequences of the subsequent hydrological changes are shown below to reduce the surface temperature change to close to zero. The importance of including as many features of the land surface processes as possible in climate models has been underlined by Shukla and Mintz (1982). Hansen *et al.* (1983) describe the land surface parameterization and the interaction between the land surface and the boundary layer in detail. Rind (1982) also uses the GISS GCM in an investigation of land surface hydrology and finds that the model reproduces the essential features of this system. It must be noted, however, that whilst the GISS GCM successfully reproduces major features of the general circulation (Hansen *et al.*, 1983) and the essential elements of regional pressure and wind fields (Figures 3 and 4) the central simulation is far from perfect. For example rainfall is underpredicted in the control run of the model (Figure 5). Despite some problems, we believe that the long-term simulation described here is at least as valid as other land surface investigations which were with much more restricted GCMs (e.g. Sud and Fennessy, 1982). The area within which the vegetation change occurred and the 4 boxes for which averaged Amazonian statistics are derived are indicated in Figure 6.

The effects of deforestation in the Amazon have been assessed by comparison with a control (standard) climatic simulation. The values shown in Figure 7 are from years 16–20 of the standard control simulation and the last five years of the ten year perturbation (deforestation) run. The results of deforestation were to alter significantly all the climate parameters considered, except the surface temperature. Figure 7 shows the departure of the hydrology from the control run particularly well. The rainfall has decreased by between 0.5–0.7 mm day^{-1} and evaporation and total cloud cover are both reduced. The effects of reduced rainfall and increased run-off are shown in the significantly lowered values of water available in both ground layers. Figure 7 shows that despite the increase of surface albedo from 0.11 to 0.19, the temperature does not decrease. This is possibly because the reduction in evaporation caused by a combination of less available water and reduced ability to transpire has offset the radiative cooling by an evaporative (or latent heat) warming. The overall effect is a negligible temperature change. It must be noted that the surface is not the only area in which the hydrosphere seems to oppose the input alterations. The decreased cloud also opposes the increase in the surface albedo, leading to a smaller overall planetary albedo increase than would have been expected from a model which did not incorporate both cloud prediction/interaction, e.g. the model used by Sagan *et al.* (1979).

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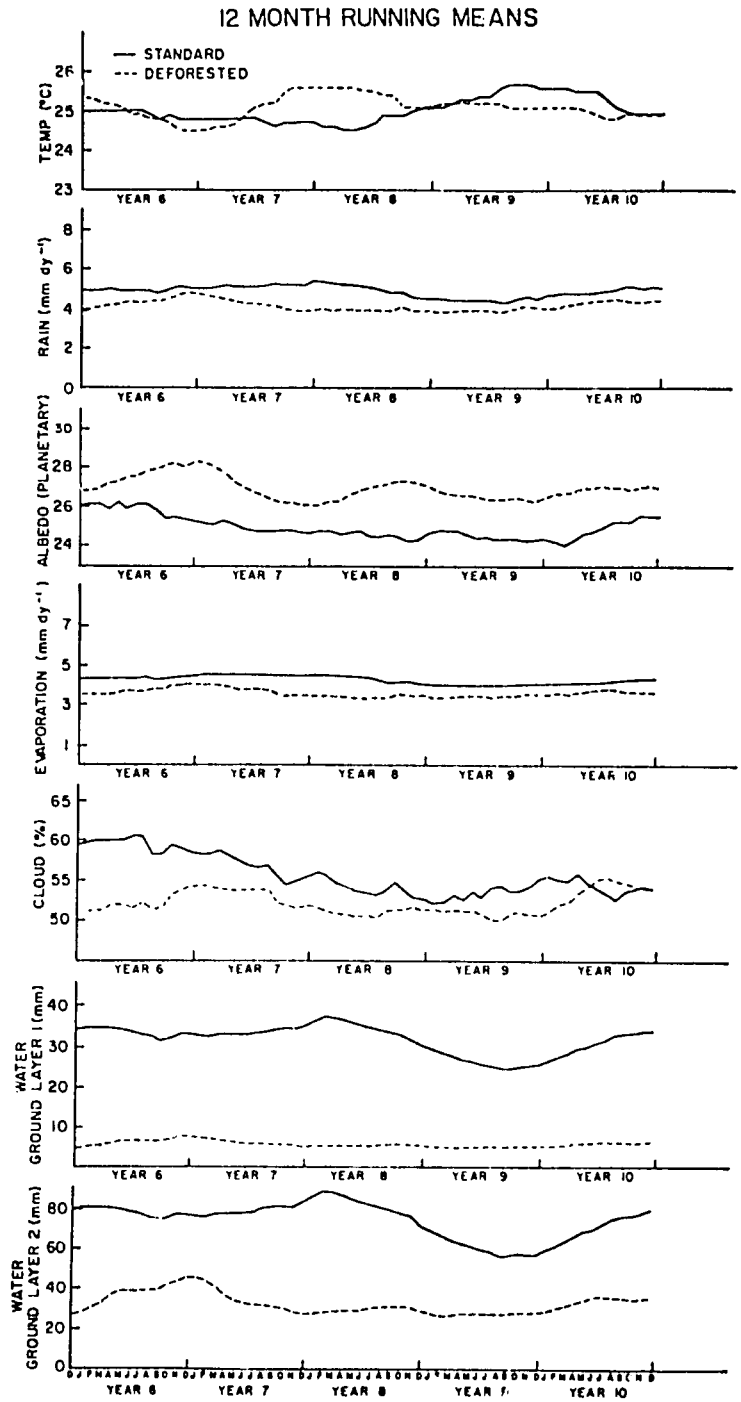

Fig. 7. 12 month running means for the last 5 yr of the control and deforestation runs (see text for description of simulation periods). The four grid elements used are shown in the upper part of Figure 6.

The local results of the deforestation experiment can be summarized as follows from Figure 7: no change in the surface temperature; precipitation decreased by around 0.6 mm day^{-1}; evapotranspiration decreased by 0.4–0.5 mm day^{-1}; planetary albedo increased by between 1% and 1.5% as a combined result of the increased surface albedo and the decrease in cloud cover (the latter is a highly variable parameter – see Figure 7) but we estimate that it has decreased by between 5% and 15%.

There are two ways in which the changes described here could affect the global climate; either through a circulation modification or, as suggested by those using simpler climate models, the local planetary albedo increase could cause an alteration in the global climatic regime. In the latter case a global temperature decrease would be expected. There is no such temperature deviation that can be detected above the natural variability of the standard climatic regime in our simulation. Even in the latitude band in which the surface change occurs (width 16°, centered on 0°) there is no climatic change detectable above the model variability. This seems to be because certain of the feedback effects resulting from the deforestation act in opposition to one another. For instance, surface temperatures rise in spite of an increased surface albedo because the reduced evapotranspiration dominates the response. Similarly enhanced convection which might be expected to result from occasional temperature rises appears to be opposed by the moist convective cooling. It seems unlikely, at least from the results of this simulation, that even massive deforestation of the Amazonian region can cause global scale temperature effects directly. The Hadley circulation, as monitored by the stream function, differed significantly from that of the standard simulation only in one area at a pressure level of ~200 mb. Even after ten years of simulation the effects of a very large deforestation could be detected only in the immediate vicinity of the perturbation.

There remains, however, the possibility that the local scale changes already noted could become coherent enough to establish a significant regional departure from the standard model's circulation. If, for example, the climatic change is large enough to affect ocean temperatures then a perturbation in the Walker circulation may occur. The possibility of such an interaction may, however, be hindered by the intervening topographic feature (the Andes). If such a regional anomaly were to be established, its effects could impinge upon the Hadley circulation. Such systematic regional departures were suggested by Potter *et al.* (1975 and 1981) from the results of their 2-D climate model. We have examined the hypothesis that the disturbance to the normal circulation pattern in the region of the Amazon could be large enough to cause a feedback effect, which feeds into the general circulation of the atmosphere, by analyzing the vertical velocity field in the area of the perturbation. Figure 8, based on the averaged data from the standard run, shows the ascent of air over the continental regions between 8° N and 16° S latitude for January and July. The classical Walker circulation cells show clearly in the January simulation (see Julian and Chervin, 1978, 1980). One of the major areas of ascent occurs over the Amazonian region. Figure 9 is a contour plot of a climatic statistic derived following Chervin and Schneider (1976), which presents the perturbed vertical velocities minus the standard vertical velocities (i.e. those in Figure 8) divided by the standard deviation of the standard run where all variables are derived from 5 yr simulations. Fol-

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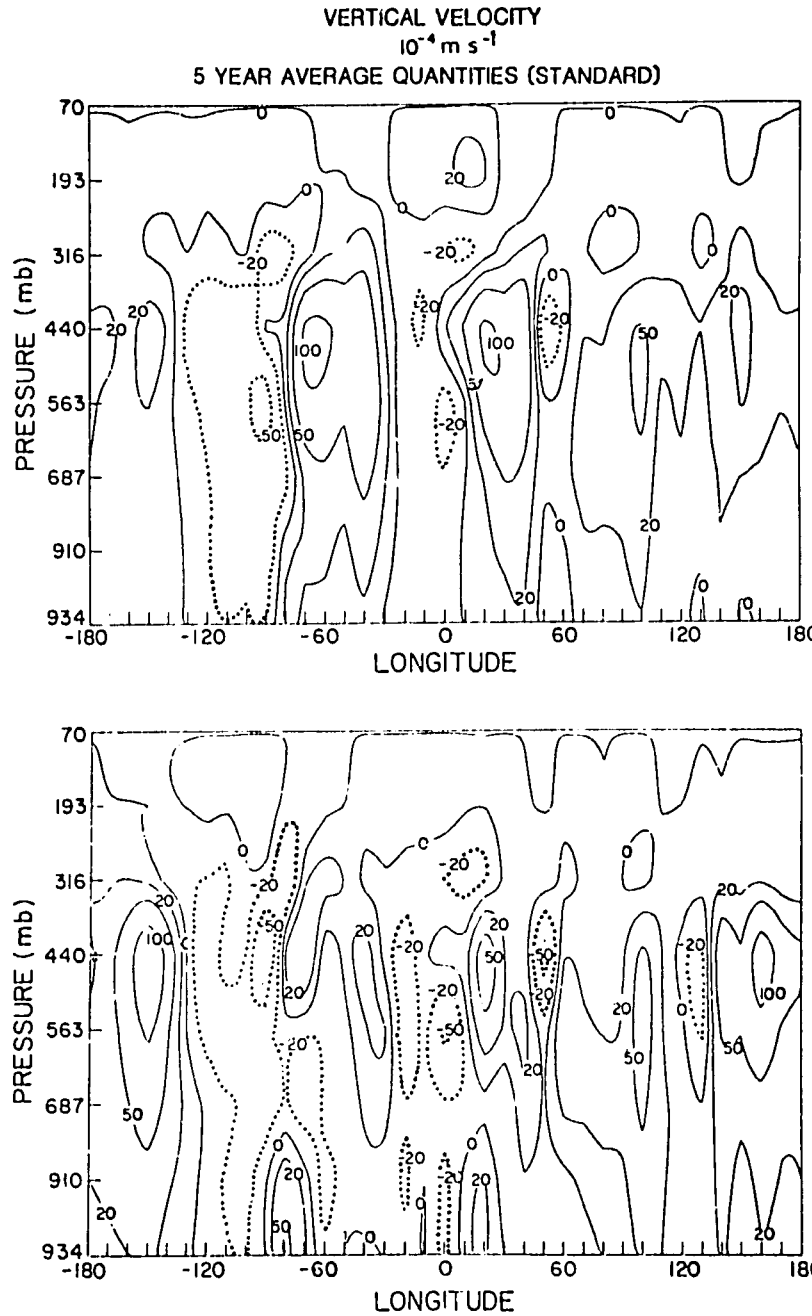

Fig. 8. Vertical velocity in the latitude band from 8° N to 15° S (10^{-4} m s^{-1}) for January (upper) and July (lower); 5 yr average from the control run. The cells of the Walker circulation are clearly seen in the January case.

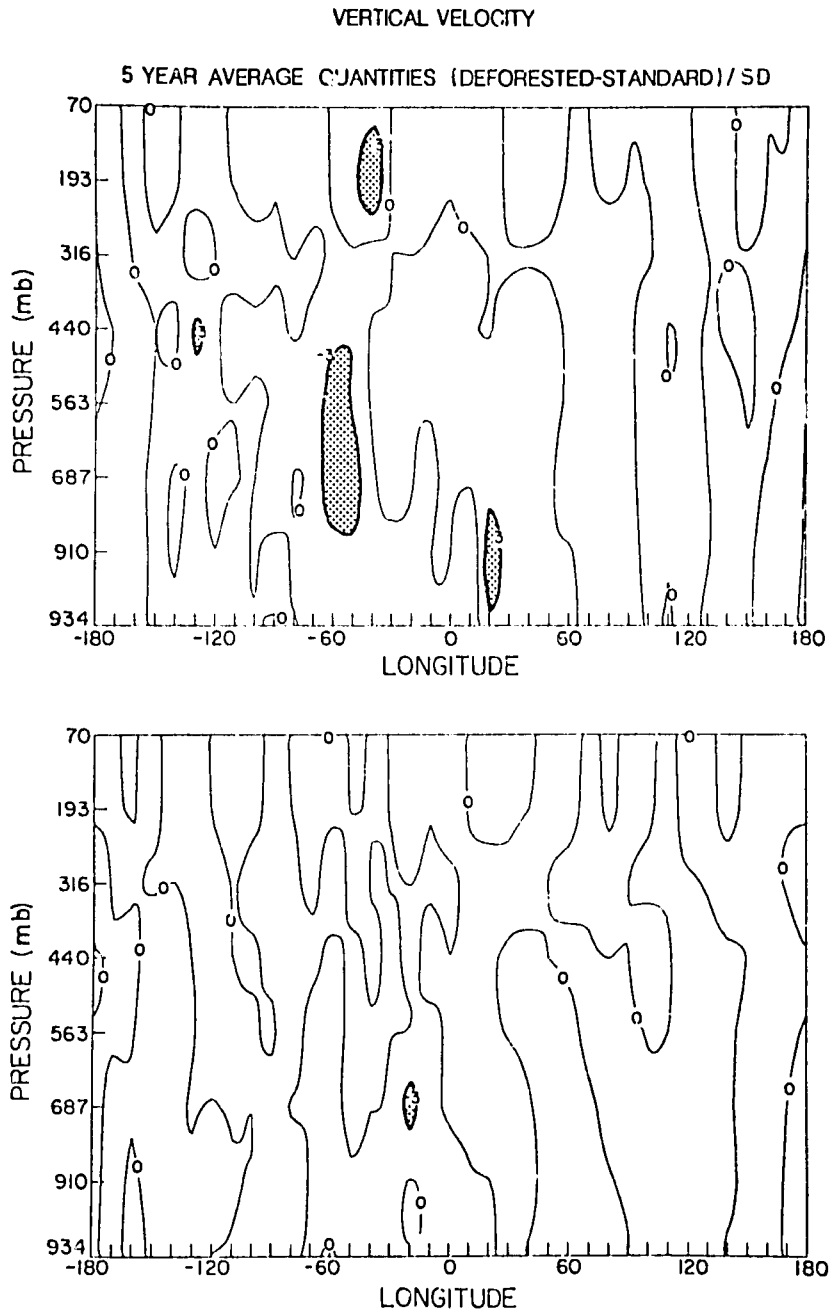

Fig. 9. 5 yr average of climatic change test statistic = (deforestation-control)/(control standard deviation). Values greater than ±3 occur in regions where the perturbation vertical velocity differs significantly from the control field (99% confidence level); January (upper), July (lower).

lowing Chervin and Schneider (1976) and Chervin (1981) it can be shown that for an absolute value of this statistic greater than 3 there is a significant difference between control and perturbation at the 99% confidence level. Figure 9 shows only a very small region of significant departure from the standard in July although in January there is a significant decrease in the vertical velocity above the deforested region. This is consistent with the reduction in the surface evapotranspiration and the reduction in the moist convective heating aloft. This latter effect becomes statistically significant at two levels in the atmosphere in January only. The resultant decrease in the vertical velocity over the deforested area is similar to that predicted by Charney (1975) as a result of desertification. In this case, however, there is no strengthening feedback and decreased upward motion never becomes descent. Nor does the effect ever penetrate beyond the area local to the perturbation. There seems to be no reason to suspect that further changes will take place in the circulation pattern in subsequent years.

In summary the GCM simulation indicates that, within the limitations of the model, there is no reason to suggest that any regional to global scale effects on the climate can be expected as a result of large-scale tropical deforestation.

5. Conclusions

The possible climatic impact of tropical deforestation has been considered using two extreme prediction modes. The calculations presented in Section 4.1 were derived from simple climate models, which must estimate the effect of a climatic forcing factor as though it were globally averaged. The results presented in Section 4.2 are from the GISS 3-D global climate model. In this case, the probable effects of a highly specific alteration to the surface and atmospheric features of one locality can be considered. In both cases the outcome for the global climate was similar, i.e. the alteration caused to the global temperature regime was undetectably small or zero. This result does not imply that local climatic changes will be unimportant. On the contrary the GCM results show that the effects upon the deforested area itself are considerable – e.g. a reduction in rainfall of around 200 mm yr^{-1}.

One of our major conclusions is that the effects of the hydrosphere, which have generally been neglected in simpler climate models, must be included if useful simulations are to be produced (see also Manabe *et al.*, 1981). The climatic effects of a surface albedo change seem to be smaller in a very moist atmospheric environment than in arid regions (cf. Charney *et al.*, 1977 and Sud and Fennessy, 1982). This conclusion cannot be drawn too strongly in view of the severe limitations of *all* climate models.

The changes, potentially important for the climate, currently being caused by man in the tropical forests can be listed under the headings of (i) surface albedo; (ii) CO_2 levels; (iii) hydrospheric changes; (iv) addition of particulates; (v) perturbations to surface atmosphere exchanges and to the atmospheric part of the hydrosphere. Estimates of the perturbations caused can only be made if the rate and mode of vegetation removal are well known.

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It is very important that we draw attention to the fact that only first order surface modifications have been considered here. No attempt has been made to consider the impact on the global carbon budget of either the initial removal of the forest vegetation or the subsequent loss of this large sink of atmospheric carbon. Also it is likely that the forest clearance itself would produce atmospheric particulates as well as additional CO_2 (e.g. Seiler and Crutzen, 1980).

Here we have sought to quantify the current extent of tropical deforestation and to calculate its possible impact upon the climate primarily as a result of the albedo change. The primary result of our studies is that in both simulations there is no discernible alteration in the global climate despite the exaggerated nature of the input perturbation. We recognize that predictive results from all types of climate models must be treated with caution. We believe that the twofold nature of our investigation adds weight to our result. It is very important to point out that we *DO NOT SUGGEST* that tropical deforestation is unimportant. On the contrary we believe that the destruction of the tropical rain forests is highly undesirable. Locally the effects are devastating. Even the coarse resolution climate model employed here shows significant decreases in precipitation and evaporation on a local scale. More importantly the effect of forest removal upon the global budget of CO_2 should be a primary concern for mankind. Despite these assertions, we cannot support the suggestion that surface albedo changes associated with tropical deforestation are significant for regional or global-scale climate.

Acknowledgements

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