THE NATURE AND SIGNIFICANCE OF THE FEEDBACK OF CHANGES IN TERRESTRIAL VEGETATION ON GLOBAL ATMOSPHERIC AND CLIMATIC CHANGE

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Abstract. The potential feedback on global atmospheric and climate change of climate-driven changes in terrestrial vegetation is examined by systematically relating the surface exchanges of energy, mass and momentum to two dimensions of vegetation, structure and taxonomy, such that the significance of climate driven changes in these characteristics can be assessed. A detailed quantitative understanding of this feedback is an important prerequisite to realistic and dynamic representations of the Earth's surface within general circulation and biological models (GCMs and GBMs). Without realistic representations of terrestrial vegetation within these models, any forecasts of future climates by these models must be suspect.

Several general conclusions are drawn. The first is that the indirect feedbacks, those associated with the clouds and aerosols of the planetary boundary layer, appear to be very powerful but as yet their behaviour and connections with the underlying surface are both poorly understood and captured within GCMs.

The physical structure of vegetation, the disposition of biomass in 3-D, is the characteristic that most strongly influences the exchange of momentum (via aerodynamic roughness) and solar radiation (via albedo). Vegetation structure and species composition determine the most important of the mass exchanges, evapotranspiration. Of all of the surface exchanges, the parameterization of evapotranspiration (E_T) and the simulation of the water balance over time is the most critical.

Lastly, the problems of scaling and spatial heterogeneity, the sub-grid variability of the modellers, looms as a difficult, but not insoluble, problem. It remains a critical problem however, and the detailed parameterization of the various 'big leaf' models stands in absurd contrast to the simplistic generalization of the spatial heterogeneity of terrestrial landscapes.

Plant ecologists can contribute to the task of improving the representation of vegetated landscapes within GCMs. There is need to simply and unify the way in which vegetation can be grouped at landscape scales. A classification that is based on function rather than phylogeny is required. The definition of Vegetation Functional Types (VFTs) would expedite research on both the impact of, and feedback on, climate change.

Introduction

This paper reviews the interactive role of the land surface in global climate change by considering the feedbacks on atmospheric and climatic change of changes in terrestrial vegetation where these may be climatic or anthropogenic.

This review concentrates on the ecological rather than the climatological aspects of global change; a strategy adopted for the following reasons. Most ecologists recognize the potential of currently forecast global atmospheric and climatic

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changes to drive changes in the structure and functioning of the biosphere. They are concerned with the ecological impacts of the rapid climate change currently forecast and are impatient for detailed predictions rather than the scenarios of change currently available. The general statements about latitudinal warming and the imprecise probabilities of shifts in rainfall systems lack the quantitative, spatial definition required by ecologists to determine the significance of these changes for terrestrial ecosystems.

To progress, to effectively plan for and manage future ecological change, either for agriculture or for species conservation, requires a substantial improvement in the capacity to forecast future climates for differing, and as yet unknown, scenarios of greenhouse gas concentrations. Because of the dynamics of the interaction of the greenhouse – modified energy budget of the planet and the circulation of its atmosphere and oceans, the only source of predictive understanding are the models that simulate the complex dynamics of this system: the general circulation models (GCMs). Thus, while most ecologists recognize that GCMs represent the only source of the required climate scenarios, they are disillusioned with the poor spatial resolution used (e.g. 500 × 500 km in the horizontal) and confused by the disagreement between models, e.g. Schlesinger and Mitchell (1987). Also the utility and validity of GCM forecasts appears questionable given the variety of representations of the Earth's surface used. Some models incorporate considerable dynamic detail of the oceans and vegetated surface while others use arbitrary, static and simplistic representations.

A constructive approach is that, as ecologists will be almost completely dependent on GCMs to provide forecasts of future climates, it is critical to their modelling of ecosystem change that these forecasts be both realistic and relevant. Realism can only come from GCMs that incorporate justifiable representations of, inter alia, the Earth's surface. Thus it becomes entirely in the self-interest of ecologists to contribute to the parameterization and modelling processes. The relevance of GCM output for ecologists, in terms of requirements for means and variances of the driving variables of precipitation and temperature and the explicit consideration of sub-grid processes, is of equal importance as realism, and must continue to be impressed on the climate modelling community by the activities of IGBP (e.g. Walker and Graetz, 1989).

Terrestrial ecologists can play a key role in the quest for predictive understanding of climatic change as a whole. Most ecological research activity will be focussed on the modelling of climate-driven change in terrestrial vegetation, particularly as it affects agriculture and conservation. If there are parallel efforts on understanding the reciprocal process, the feedback of vegetation on climate, then a complete loop of interaction is closed, and the best quantitative understanding can be included in the evolution of GCMs over the next decade or so.

These efforts will only be successful if there is a commonality of description between the disciplines of terrestrial ecology and climate modelling. This commonality can be achieved if the great taxonomic detail of vegetation essential to understanding ecological processes can be simplified into fewer groups that have functional equivalence in terms of the exchange of energy, mass and momentum between landscape and the atmosphere. The principal objective of this review is to contribute to this simplification, and to the derivation of Vegetation Functional Types (VFTs).

The Climate System

In general terms the complete climate system consists of five physical components – the atmosphere, oceans, cryosphere, land surface and biomass, GARP (1975). Figure 1 is a simplification of that system which illustrates its nature and from which we can assess the relative importance of the terrestrial component of biomass, vegetation.

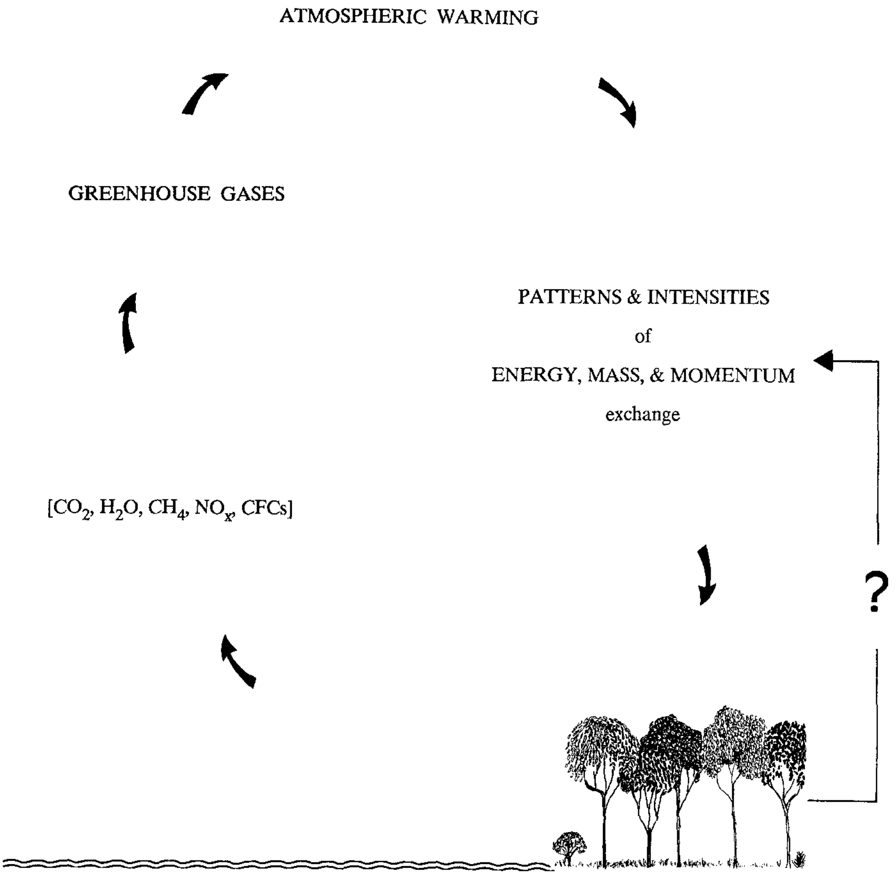

Fig. 1. A diagrammatic simplification of the biosphere-atmosphere interaction in the climatic system. The relative surface areas of the globe that are ocean (71%) and land (29%) are represented and the vegetation is intentionally depicted as sparse. There is no doubt that changes in the patterning of energy, mass and momentum exchange will derive changes in the vegetation cover of the land's surface: the impact. What is not yet clear is the nature, dynamics and significance of these changes in vegetation for further climate change: the feedback.

The greenhouse gases, CO_2, H_2O, CH_4, NO_x, CFCs, from a variety of sources contribute to an atmospheric warming by modifying the radiation budget. This atmospheric warming, in conjunction with a dynamic interaction of the atmosphere with the Earth's surface, will produce modified temporal and spatial patterns of energy, mass and momentum exchange between the atmosphere and the ocean and land surfaces. The magnitude of the greenhouse warming of the atmosphere has been predicted simply on the basis of the radiation budget of the planet, e.g. Schlesinger and Mitchell (1987). The resultant changes in patterns of atmospheric behaviour are complex; those in space are largely determined by hydrodynamics while those in time (intensities) are determined by the thermodynamics, which in combination are the focus of GCMs. The exchange of energy, mass and momentum, i.e. weather, becomes the climate when averaged over time for any one location in space. Thus for the relatively immobile terrestrial ecosystems, and less rigidly for oceanic ecosystems, the rate of climate change experienced by these two components will be the relative rates of movement in space of the average weather and of the biomass.

There is no doubt that changes in the magnitude and patterning of energy, mass and momentum exchange between the atmosphere and the land will drive changes in the vegetation cover of the terrestrial surface (the impact). What is yet to be assessed is the nature, dynamics and significance of these changes in vegetation for further climate change (the feedback).

Several points need to be emphasised from Figure 1. The first is that the relative areas of sea and land surface are proportionally represented; in the Southern Hemisphere it is approximately 0.80/0.20, in the Northern Hemisphere 0.60/0.40, which together give a global average of 0.71 ocean surface to 0.29 land surface. Obviously the fundamentally different physical characteristics of land and oceans will influence their respective interactions with the atmosphere. The ocean greatly surpasses the capacity of the land in terms of a source of water, a sink for solar radiation and CO_2 but is exceeded by land as a sink for momentum. In spite of these great differences in surface properties, the relative importance of ocean and land surfaces in the interaction with atmosphere cannot be assessed by their respective areas alone. The contribution of both to the global energy balance is considerably moderated by the presence of clouds and dust in the planetary boundary layer (PBL).

Lastly, in Figure 1 the vegetation is deliberately depicted as sparse to emphasize that the majority of the Earth's surface presents both bare soil and vegetation surfaces to the atmosphere. Vegetation cannot be realistically generalized as either bare soil or vegetation or as a green film of one dimension, depth, as is done in the 'big leaf' parameterizations. The importance of vegetation structure will be enlarged upon later.

The Earth's climate has been remarkably stable over geological time scales having remained temperate for at least 3.5×10^9 yr; a time span comparable to the age of the solar system (Kasting, 1989). Climate stability is defined on this time scale by a continued presence of liquid water and life on the Earth's surface, the first condition being a prerequisite of the second. Runaway glaciation is prevented by the negative feedback inherent in the carbonate-silicate biogeochemical cycle, and a runaway greenhouse effect is apparently impossible with present (or past) solar fluxes (Kasting, 1989). A remarkable demonstration of the short-term stability of Earth's climate is that published by Idso (1989), Figure 2. The analysis is repeated here because it demonstrates the damped oscillatory response of surface air temperatures to the perturbation of volcanic eruptions; a response function that can be reasonably interpreted as representing system interactions that not only resist changes but also corrects for disturbances of either sign. But what is particularly relevant to this discussion is the demonstrably different behaviour of the Southern compared with the Northern Hemisphere, a difference interpreted by Idso as being due to the larger surface area of oceans.

The uncertainty associated with the feedback of terrestrial vegetation on the atmosphere is quantitative only. In qualitative terms the feedback is readily demonstrable; e.g. Lockwood (1983), Balling (1988). The most general example is the recognition by climatologists of the relative importance of surface (boundary) con-

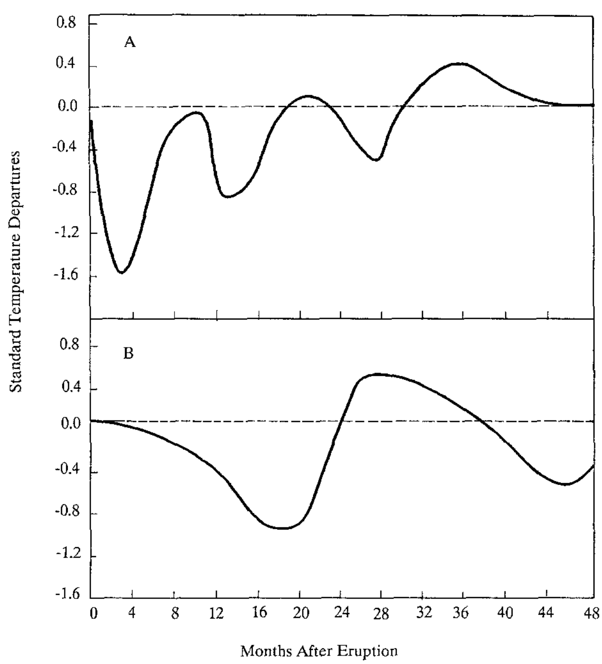

Fig. 2. A comparison of the time responses of the (a) Northern and (b) Southern Hemisphere surface temperatures to volcanic eruptions. The data have been initially grouped using superimposed epoch analysis by Sear *et al.* (1987); redrawn from Idso (1989). In the Northern Hemisphere there are six temperature trend changes before returning to equilibrium whereas in the Southern Hemisphere there are but three such changes in the 48-month data period. The climate systems of the two hemispheres both exhibit damped oscillatory responses but they are very different in characteristics. This difference is most probably due to differences in areas or circulation patterns of the ocean.

ditions compared with the internal dynamics in determining the characteristics of the climate systems of tropical versus extratropical regions (Nicholson, 1988). Over hundreds of millions of years and on global scales, biogeochemical controls of atmospheric CO_2 concentration, and thus of global temperatures, can be demonstrated (Volk, 1989). Over shorter geological time scales, the dramatic decline in atmospheric CO_2 over the last 10 million years has undoubtedly been influenced by the continuing evolution of the photosynthetic systems of plants (Goudriaan, 1987).

On time scales of years but on global scales, the interannual dynamics of atmospheric CO_2 can be related to the latitudinal variations in plant photosynthetic activity as observed by satellite (Tucker *et al.*, 1986a; Fung *et al.*, 1987). Conversely the same satellite data set at continental and global scales demonstrates how terrestrial ecosystems wax and wane as they are driven by the climate factors of precipitation and temperature (e.g. Justice *et al.*, 1985; Malingreau, 1986; Tucker *et al.*, 1986b); Townshend and Justice, 1986).

Thus, there is little doubt that terrestrial (and oceanic) vegetation plays an interactive and significant role in influencing the atmospheric CO_2 level and thus of the planetary energy balance over ecological and geological time scales. The task is to assess the magnitude and dynamics of the three exchanges between atmosphere and vegetation, momentum, mass and energy, and to determine how they are related to vegetation characteristics.

Indirect Feedbacks

The principal feedback is the influence of terrestrial vegetation on the exchanges of momentum, mass and energy between the land surface and the atmosphere; the direct feedback. There is also an indirect feedback; indirect because it involves components of the planetary boundary layer, clouds and dust, rather than the land surface itself. The indirect feedbacks will be considered first.

Dust

Two components of the planetary boundary layer can exert a feedback on climate, dust and clouds. Dust, which strictly includes all water-insoluble aerosols such as smoke, can play an important climatological role (Coakley and Cess, 1985). The role of smoke has been very closely examined in models dealing with the potential impacts of nuclear warfare which predicted global surface cooling under conditions of 'nuclear winter' (Turco*et al.*, 1983).

Smoke in the atmosphere produces cooling in the daytime, has little effect at night, and thus results in a net cooling of the surface. In this modification of the surface energy balance, smoke is more influential than, and different from, dust. The spatial dimensions of this phenomenon are local; the order of tens or hundreds of kilometres, e.g. Robock (1988). Even the extensive burning currently underway in the Amazon basin, which is perhaps the most extensive and intense vegetation modification in modern human history, has not been identified as global in influence and will be of a temporary nature. Smoke therefore is best regarded as a powerful feedback on climate that, by virtue of areal extent and duration, is but seldom realized.

The frequency, extent and duration of wildfire will be closely related to vegetation and land-use type. However, the climatological consequences of biomass burning have still yet to be quantitatively assessed on either regional or global scales.

Dust differs from smoke by being a truly global phenomenon with the two largest sources being Africa and Asia (Iwasaka *et al.*, 1988). The Saharan dust layer alone contributes about 25% of the global water-insoluble aerosols (Legrand *et al.*, 1988). Dust, as with smoke, modifies the surface energy balance by reducing incoming solar radiation, thereby lowering daytime surface temperatures, and longwave loss at night, and increasing nighttime surface temperatures (Legrand *et al.*, 1988; Nicholson, 1988). The generation of dust is clearly related to land-surface characteristics, e.g. Gillette (1988) and McTainsh *et al.* (1989), while its transport and distribution is determined by the characteristics of the atmosphere (Gillette and Hanson, 1989).

For the extensive arid landscapes such as Africa, the capacity of dust to alter the energy balance has been hypothetically linked to changes in atmospheric circulation patterns, e.g. subsidence, which reduce the probability of rainfall, thus generating drought conditions, reducing vegetation growth and increasing dust production. This positive feedback loop has been invoked to explain the persistence of drought in the African Sahel, but conclusive evidence is still lacking (Lockwood, 1986; Rasool, 1984).

Lastly, in the current debate on the origins of global climate change, the episodic massive injections of dust into the stratosphere by volcanoes has been implicated (Sear *et al.*, 1987; Self and Rampino, 1988). This hypothesis is not considered here because it can be regarded as a static geological characteristic of the surface rather than a dynamic ecological one.

In summary, the interactions between the land surface and atmospheric dust are much better understood than those between atmospheric dust and climate.

Clouds and Cloudiness

Cloud type and cover are the two most important characteristics of a system that regulates the radiative heating of Earth (Ramanathan, 1988; Ramanathan *et al.*, 1989). The presence of clouds effectively doubles the albedo of a cloudless Earth, thereby tending to cool the surface, and differentially alters the thermal radiation flux by reducing the loss of longwave radiation to space, thereby warming the surface. The greenhouse effect of clouds is equivalent to a 100-fold increase in the atmospheric CO_2 concentration, Ramanathan *et al.* (1989). The two opposing effects of clouds vary with cloud type and structure; low-level clouds (cumulus) tend to cool whereas high-level clouds (cirrus) warm (Platt and Harshvardan, 1988; Roeckner *et al.*, 1987; Platt, 1989). The net effects of interactions of cloud type and amount by geographic location in time on climate are not yet quantitatively assessed. In fact, the fundamental question of whether clouds cool or warm the Earth cannot be unequivocally answered (Ramanathan *et al.*, 1989).

It follows that the representation of clouds and cloudiness within GCMs becomes critical. If it can be done with reality, then these models can be used to explore and untangle the complex interactions outlined above (Cess *et al.*, 1989; Roeckner *et al.*, 1987; Schlesinger, 1988). The reality of such modelling has been both improved and tested using satellite data from the Earth's Radiation Budget Experiment (ERBE), e.g. Slingo *et al.* (1989). All of these models indicate the strength of the influence of clouds and cloudiness on climate even if the sign of the interaction is not universal.

In contrast, the influence of the land surface on clouds and cloudiness is relatively very weak. There are few scattered reports of connections between changing land surface conditions and consequent changes in cloud type and amount, e.g. Anthes (1984) and Nicholson (1988), although the anthropogenic emissions of SO_2 appear much more closely linked to cloudiness and climate change (Schwartz, 1988; Witley, 1989). The situation for the oceans is however completely different because here there is evidence, albeit disputed, that oceanic surface conditions influence cloudiness via the production of dimethyl-sulphide (DMS) by phytoplankton (Charlson *et al.*, 1987; Schwartz, 1988; Slingo, 1988).

Direct Feedbacks

The land surface exchanges energy, mass and momentum with the atmosphere and vegetation influences the magnitude of these exchanges by its presence. It follows therefore that changes in the vegetation will alter the rates of exchange between surface and atmosphere. This qualitative argument is logical. The difficult task is to quantify the relationship between vegetation and energy, mass and momentum exchange, thereby determining the sign and strength of the feedback on these exchanges of potential changes in vegetation.

For this task, vegetation is assumed to have just two relevant dimensions; structure, the distribution of biomass in 3-D space, and taxonomy, the species composition of that structure. A simplification such as this is a prerequisite for generalizing the micrometeorology of vegetated surfaces. The functional simplification of the description of vegetation appears critical to the modelling of the interactions of climate and vegetation and will be developed further later.

A summary of the exchanges and the key or essential variables with which the process can be parameterized is provided in Table I. The emphasis is placed on modelling the exchange processes within GCMs. Thus the controlling parameters for mass exchange are the resistances or conductances to the fluxes and not the

TABLE I: A list of the three exchange processes between the atmosphere and the land surface, the key parameters that control these exchanges and an indication of the variability of these parameters in time and space

Exchange process	Parameter	Pattern	
		Space	Time
Energy			
$R_n = (1 - \alpha)R_s - R_1$	α	+	+
Mass			
H_2O	$r(c, p)$	?	+
CO_2	$r(a, p, m)$	+	+
NO_x	$r(a, m)$	+	?
CH_4	$r(a,m)$	+	?
Momentum			
z_0		+	?

The parameters are albedo (α), exchange 'resistances' that are atmospheric (r_a) or physiological (r_p) or metabolic (r_m) in nature, and z_0, the roughness length.

gradients driving the flux. It is assumed that the latter can be continuously computed within the model.

In Table I the exchange of radiation, the ultimate driving force for all other energy and mass exchanges, is basically parameterized by albedo (α), while for momentum it is roughness length (z_0). For the mass exchanges, the key parameters are the resistances, which may be aerodynamic in nature (r_a, the effectiveness of turbulent transport away from soil and leaf surfaces) or physiological (r_p, the resistance to water vapour flux from within the leaf to the atmosphere via the stomata which are under physiological control) or metabolic (r_m, determined by the rate of some chemical reaction such as is the fixing of CO_2 within the leaf or the production of NO_x within the soil). Where all three resistances are appropriate, they act in series, e.g. Choudhury and Monteith (1988).

As will be discussed in greater detail later, contemporary parameterizations of the surface exchange processes involve many more variables than just those listed in Table I, especially in the case of evapotranspiration. Nevertheless the partitioning of net radiation by most of the world's vegetation for most of the time is determined by r_a and r_p (Jarvis, 1986; Jarvis and McNaughton, 1986). The r_p of a canopy, however spatially aggregated, is not static, and the capturing of the behaviour of stomatal resistance, which is both dynamic within a day and progressive on a scale of tens of days, within GCMs has not been widely achieved.

The simplest yet still realistic GCM requires the global spatial and temporal patterns of albedo, surface roughness and just one mass flux, evapotranspiration (Henderson-Sellers and McGuffie, 1987). With the present level of sophistication and integrity of GCMs, the fluxes of the trace gases CO_2, CH_4, NO_x, etc., are not required. These fluxes are, however, becoming the focus of GBMs and are mentioned for that reason.

Recent advances in parameterizing the land surface exchange processes have concentrated on substantially refining the evapotranspiration (Abramopolous *et al.*, 1988; Wilson *et al.*, 1987a, b; Sellers and Dorman, 1987; Dickinson and Henderson-Sellers, 1988). Evapotranspiration represents the largest component of the energy balance and the mass balance, and has other interactions within the climate system via clouds. The two key feedbacks in the global climate system are those which involve the phase transitions of water (Ramanathan, 1988).

Aerodynamic Roughness (z_0)

The exchanges listed in Table I are considered in order of increasing importance and difficulty to parameterize. The first and simplest transfer is that of momentum from the free atmosphere to the surface. The surface stress can be modelled as a function of windspeed and a drag coefficient (C_D); itself as a function of the surface roughness length (z_0) and a stability measure such as the bulk Richardson number, e.g. Sud *et al.* (1988), or including a displacement height (z_d), e.g. Garratt (1978b). It is recognized that the quantitative understanding of momentum transfer over complex terrain is comparatively much less well understood.

The magnitude of surface roughness has been demonstrated to have a substantial influence on the modelled general circulation of the atmosphere; particularly the contrast in aerodynamic roughness between the oceans and the land surfaces, which have drag coefficients of the order of 1×10^{-3} and 4×10^{-3}, respectively (Carson, 1982). Sud and Smith (1985a) found that the consequences of a dramatic change in (z_0) from 0.45 to 0.0002 m for the desert areas of the globe was a substantial shift in circulation and rainfall patterns over and adjacent to the Sahel region of Africa. Similarly, Sud and Smith (1985b) demonstrated that roughness changes had a large influence on the simulated monsoon rainfalls over India and that for this area, in this model, the surface roughness and the albedo of the surface were equally influential in determining rainfall patterns.

Sud *et al.* (1988), in a global extension of modelling this same roughness change, which is equivalent to making to land surfaces as aerodynamically smooth as the oceans, found an expected increase in boundary layer wind speed but no change in evapotranspiration or sensible heat flux. However, very large changes in rainfall distribution resulted, and Sud *et al.* (1988) propose that global vegetation patterns, which in the absence of topography determine the aerodynamic roughness of the land surface, have an appreciable influence on global rainfall distribution.

Under conditions of horizontal uniformity, the aerodynamic roughness length can be empirically related to the geometry of vegetation and soil (Garratt, 1978a, b, 1980; Hatfield, 1989; Kustas *et al.*, 1989). Roughness lengths have also been derived or estimated for landscape-scaled features such as crops, hedges, etc.

TABLE II: The relationship between various surfaces and their measured aerodynamic roughness (z_0). The values are summarized from many sources

Surface type	$z_0(m)$
Sea	0.0002
Grasslands	0.07–0.10
Crops	0.10–0.10
Woodlands	0.30–0.40
Forests	1.00–2.00

(Pielke, 1984; Weiringa, 1986). The flexibility of some vegetation types, e.g. grasslands, means that z_0 can be a function of windspeed, but this is insignificant on any scale other than local. Illustrating values of z_0 are collected in Table II.

Aerodynamic roughness is entirely determined by the structure of vegetation, i.e. the disposition of the biomass in 3-D space. There are many empirical relationships that relate z_0 to vegetation height but very few that relate z_0 to both height, frontal area and spacing of vegetation. Based on field and wind-tunnel measurements, a useful relationship between aerodynamic roughness and vegetation structure has been determined by Garratt (1977) and used by him to compute the aerodynamic roughness and mean monthly stress for the continent of Australia. The relationship is reproduced here as Figure 3. It should be noted that it is sharply non-linear in a logarithmic plot.

The most general interpretation that can be applied to the relationship in Figure 3 is that z_0 for a vegetation type of fixed height increases as element density increases, reaching a maximum at a value of 0.4. Thereafter the canopy rapidly becomes aerodynamically 'smoother' as vegetation density increases. From Table II it can be seen that the largest potential changes in roughness would result from the conversion of a woodlands to grasslands, or vice versa. This is both ecologically possible and feasible. In the case of deforestation, Dickinson and Henderson-Sellers (1988) employed changes in z_0 from 2.0 to 0.05 m. Far more widespread, but less dramatic changes in roughness can result from changing fire frequencies which can transform boreal and temperate forests or savanna woodlands into sparsely wooded grasslands. This type of surface transformation is quite possible as the result of climate change and has the potential to affect some 30% of the global land surface. Increases in fire frequencies, both natural and anthropogenic, will be a consequence of drying climates. Conversely, increasing rainfall in the warm, dry regions of the globe will substantially increase the size and density of woody vegetation, and consequently z_0, in the deserts and grasslands which together comprise some 40% of the world's vegetation.

The problems of the parameterization of momentum transfer within GCMs are two; how to include orography and how to scale up, i.e. how to areally average the aerodynamic roughness of the mosaic of different surfaces that comprise any land-

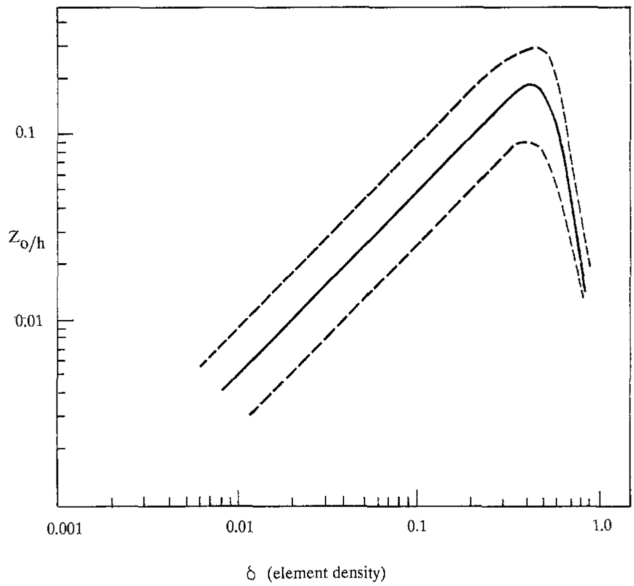

Fig. 3. The non-linear relationship between aerodynamic roughness scaled by height (z_0/h) and the dimensionless vegetation element density δ, where δ is defined as the silhouette area normal to the wind per unit surface area occupied by each element. The relationship is drawn with uncertainty limits. Note that both scales are logarithmic. Redrawn from Garratt (1977).

scape. These problems are currently being addressed by modellers using a mixture of empirical and theoretical solutions (Andre and Blondin, 1986; Carson, 1986 and 1987; Mason, 1988).

Albedo (a)

The exchange of electromagnetic energy that occurs at the Earth's surface is fundamentally, but not completely, parameterized by the (shortwave) albedo of that surface. It is a critical parameter because it represents the amount of radiant energy that is absorbed to drive both biospheric and atmospheric exchanges and dynamics (Henderson-Sellers and Wilson, 1983; Dickinson, 1983). The simplicity of the basic concept of albedo is often lost in its application (Hughes and Henderson-Sellers, 1982). The global or system albedo, which must be accurately simulated by GCMs, is an additive function of the albedos of three components: the vegetated surface, clouds and the atmosphere (dust) of the planetary boundary layer. The latter two have been dealt with earlier and only surface albedo is considered here.

The range of albedo values for various surfaces are provided as Table III. These values are to be appreciated in general terms only because, as it will be demonstrated later, these values are subject to diurnal and seasonal variation and have differing climatic significance depending on latitude of occurrence. It is only the contrasts in Table III that can be emphasized; particularly that between the dark

TABLE III: A tabulation of short-wavelength (150–4000 nm) albedos for various surface types; compiled from several sources

Surface type	Measured albedo
Clouds; cumulus	0.85
Clouds; cirrus	0.35
Snow; ice	0.70–0.90
Sands; dry	0.40–0.50
Sands; wet	0.20–0.25
Grasslands	0.15–0.35
Forests	0.10–0.20
Ocean	0.02–0.07

oceans ($\alpha < 0.10$), clouds and ice ($\alpha > 0.70$) and the terrestrial surfaces ($0.10 < \alpha < 0.50$). The oceans, the most extensive surface feature of the planet, have very low albedos and are responsible for most of the heat absorbed by the planet (Henderson-Sellers and Hughes, 1982). In contrast, the cryospheric surfaces of snow and ice have extremely high albedos but their influence on global albedo is much less than that of the oceans because of their restricted occurrence to high latitudes. In spite of this, the polar regions are critical to the overall energy budget and climate system of the planet (Harvey, 1988).

Within terrestrial surfaces, the bare white sands of deserts have albedos comparable to those of cirrus clouds, while forests have the lowest albedos of all. Equatorial rainforests, with albedos of <0.10, are amongst the darkest, most energy-absorbing terrestrial surfaces of the planet. Simulations of tropical deforestation indicated that the consequent increase in albedo, a doubling from 0.11 to 0.19, had climatic consequences that were only local in significance, and having a negligibly small effect on global scales (Henderson-Sellers and Gornitz, 1984).

The albedo of any surface is determined by the spectral properties of the surface components, the physical arrangement or structure of those components and by the conditions and geometry of illumination (Dickinson, 1983). Thus for any one surface type, albedo will vary on a daily and yearly cycle as a function of solar position and thus illumination geometry. Superimposed on this systematic variation will be the random influences of variations in the ratio of direct to diffuse radiation and the seasonal or episodic variation due to phenological changes in the spectral and structural properties of the vegetation cover. The magnitude and seasonality of these responses on a global scale have, as a consequence of regular satellite observation, only recently been appreciated (Tucker *et al.*, 1986a, b).

One such example of the annual phenological cycle is the data of Brest (1987), Figure 4. Here the albedo for a rangeland site has been separated into visible and near-infrared components, and the divergence of the two during the period of active growth is quite dramatic. This divergence is characteristic of all actively growing vegetation where there is strong absorption by chlorophyll in the visible wave-

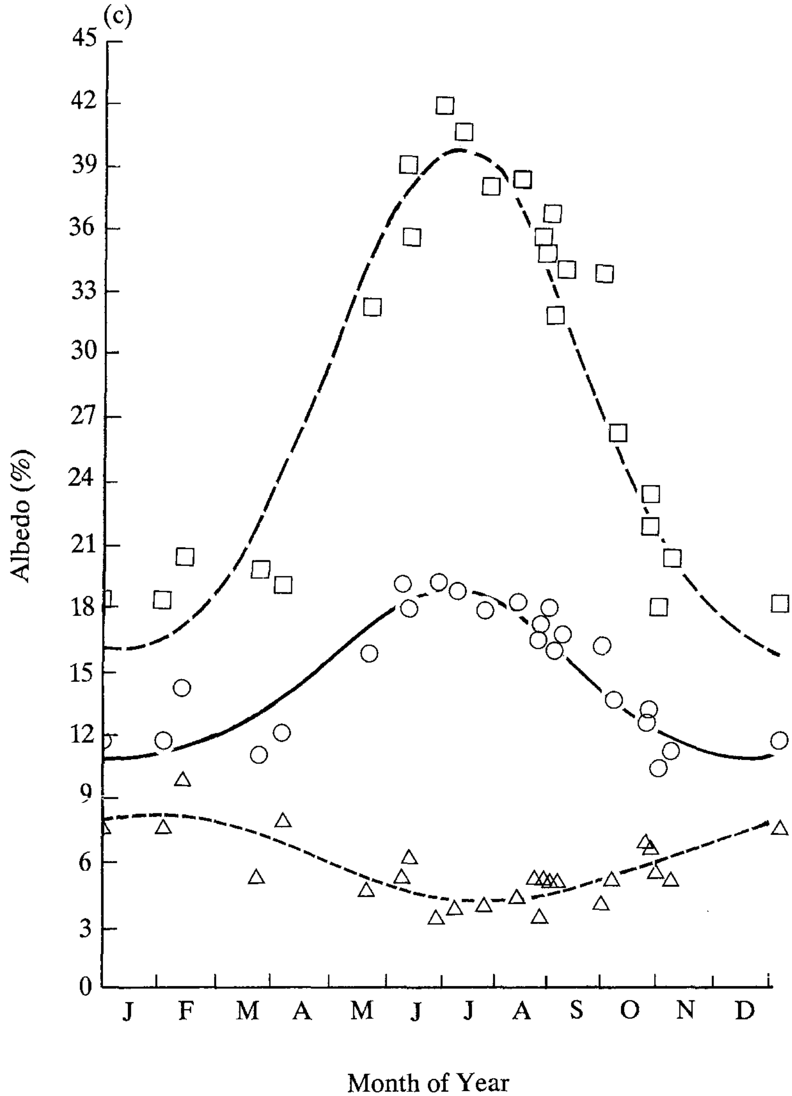

Fig. 4. The seasonal variations in visible and near-infrared reflectance, and their weighted combination as an albedo for a tree-less rangeland site in North America. Redrawn from Brest (1987).

lengths and a high reflection of near-infrared by the cell walls of the expanding leaves. The significance of this divergence in albedo can be assessed by noting that incoming solar radiation is approximately equally partitioned into these two broad wavebands (Dickinson, 1983).

The albedo of terrestrial surfaces can, using satellite data, be observed to show significant variation in time and space at the level of areal aggregation used within GCMs. This can be illustrated by an observed linear contrast across the Sahel of Africa, Figure 5. The transect traces are of bidirectional reflectance (BDR) derived

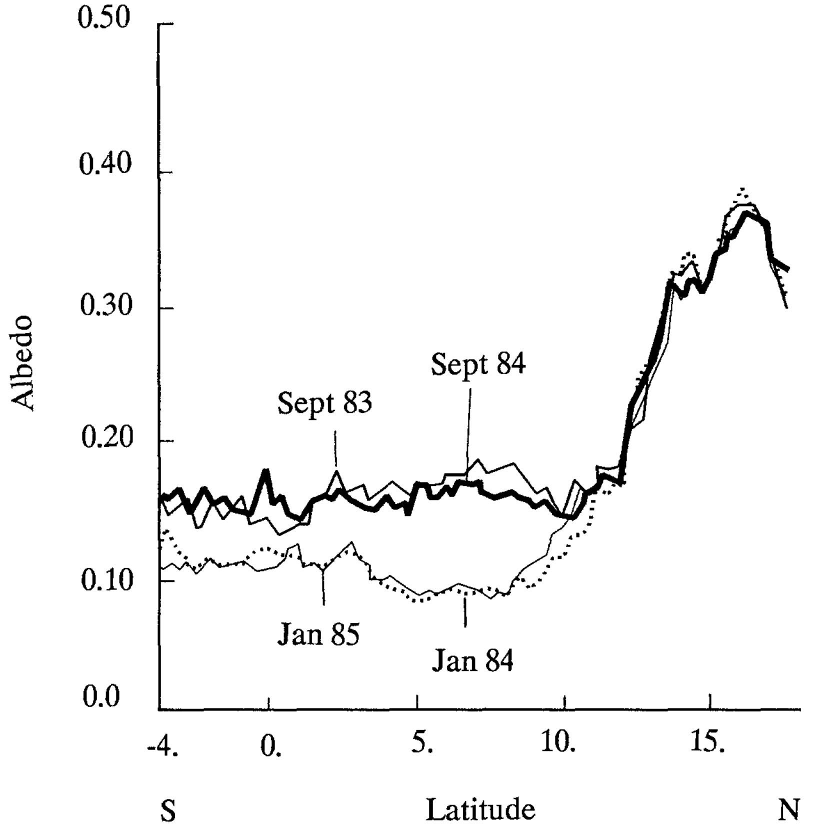

Fig. 5. Seasonal variations of albedo along a N–S transect at longitude 20° E in the African Sahel, where albedo was derived from METEOSAT data. Redrawn from Becker *et al.* (1988).

from METEOSAT data which, while not strictly albedo, adequately illustrates two points. The first is the dramatic spatial contrast in albedo in space; from the low values of 0.15 for the equatorial forests to the very high values of the Sahara desert at latitude 15° N. The second point is the variation of the albedo of the vegetated areas with time of year, between January and September. Here the difference in albedo (0.11–0.16) can be interpreted as a combination of variations in illumination geometry (solar elevation of September vs. January) and phenological response of the vegetation to seasonal rainfall.

There have been few studies that have systematically related albedo to the structure, composition or phenological status of the vegetated surfaces (Ahmad and Lockwood, 1979; Irons *et al.*, 1988). Given that the considerable variations in illumination geometry and conditions are often averaged, only generalizations can be made. The most basic of these is that the greater the depth of vegetation through which the radiation must pass, the greater the chance of absorption of the visible wavelengths and the increase in reflectance of the near-infrared wavelengths. Thus equatorial rainforests which have very high leaf biomass have the lowest albedos, while the sparse desert vegetation covering brighter desert soils has the highest albedo. Even if the canopy is not closed (projected foliage cover, PFC < 1), such as woodlands or tall boreal forests, multiple scattering produces low albedos (Dickinson, 1983). Albedo generally declines with increasing vegetation height (Stanhill, 1970). Typical phenological changes are illustrated in Figure 5.

Because of its critical role in determining the amount of solar radiation absorbed, albedo and albedo changes have been cited as part of many climatic change systems that range in time and space scales from the geologic-global scales of the ice ages to the regional-decade scales of desertification (Gornitz, 1985; Henderson-Sellers and Hughes, 1982; Otterman and Tucker, 1985). Albedo not only provides a fundamental source of variability within the climate system but also reflects climatic change at surface and system levels. This complexity and the difficulty of its incorporation into GCMs is well summarized thus: 'One conclusion ... is that even for (cloud) clear regions there exist differences in our knowledge as to how to model the directional planetary albedo, with the notable deficiency pertaining to land surfaces' (Potter *et al.*, 1988).

This deficiency could be in part alleviated by a systematic albedo measurement program using satellite data as recommended by Henderson-Sellers and Wilson (1983) and achieved by Brest (1987), Brest and Goward (1987) and Gutman (1988). However, for predictive understanding such observations need be closely related to the vegetation characteristics of the surface. The global data base assessed by Matthews (1983) can be substantially improved.

There have been two salient reviews of the sensitivity of modelled climates to changes in surface albedos. Mintz (1984) reviewed 11 simulation experiments examining the sensitivity of various models to changed boundary conditions with the emphasis on surface albedo and soil moisture. To quote from his conclusions: 'All of the experiments show that the atmosphere is sensitive to land-surface evapotranspiration: so that changes in the available soil moisture or changes in the albedo ... produce large changes in the numerically simulated climates.'

Most recently, Rowntree (1988), in a review of GCMs and the relation between vegetation and climate, briefly summarized GCM sensitivity studies relating to surface albedo and moisture. For albedo increases both globally and regionally, the major effects (for both non-interactive and interactive soil moisture) were decreased land evaporation and decreased precipitation over land (see Table 2 of Rowntree, 1988).

The difficulty in assessing the significance of these findings lies in the unreality of many of the changes in surface conditions; the changes are often unrealistically large and imposed independently of other commonly correlated surface conditions, such as albedo and surface soil water content.

Evapotranspiration

The exchange of mass between the vegetated land surface and the atmosphere that

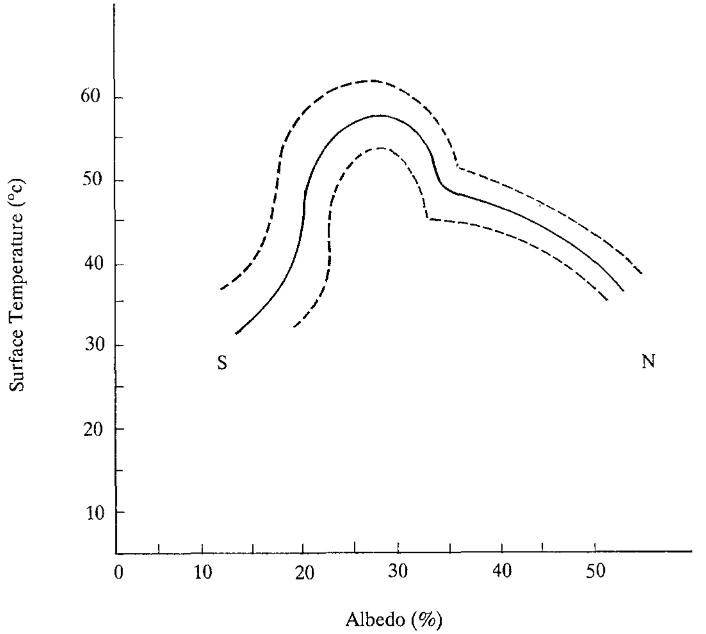

Fig. 6. The relationship between mid-summer surface albedo and surface temperature along a S–N transect in Senegal, July 1979. The data were derived from METEOSAT-6 and are here represented by a mean trend line (continuous) bracketed by the (discontinuous) outer limits. At the southern end of the transect vegetation cover is high and albedo and surface temperatures are low, presumably because evapotranspiration is also high. The northern end of the transect is desert where the vegetation cover is low and albedo and temperature are high. The middle of the transect can be interpreted as being vegetated, therefore having a low albedo, but having limited evapotranspiration, and thus high surface temperatures. Redrawn from Becker *et al.* (1988).

is of greatest significance to meteorological and climatological processes is that of water; collectively called evapotranspiration (E_T).

Evapotranspiration is a critical process because it dominates the partitioning of the net energy balance at the surface. Because evapotranspiration involves the latent heat of vaporization, small changes in the availability of water for evaporation will determine very large changes in the partitioning of the net radiation between evapotranspiration and sensible heat with surface temperatures indicating the balance between these two fluxes. The interaction of albedo and evapotranspiration in determining surface temperatures of vegetated landscapes can be illustrated using satellite data, Figure 6. As with albedo, the evapotranspiration characteristics of landscapes vary quite dramatically over spatial scales that are considerably smaller than the grid cell sizes of GCMs.

The climatological consequence of differences in the evapotranspiration characteristics of landscapes can be evaluated by both observation and simulation. The observational evidence comes from areas wherein large-scale contrasts in evapotranspiration have been created by irrigation. Such an example is the Great Plains of the USA, where it has been established that irrigation has increased rainfall by

10–25% in the rainy months and increased the frequency of hailstorms and tornadoes in the dry months (Nicholson, 1988). Similarly simulation studies using mesoscale models of the atmosphere indicate that spatial variations in surface evapotranspiration are reflected in altered mesoscale circulation patterns (Avissar and Pielke, 1989; Mahfouf *et al.*, 1987; Segal *et al.*, 1988; Yan and Anthes, 1988).

Evapotranspiration is a critical component of the simulated water balance of any point in space. The net balance between the precipitation of water from the atmosphere and evapotranspiration returning that water determines the water balance, i.e., the difference between that stored in the soil or water table and that which runs off. The water balance of the entire land surface is an expression of the spatial patterns of differences between two large variables, precipitation and evapotranspiration.

As a process, evapotranspiration is driven by the energy available for vaporization, the gradient in specific humidity and controlled by the resistances offered by the turbulent transport characteristics of the atmosphere (r_a) acting in series with the collective resistance offered by the stomata of leaves (r_p) (Choudbury and Monteith, 1988; Finnigan and Raupach, 1987). The stomatal resistance operates in a physiological feedback loop between soil water and leaf water content (r_p). Or, if soil water is not limiting, r_p is a function of radiation, temperature and vapour pressure deficit. As a consequence, the stomatal resistance of a canopy can be both dynamic, on a time scale of hours, and progressive, on a time scale of days, as the soil water store is depleted. The relative importance of these resistances vary as a complex function of climate, soils and the physiognomy, dynamics and taxonomy of the vegetation. This complex function of interaction is best regarded as an exercise in economics whereby vegetation pursues a strategy to maximize CO_2 uptake with respect to water loss, or to light intercepted, etc. (Cowan, 1986; Dunin *et al.*, 1989; Verma *et al.*, 1989).

The mechanistic understanding of evapotranspiration is advanced at the level of environmental plant physiology, e.g. Jones (1983), but decreases with increasing scale of spatial aggregation (Brutsaert, 1988). The representation and parameterization of this process within GCMs has proven to be a complex, difficult and critical task (Dickinson, 1984). It is critical because the bulk of the evidence from GCM sensitivity studies indicates that forecast climates are very sensitive to the parameterization of evapotranspiration. Far more so than for, say, albedo (Dickinson, 1988).

The task of parameterizing evapotranspiration for GCMs has been tackled in several ways with differing complexity depending upon the purpose of the modelling exercise. Almost all formulations are of a 'big leaf', that is, of GCM cell size, i.e. 100×100 km up to 500×500 km in size. Within each grid cell the physics and physiology of all of the radiant energy exchange, soil water storage, movement and transport processes are captured and parameterized with varying degrees of complexity. Choudhury and Monteith (1988) set out a relatively simple model of the main component fluxes and their representation as a network of resistances and potentials. Sellers *et al.* (1986), Henderson-Sellers *et al.* (1988) and Wilson *et al.* (1987a, b) equate almost identical models with the structure of a 'big leaf' to provide a guide on how vegetation can be treated in global models given the small consistent descriptive data sets available for terrestrial vegetation. Of the two research groups, the Biosphere-Atmosphere Transfer Scheme (BATS) within the NCAR Community Climate Model, e.g. Wilson *et al.* (1987a, b) and Dickinson and Henderson-Sellers (1988), appears the simpler and more generic model compared to the Simple Biosphere Model (SiB) of Sellers *et al.* (1986) and Sellers and Dorman (1987).

For complete details of the intricacy of the parameterizations, the reader is referred to the original papers cited above. However, for the purposes of this paper, the relationships between vegetation structure and evapotranspiration can be explored by a much simplified 'big leaf' model, Figure 7.

The basic assumption of all 'big leaf' models is that the vegetation is distributed uniformly in space such that all exchange processes need be modelled in the vertical direction only. This assumption greatly simplifies the model formulation and facilitates progress. It is, however, a critical assumption and, in the experience of all landscape ecologists, a drastic simplification. But, at this stage of the development and testing of such models, the significance of this critical assumption cannot be evaluated.

Adopting the symbolism of the NCAR BATS model, the essential structure of a 'big leaf' canopy can be represented by (a) in Figure 7. The canopy is structurally simplified into three lobes which represent the vertical aggregations of three processes, the most important of which is the (RHS) physiological control of water loss (and CO_2 uptake) by stomata. The size of the upper LHS lobe represents the interception of precipitation by the canopy, while the lower indicates the size and seasonal dynamics in projected foliage cover (PFC). Root distributions and dynamics are specified for a minimum of two discrete soil horizons. The upper may contribute directly to E_T and change radiative characteristics as a function of water content, while the lower represents the principal soil water storage to be exploited by roots.

Evapotranspiration can be reasonably simplified by the potential-resistance representation of Figure 7(b). All resistances in the circuit are noted as variable on the time scale of hours and differ greatly in absolute magnitude. It can be appreciated from Figure 7(a) that E_T is usually controlled by either r_a or r_p, given that direct evapotranspiration from the soil is usually small in the long term. The aerodynamic resistance (r_a) will be related to roughness (z_0), and thus to canopy structure, while stomatal resistance (r_p) and its dynamics will be related to the species composition of the canopy, all other environmental influences being equal.

In summary the relationship between the existing model parameterizations of evapotranspiration and vegetation is explicit, but simplified in time and space. The structural aspects of vegetation are essentially captured by the values used for z_0, albedo, PFC, etc.; see particularly Figure 2 of Wilson *et al.* (1987b). The taxonomic

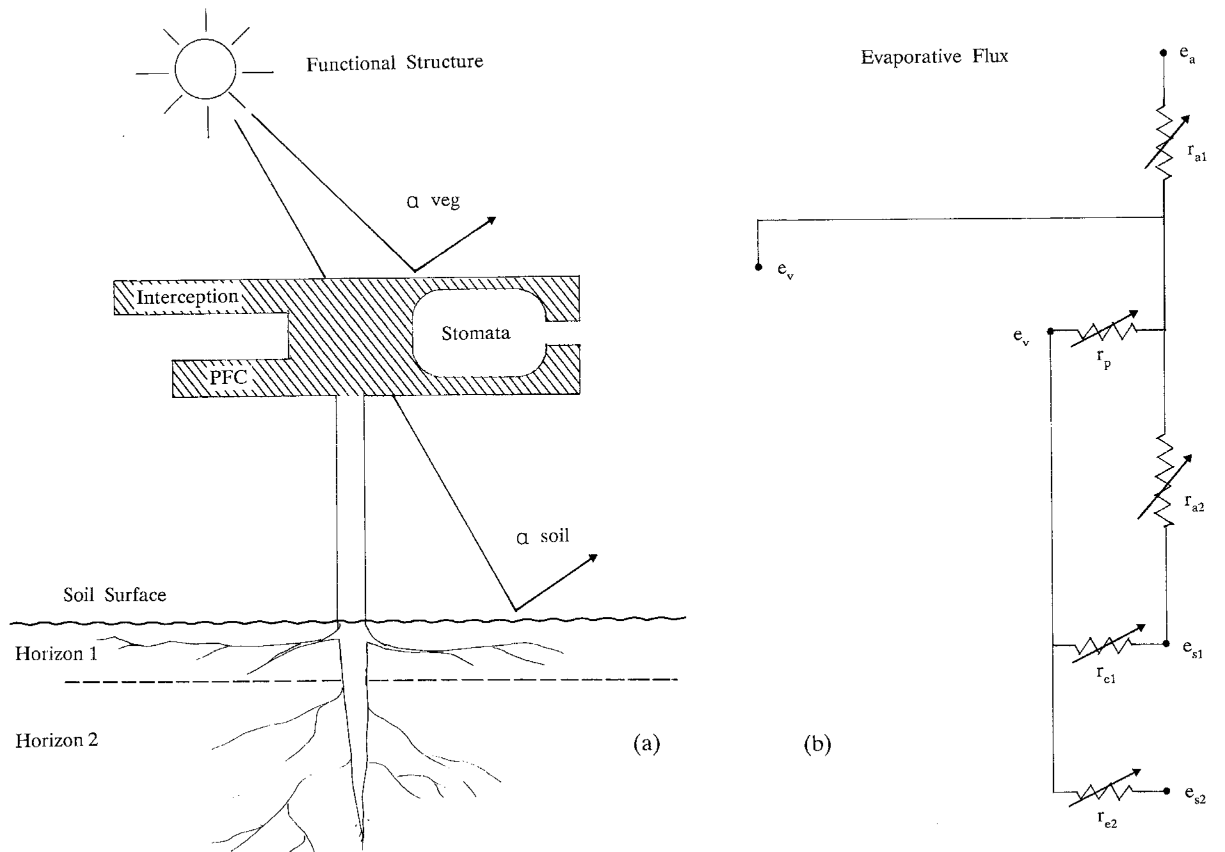

Fig. 7(a). A stylized representation of the functional structure of a 'big leaf' canopy. The real-life vertical distribution of processes and variables is reduced to just one level, i.e. one stomatal resistance (r_p) aggregated for the canopy. The soil water reservoir is partitioned into a minimum of two layers which are differentially exploited by roots. The processes of interception of precipitation and dynamics of projected foliage cover (PFC) are represented by the LHS lobes of the leaf. Incoming radiation is partitioned by a canopy and soil albedo; longwave exchanges are omitted for clarity; redrawn from Wilson *et al.* (1987b). (b). A potential and resistance network representation of evapotranspiration from the canopy (a) above. The fluxes are driven by water vapour pressure differences between the soil (e_{s1}, e_{s2}), the canopy (e_v) and the ambient air (e_a). The resistances are all symbolized as variable on a time scale of hours and vary greatly in absolute magnitude.

composition, the species-specific information, is only captured by the values of stomatal resistance r_p used, and its functional relationship with soil water content, temperature and specific humidity deficit (Gash *et al.*, 1989; Monteith, 1988). There are few data to support this last critical modelling step, except for crops and coniferous forests.

The critical assumption of the 'big leaf' parameterization approach is that of spatial uniformity, i.e. that all modelling need have but one dimension, the vertical (hereafter called the 1-D assumption). The limitation of this assumption is demonstrated in Figure 8. Based on several sources, global vegetation is grouped into eight classes on the basis of the domains they occupy in the canopy height/cover space. A critical divide is drawn where the PFC reaches the value of 1 or greater (closed canopy) because it is only at, or beyond, this point that the exchanges of energy and mass within the canopy can be legitimately modelled in just the vertical dimension. In open or sparse canopies where PFC < 1, the modelling should have at least two dimensions, height and spacing or density. In reality the measure of leaf area index (LAI) is meaningless where canopies are not closed. Some 70% of terrestrial vegetation has open canopies with PFCs < 1 (Figure 8). Almost all of the experimental evidence for the behaviour of closed canopies comes from crops, man-made annual grasslands, with a relatively small and recent contribution from rainforests and boreal forests. It follows then that for GCMs to improve their realism and acceptance, the seriousness of the violation of the 1-D assumptions needs be examined. The level of understanding that now exists for crops and forests also needs to be extended by concerted effort to the other 70% of global vegetation, those with sparse open canopies.

The same modelling and analysis that can be brought to bear on the 1-D assumption can also serve to examine the problems associated with spatial aggregation within GCMs. This problem, more often called sub-grid variability, relates to the difficulty of aggregating and averaging non-linear processes and parameters (Abramopoulos *et al.*, 1988; Avissar and Pielke, 1989; Mahrt, 1987; Wetzel and Chang, 1988) as well as realistic modelling of the processes themselves (Rind, 1988). Given that, because of computational limits, GCM grid cell resolution is unlikely to fall much below 250 × 250 km in the next decade, and that considerable surface soil and vegetation data exist at approximately 100 × 100 km, e.g. Matthews (1983), then the problem of aggregation of mixtures of landscapes is obviously limiting.

Lastly, the task of forecasting future climates with GCMs has generated questions of plant ecologists that are both new and challenging. One such question is whether much of taxonomic detail with which vegetation is traditionally described can be simplified. Vegetation is most commonly grouped using phylogenetic criteria. That is, importance is given to evolutionary history and evolutionary relationships. Obviously what is important in this context is not phylogenetic but functional similarity. Thus one contribution that ecologists can make is to simplify the characterization and description of vegetation to a functional basis; to devise a clas-

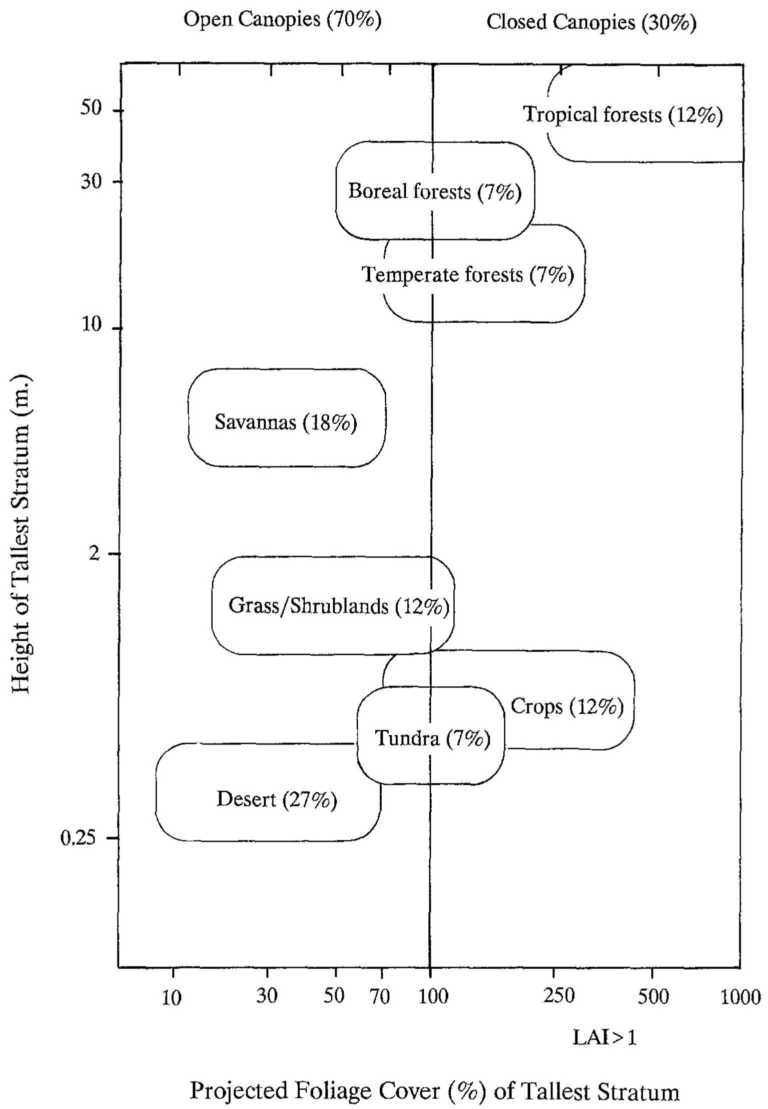

Fig. 8. The domains of world vegetation types distributed in the structural space of height and projected foliage cover (PFC). The point at which canopies become closed (PFC > 100%) is a critical divide. Only in closed canopies can the transfer of radiation, momentum and mass be legitimately modelled in one dimension. In sparse canopies, the bulk of the world's vegetation, modelling of these fluxes must be two- or preferably three-dimensional.

sification of Vegetation Functional Types (VFTs), thereby expediting the progress of forecasting future climates with GCMs. It follows then that if the outputs of GCMs could be interpreted and modelled in terms of VFTs, then the loop between the impact of climate on vegetation and the feedback of those vegetation changes on further climate change would be closed.

Summary and Conclusions

This review has related the energy, mass and momentum exchange characteristics of landscapes to the structure and taxonomy of vegetation.

The first conclusion that emerges from this systematic examination is that the indirect, non-surface feedbacks of clouds and dust appear to be potentially more significant than most foreseeable changes in vegetation. This conclusion cannot be uneqivocal because of the current difficulties in representing clouds and cloudiness within GCMs.

Vegetation structure, rather than species composition, exerts the greatest influence on the exchange of momentum and radiation. However, for the pivotal mass exchange, evapotranspiration, both structure and species composition are tightly interwoven. The representation of real, spatially diverse landscapes as a single leaf is a necessary first approximation that has been rewarding, but the degree of this simplification should not be forgotten. Progress will most likely come from a dynamic parameterization of considerably reduced intricacy with implicit sub-grid variability.

Climate modellers seeking the simplifications they require have started to simplify on a global scale the diverse vegetation literature. Plant ecologists could greatly assist in this process by changing vegetation classification from one based on phylogenetic or geography to a unified one based on function. The definition of Vegetation Functional Types (VFTs) would expedite research on both the impact of, and feedback on, climate change.

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