MODELLING LAND SURFACE–ATMOSPHERE INTERACTIONS AT DIFFERENT SPATIAL SCALES.

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1. INTRODUCTION

Atmospheric General Circulation Models (AGCMs) are valuable tools for investigating the effects large scale perturbations (e.g. increasing greenhouse gases, Mitchell *et al.* [1]; volcanic dust in the atmosphere, Hansen *et al.* [2]; deforestation, Dickinson and Henderson–Sellers [3]; Lean and Warrilow [4]; and desertification, Charney *et al.* [5]) might have on the Earth's climate. Although AGCMs were specifically designed for global scale climate simulations, they are now being used to evaluate climatological and hydrological quantities at or near the land surface and at sub–continental scales (Rind [6]; Sato *et al.* [7]; Wilson *et al.* [8], [9]). They are also being used to provide input data for high resolution climate impact simulations, to predict, for example, the effects of increasing atmospheric CO_2 levels on energy demands, food production, tourism and recreation at regional scales (Cohen and Allsopp [10]).

The surface of the Earth is an essential element of AGCMs. Early models considered the surface as a simple boundary for the atmosphere. However a series of experiments over the last decade have shown that the atmosphere is sensitive to the parameterization of the land surface (Mintz [11]). For instance, changes in the state of the land surface (e.g. albedo, Chervin [12]; roughness length, Sud and Fennessy [13]; soil moisture, Charney [14]; soil texture, Wilson *et al.* [9]) have all been shown to affect the simulation of the Earth's climate by AGCMs. Recently Pitman *et al.* [15] showed that surface climatologies derived from

AGCMs may be remarkably sensitive to slight modifications in the land surface–atmosphere coupling.

In this paper, current approaches and recent advances in modelling the land surface in AGCMs will be discussed. The problems of coarse grid resolution and how to incorporate sub–grid–scale processes are emphasised, with reference to simulations using a variety of different land surface parameterizations with a number of different AGCMs.

2. THE LAND SURFACE COMPONENT OF AGCMS

2.1 INTRODUCTION

The realisation that the atmosphere and land surface are closely linked through a variety of feedback mechanisms has led to a number of models being developed for AGCMs which aim to represent the land surface at a variety of levels of realism. Early attempts considered the surface very simply, typically as a single soil level of globally uniform depth, with a constant moisture holding capacity and with specified roughness length and albedo. Since early AGCMs did not resolve the diurnal cycle in solar radiation, considering the surface in this primitive way was adequate, but as computer power increased, and when the diurnal cycle was resolved, these simple schemes became inappropriate for climate simulations.

In an attempt to provide improved land surface models for climate modelling, Deardorff [16], [17] used the "force–restore" model for soil temperature in two discrete layers, with associated equations for soil moisture and, significantly, a description of vegetation as a single "bulk" layer. Dickinson [18] extended Deardorff's model in a number of ways, in particular by linking it with global data for soil texture and vegetation (from Wilson and

Henderson–Sellers [19] to enable each individual point on the Earth's surface to be represented. Dickinson's model (Dickinson *et al.* [20] has been used in AGCMs successfully and simulates surface energy and moisture fluxes realistically. Sellers *et al.* [21] developed a second land–surface model which also incorporates a canopy parameterization. Although considerably more complex than Dickinson *et al.*'s [20] model (for instance it incorporates a two layer canopy description) it also has been incorporated into AGCMs successfully.

A number of other land surface models have been developed for AGCMs which do not explicitly incorporate a canopy parameterization, but are important as intermediate links between the two current extremes (i.e. Manabe [22]; Sellers *et al.* [21]. Hansen *et al.* [2]) developed a two soil layer model which was analogous to that developed by Deardorff [16]. However, in an attempt to represent canopy transpiration very simply, Hansen *et al.* [2] incorporated a factor (γ) which allows instantaneous transfer of moisture from deep within the soil into the atmosphere where vegetation exists. It represents, conceptually, the root-stem-leaf pathway for moisture. Warrilow *et al.* [23] also developed a soil based model for AGCMs. This model contained four soil layers and represented the canopy by use of a single resistance term. Arguably, Hansen *et al.*'s [2] and Warrilow *et al.*'s [23] models could be considered attempts to represent the soil–plant–atmosphere system by parameterizing the plant–atmosphere part as a simple extension of the soil system. Their models fundamentally miss the hydrological, thermal and momentum peculiarities of vegetation which make its inclusion in AGCMs essential. However, climate simulations using the Hansen *et al.* [2] or Warrilow *et al.* [23] models appear to be successful, and it may be that these models represent valuable approaches when the surface and near surface climatology is not to be considered in detail.

2.2 RECENT ADVANCES IN LAND SURFACE MODELLING

Cogley *et al.* [24] have recently developed a model which attempts to simplify the parameterization of the surface–atmosphere interaction as far as possible, while still retaining an adequate description of land surface processes. Since computational resources are at a premium in AGCM integrations, it is important to limit the level of complexity in land surface models. However, in order to model surface–atmosphere interactions realistically, a reasonable level of complexity has to be accommodated within AGCMs. Figure 1 shows a schematic representation of the processes simulated in the model developed by Cogley *et al.* [24].

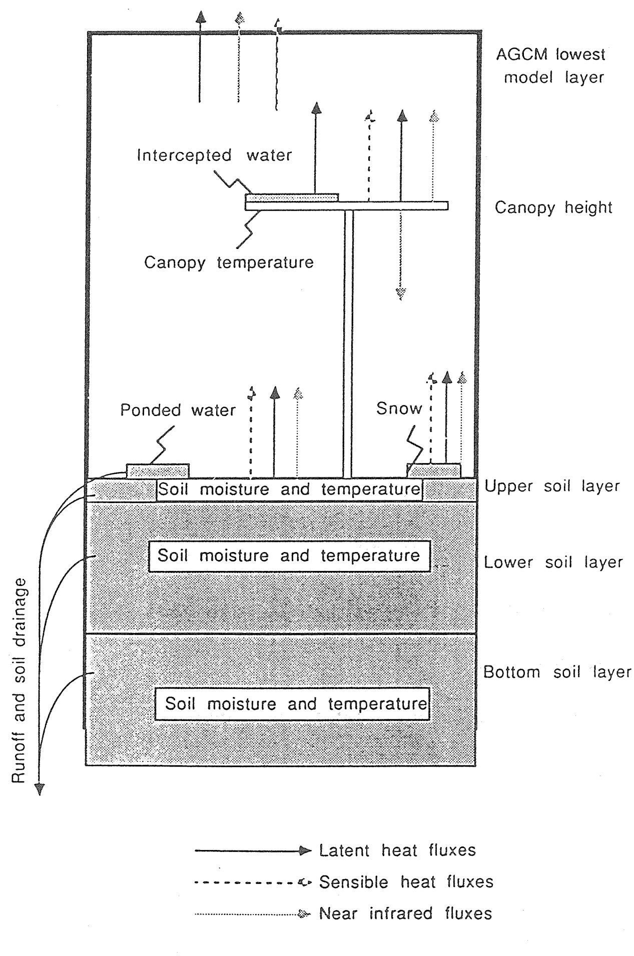


Figure 1 Schematic representation of the land surface model developed by Cogley *et al.* [24], showing the position of heat and moisture stores and the main energy fluxes simulated by the model.

The canopy is represented using a single layer with a seasonally variable albedo, roughness length and leaf area index. In the soil, a moisture holding capacity is specified, snow depth is modelled explicitly and water can pond on the surface if the precipitation rate exceeds the soil infiltration rate. At this level, Cogley *et al.*'s model is conceptually similar to Dickinson *et al.*'s [20] model. However the individual components: how the canopy is parameterized; how snow and soil temperatures are accounted for and how many minor parameters are calculated differ fundamentally. Cogley *et al.*'s [24] model will be described in the subsequent section in order to indicate the level of complexity currently realised in AGCMs.

All those land surface models which incorporate a parameterization of soil and vegetation require information for each land based grid element (soil and vegetation characteristics). Cogley *et al.* [24] use the Wilson and Henderson–Sellers [19] data set to provide soil texture (course, medium, fine) and vegetation type (one of 22 primary vegetation types) at the appropriate model resolution. From soil texture, the moisture holding capacity and soil drainage characteristics are derived. The vegetation type information, coupled to a look up table, provides roughness length, albedo (two stream approach, split at 0.7 μm), leaf area index and some other minor parameters. These data are crucial to how successful the models simulate the surface–atmosphere interactions, while the lack of reliable estimates for these data at an appropriate spatial scale limits how realistically the land surface can be represented in AGCMs.

2.3 THE SOIL MODEL

The soil model is composed of three layers. For the soil temperature, the top soil layer (~0.1 m deep) contains the diurnal temperature wave, the second soil layer (1–2m deep) contains the seasonal temperature wave while the lower soil layer (1–5m deep) contains inter–annual temperature variations. The lowest soil layer may be initialised as frozen to represent permafrost in areas of tundra vegetation.

The soil heat and moisture model is developed from the Philip–deVries theory for heat and water transfer within the soil. The heat and moisture transfer in the soil is not coupled, and vapour transfer in the soil is zero, except at the soil–atmosphere interface. The Philip–DeVries [25], [26] theory is extended to explicitly account for soil ice, snow and glacier ice. In the model there are heat and water sources at the surface, internal sinks in the form of latent heat of melting of soil ice and withdrawal of water by the roots of plants.. Simply boundary conditions exist at the base of the soil where the heat flux is zero and excess water can drain to an imaginary water table at indefinite depth. Heat diffuses down the soil temperature gradient and water follows gradients of gravitational and pressure potential. The three conservation equations which describe this system are:

$$(1) \quad C_v \frac{\partial T}{\partial t} = \nabla G + \mu \, \Gamma$$

$$(2) \quad \rho_w \frac{\partial X_w}{\partial t} = \nabla(R + [1-f_i] \, E) - \Gamma - \theta$$

$$(3) \quad \rho_i \frac{\partial X_i}{\partial t} = \nabla(f_i \, E) + \Gamma$$

where C_v is a volumetric heat capacity (J m^{-3} K^{-1}),
T is the soil temperature (K),
t is time (s),
G is the heat flux density,
μ is the latent heat of fusion (333 kJ kg^{-1}),
Γ is the potential ice production rate (kg m^{-2} s^{-1}),
∇ denotes differentiation with respect to depth, z,
ρ_i is the density of ice = 900 kg m^{-3},
ρ_w is the density of water = 1000 kg m^{-3},
R is the water flux density (kg m^{-2} s^{-1}),
E is the water vapour flux density (kg m^{-2} s^{-1}),
θ is the rate of withdrawal of water due to transpiration (kg m^{-2} s^{-1}),
f_i is the fraction of the total soil moisture store which is frozen,
X_w is the volume fraction of water (m^3 of water / m^3 of soil), and
X_i is the volume fraction of ice (m^3 of ice/ m^3 of soil).

In the case of soil moisture, a moisture holding capacity is calculated according to soil depth and soil texture which is prescribed from Wilson and Henderson–Sellers [19]. These depths for soil moisture need not be identical to the soil depths for temperature. Indeed, the top soil layer for soil moisture is typically only a few mm deep to allow rapid drying during the day, to form a dry crust and thereby inhibit evaporation. The understanding of soil physics with respect to soil temperature and moisture variations with time and depth are reasonably well known. However, spatial variability in soil properties are the severe limiting factor in the level of realism attained in soil based land surface modelling.

Although the basic soil model works well, it is limited by how accurately C_v, G, Γ, R, E and θ can be calculated. The moisture dependency of C_v must be taken into account, but this is relatively straightforward. Each of the other parameters are interdependent and are therefore hard to represent. R is dependent on precipitation intensity (see Section 3), soil moisture levels and soil characteristics. E and θ depend on vapour pressure gradients, surface resistances, and in the case of θ on stomatal resistance. Simulating G implies the need to simulate all surface energy fluxes into, out of and within the ground realistically.

2.4 THE CANOPY MODEL

The parameterization of the canopy in AGCMs is extremely difficult. Typically, a canopy is well ventilated, has a large surface area of leaf in contact with the atmosphere and is hydrologically and thermally dynamic. They are also biologically active, and hence represent important elements in the climate system (for instance by providing a major sink for atmospheric carbon dioxide). Unfortunately, developing simple, large scale models to incorporate canopies in AGCMs has proved difficult. Sellers *et al.* [21] achieved success by using a comparatively complex canopy model based on a large number (around 40) of empirical constants which need to be specified for each vegetation type. Dickinson *et al.* [20] and Cogley *et al.* [24] used a slightly simpler approach, with fewer empirical constants. It is impossible to assess the relative merits of each approach due to a lack of available quantitative data. However, all these models are derived from the basic Penman–Monteith "big–leaf" model (e.g. Monteith [27]). In all three canopy models, a leaf energy balance is solved to calculate the canopy temperature by using a Newton–Raphson iterative technique. This numerical technique seems to work well, but it is computationally expensive and is therefore not ideal for climate modelling purposes.

In order to calculate the canopy temperature, Cogley *et al.* [24] assume that the canopy has no capacity to store heat. This requires the heat flux into the canopy (G_{sc}) to be zero, which is a reasonable if crude way of closing the problem. The canopy temperature (T_c) can be written in terms of an energy balance

$$(4)\quad G_{sc}\,[T_c(t)] = 0 = R^*_c - H_{sc} - \lambda E_{sc}$$

where R^{*}_{c} is the net radiation of the canopy (W m^{-2}),

H_{sc} is the canopy sensible heat flux (W m^{-2}), and

λE_{sc} is the canopy latent heat flux (W m^{-2}).

To solve equation (4) the Newton–Raphson method is used, proceeding iteratively to a solution in which the left side has an acceptably small value. A first guess of T_c is made (T^{0}_{c}), then a Taylor–series expansion of equation (4) gives

$$(5)\quad 0 = G_{sc}(T_c^{\,n+1}) \qquad = G_{sc}(T_c^{\,n} + \sigma) \qquad \approx G_{sc}(T_c^{\,n}) + \sigma G^{*}_{sc} + \ldots$$

which implies, if σ is not too large, that

$$(6)\quad \sigma = -G_{sc}(T_c^{\,n}) \,/\, G^{*}_{sc}$$

and thus

$$(7)\quad T_c^{\,n+1} = T_c^{\,n} - G_{sc}(T_c^{\,n}) \,/\, G^{*}_{sc}$$

The first guess at T_c has been found to be important, and that the most reliable guess is usually the result of the iteration at the previous timestep. Although this implies a cost in terms of computer memory it saves computer time. It should be emphasised that it is essential to take great care in calculating G^{*}_{sc}. Attempts to save computer time by restricting the iteration can lead to unreasonable values of G_{sc} and therefore T_c. It has proved surprisingly difficult to ensure that this iteration for T_c will converge under all circumstances. The shape of the function for the surface drag coefficient and the Richardson number is such that when the lower atmosphere is in transition from a stable to an unstable state, or vice versa (commonly at dawn and dusk) the Newton–Raphson method is very inefficient at finding the desired root. It is essential to have a sophisticated means of a) recognising and b) treating the occurrence of cycles of nonconverging steps. These cycles invariably involve switches from stable to unstable conditions between successive steps of the iteration. To break out of them it is necessary to abandon the Newton–Raphson method for one timestep, and to switch instead to a version of the cruder bisection method. In this method, one chooses one of the two most recent estimates of T_c. Normally the bisection method implies that one would chose their average as the next estimate of T_c, but it has been found that this does not guarantee that the cycle will be broken, and that it is necessary to select a new estimate *at random* within the range defined by the two old ones. Since adopting this method for solving the canopy temperature calculation, the iteration has never failed to converge.

2.6 SUMMARY

There are a variety of other parameters which have to be calculated to permit these types of model to be used in AGCMs. For instance, canopies suck water up through their roots, transfer it through stems and transpire it through stomata on the leaves. Modelling these processes at a level of complexity suitable for AGCMs necessitates considerable simplification. A major problem is in the representation of the stomatal resistance to transpiration. Stomatal resistance is a complex function of air and leaf temperature, light intensity, ambient carbon dioxide levels, water stress etc. but is included in AGCMs in a highly parameterized form, if at all. It is doubtful whether, at the spatial scale of AGCMs, stomatal resistance to transpiration can be incorporated in a more physically realistic way at present.

Although this has been a very brief discussion of how some aspects of the land surface are now being modelled in AGCMs, it hopefully illustrates how complex the problem is. The resolution of climate models are far too coarse to allow many important processes to be modelled explicitly. Hence there is a continuing need to develop sub-grid-scale parameterizations of important land surface processes. An example will be discussed in the following section.

3. SUB-GRID-SCALE HETEROGENEITY

3.1 INTRODUCTION

All processes which exist and are deemed to be important at spatial scales below that of the AGCM must be represented in some way. Typically, parameterizations are developed which attempt to describe the behaviour of the sub-grid-scale system with respect to time (i.e. the system response to diurnal, seasonal and annual climate forcing). Recently the spatial variability of a number of land surface quantities (roughness length, soil properties and precipitation) have been investigated. Here, the importance of accounting for sub-grid-scale precipitation patterns will be discussed.

3.2 PRECIPITATION IN AGCMS: A CASE STUDY

Although models of the land surface models are probably reasonably realistic, the atmospheric quantities output by the AGCM are not always compatible with those required by the land surface models. For instance, virtually all AGCMs currently predict the precipitation rate as a grid element average, i.e. a single value for precipitation is calculated for each grid element at each timestep. Both large scale (e.g. frontal) and small scale (e.g. convective) precipitation is assumed to fall uniformly over the *entire* grid element. In the case of large scale precipitation this may be satisfactory (particularly for models with spatial resolutions of around 300 km x 300 km). However, in the case of small scale precipitation simulated by the AGCM this grid–average assumption leads to a major loss of physical realism.

The evaporation of precipitation intercepted by a canopy occurs rapidly in comparison to the evaporation of water which reaches the soil surface and infiltrates. The low precipitation intensities "predicted" by AGCMs when precipitation is spread out over the entire grid square could therefore lead to fundamentally misleading hydrological simulations because precipitation recycling from the surface to the atmosphere, in any AGCM incorporating a vegetation canopy, will be overestimated at the expense of increasing soil moisture or runoff. To overcome this problem it is crucially important to account for both the *spatial extent* and *intensity* of precipitation in AGCMs which incorporate the new generation of land surface schemes.

Using the Biosphere–Atmosphere–Transfer–Scheme (BATS, Dickinson *et al.* [20]), it is possible to investigate a method of retaining the spatial extent and intensity of precipitation for AGCMs suggested by Warrilow *et al.* [23] and Shuttleworth [28]. Throughout this discussion water fluxes are in units of kg m^{-2} s^{-1} (1 kg m^{-2} s^{-1} = 1 mm s^{-1}).

In the standard version of BATS, the surface runoff rate, R_{surf}, is given by

$$(8) \quad R_{surf} = \beta P_s$$

where P_s is the net flux of water at the surface in kg m^{-2} s^{-1} and β is a depth–weighted ratio of the soil wetness. Note that equation (8) predicts $R_{surf} > 0$ when *any* water falls to the surface. The canopy drip, R_{drip}, is determined by use of a storage capacity (S, in kg m^{-2}). In a timestep, Δt, when the mass of intercepted water per unit leaf area (C, in kg m^{-2}) on the canopy exceeds S, all water in excess of S falls to the surface

$$(9) \quad R_{drip} = (C - S) / \Delta t$$

Following Warrilow *et al.* [23], it is possible to derive an expression for surface runoff (R_{surf}) and, following Shuttleworth [28] for canopy drip (R_{drip}) by assuming that the local precipitation rate (over a fraction, μ, of the grid element) is described by a decaying exponential probability distribution. This leads to values of R_{surf} and R_{drip} given by

$$(10) \quad R_{surf} = P_s \, exp(-\mu \, F_s / P_s)$$

where F_s is the maximum surface infiltration rate, assumed constant over the grid element in kg m^{-2} s^{-1}. For the canopy, the drip rate (R_{drip}) is given by

$$(11) \quad R_{drip} = P_c \, exp(-\mu \, F_c / P_c)$$

where P_c is the precipitation intercepted by the canopy and F_c is the maximum canopy infiltration rate, assumed constant over the grid element, in kg m^{-2} s^{-1}.

The expressions for R_{surf} and R_{drip} in BATS [equations (8) and (9)] which always assume uniform precipitation over the entire grid square have been replaced by expressions (10) and (11). Using the modified version of BATS in a stand–alone mode, and by prescribing atmospheric forcing (including precipitation, air temperature, solar radiation, downwelling near–infrared radiation and wind speed cf. Wilson *et al.* [9], the model was integrated for two years in order to investigate the sensitivity of the land surface to changes in the definition of μ. Results of the second year of integration (these results are independent of initial conditions) for a tropical forest, with μ set to 1.0 (uniform rainfall over the whole grid element), 0.5 and 0.1 are discussed here. The results from the control simulation using the standard version of BATS are also included. Figure 2 shows the potential effects of these prescribed changes in μ.

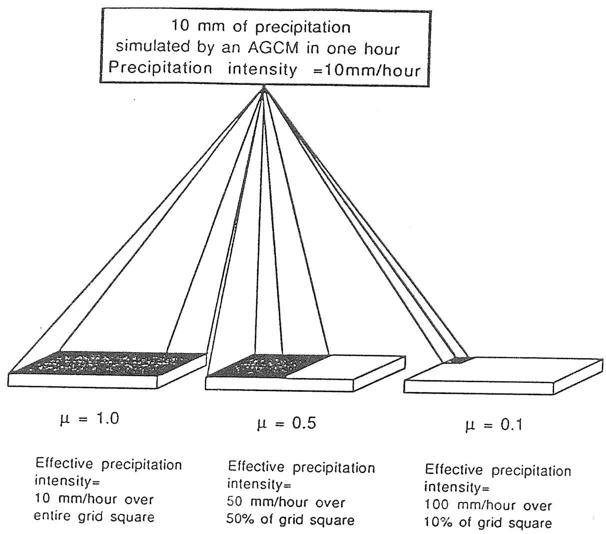


Figure 2 Schematic of the effect of changing the prescribed value of μ on a hypothetical rainfall event. As μ decreases, the effective precipitation intensity increases since the area over which the rainfall occurs is reduced.

When μ = 1.0, and the precipitation falls uniformly over the grid square the precipitation rate remains at 10 mm h^{-1} in this contrived case. However, in reality the precipitation event discussed here should only fall over a fraction of the grid square. When it is prescribed to fall over half the grid square (μ = 0.5) the effective precipitation rate increases from 10 mm h^{-1} to 50 mm h^{-1} and when μ = 0.1, the effective precipitation rate increases to 100 mm h^{-1}.

As the effective precipitation rate increases, the fraction of the incoming precipitation intercepted (and subsequently re–evaporated) is reduced, while the amount of water which reaches the surface increases.

In the following simulations, tropical forest and tropical rainfall are considered because the dense canopy and the high leaf area index typical of this ecotype leads to a high sensitivity to the type (and here the spatial extent) of precipitation simulated by AGCMs. Figure 3 shows four sets of histograms of precipitation, evaporation and runoff for the four simulations. In each case, only (μ) is variable.

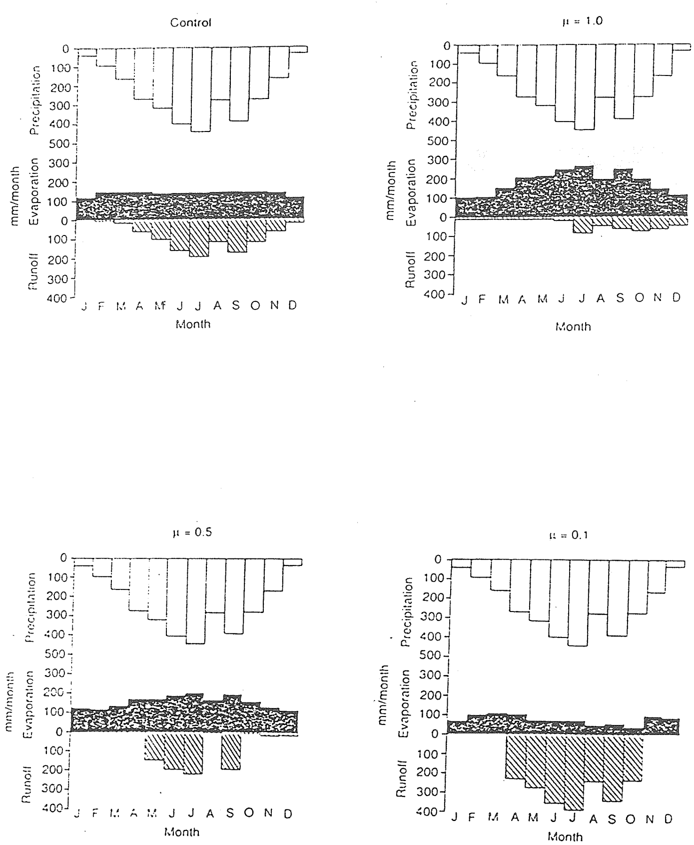


Figure 3 Results from the second year of 4 two–year stand–alone simulations showing the seasonal variation in the precipitation (open downward bars), evaporation rates (solid upward bars) and runoff (hatched downward bars all in mm month^{-1}) for (a) the control simulation (standard BATS without the μ parameterization), (b) BATS plus the μ parameterization with μ = 1.0, (c) μ = 0.5 and (d) with μ = 0.1. All the components of the atmospheric forcing, including precipitation, air temperature, wind speed, solar radiation and downwelling near–infrared radiation are identical in all cases which permits direct comparisons of the four simulations to be made.

The control simulation with "standard BATS" (Figure 3a), shows that the seasonal variation in evaporation is limited, with the precipitation in excess of evaporation during the summer months (JJA) forming runoff. In contrast, BATS incorporating the μ parameterization produces very different simulations. Figure 3b (μ = 1.0) shows much higher evaporation fluxes and very little runoff. When μ = 1.0, the precipitation is evenly distributed over the entire grid square leading to relatively low intensities. The dense tropical forest canopy thus intercepts much of the rainfall, permitting re–evaporation and thus leading to higher monthly evaporation, but much lower runoff rates.

Although the spatial extent of precipitation in Figures 3a and 3b is identical (uniformly distributed over the entire grid element) it is clear that the resulting simulation of evaporation and runoff is quite different because the actual formulation of R_{surf} and R_{drip} are different. The change from the control version of BATS to the version incorporating the μ parameterization leads to a change in the amount of precipitation which is intercepted and evaporated, or which reaches the soil surface. The differences between Figures 3a and 3b show how sensitive the simulation of the partitioning of precipitation between runoff and evaporation is to minor changes in the formulation of the land surface in AGCMs.

As μ is reduced from 1.0, the effective precipitation intensity is increased and interception rates are reduced, leading to lower evaporation and higher runoff. Figure 3c shows the case for $\mu = 0.5$. Evaporation shows some seasonality (though less than in Figure 3b) and high runoff during the summer months. However, Figure 3c shows negligible runoff in April, October and November in contrast to Figure 3a. Finally, Figure 3d shows the simulation for $\mu = 0.1$. Here precipitation is concentrated over 10% of the grid element, hence its effective intensity is relatively high. Figure 3d shows that there is relatively little evaporation through the year except in February, March, April, November and December when evaporation rates are comparable with the *minimum* rates in Figure 3a–3c. The runoff simulated with $\mu = 0.1$ is dramatically higher than in the previous simulations. The high precipitation intensities lead to very low interception and high runoff rates for seven months of the year, with most of the incident precipitation reaching the soil surface and forming runoff. In this case the canopy becomes saturated rapidly in a precipitation event, leading to high throughfall and subsequently large runoff rates.

The quantitative values in the histograms of runoff and evaporation in Figure 3 are of less importance here (since minor changes in the land surface parameterization, or in the prescribed atmospheric forcing would alter magnitudes) than the relative shapes of the seasonal distributions. The differences in the seasonal distributions shown in Figures 3a–3d however, are of considerable importance to potential users of AGCM simulations of quantities at or near the land surface. Figure 3 shows clearly that the surface hydrological climatology is highly sensitive to atmospheric "forcing", in this case precipitation. Simply by changing the area over which precipitation is distributed, the surface climatology can be changed from an evaporation dominated regime (Figure 3b) to one dominated by runoff (Figure 3d).

3.3 SUMMARY

These simulations suggest that it may not be possible to improve the simulation of the land surface climate by improving the new generation of land surface models alone. Advances in land surface modelling must be made in tandem with improvements in the land–surface–atmosphere coupling. The addition of vegetation and soil processes to AGCMs improves physical realism, but it also has the potential to increase the sensitivity of the surface to the atmosphere. If incorporation of these more complete land surface sub–models is considered desirable then the simulation of precipitation, including its sub–grid–scale variability, must be improved.

The distribution of precipitation within grid elements, expressed here through μ, needs to be incorporated into those AGCMs which include vegetation. Its calculation is fairly straightforward (as discussed by Entekhabi and Eagleson [29]) but even specifying precipitation intensities of large scale and small scale precipitation events as different but constant values following Warrilow *et al.* [23] and Shuttleworth [28] would improve the simulation of near surface quantities. Figures 3a–3d showed that using AGCM simulations of near surface variables from models which incorporate a parameterization of vegetation, but do not consider sub–grid–scale variability of precipitation are misleading. Our results underline the fact that using AGCM results for regional scale impact studies of, for example, greenhouse warming is extremely dangerous unless current model limitations are clearly understood.

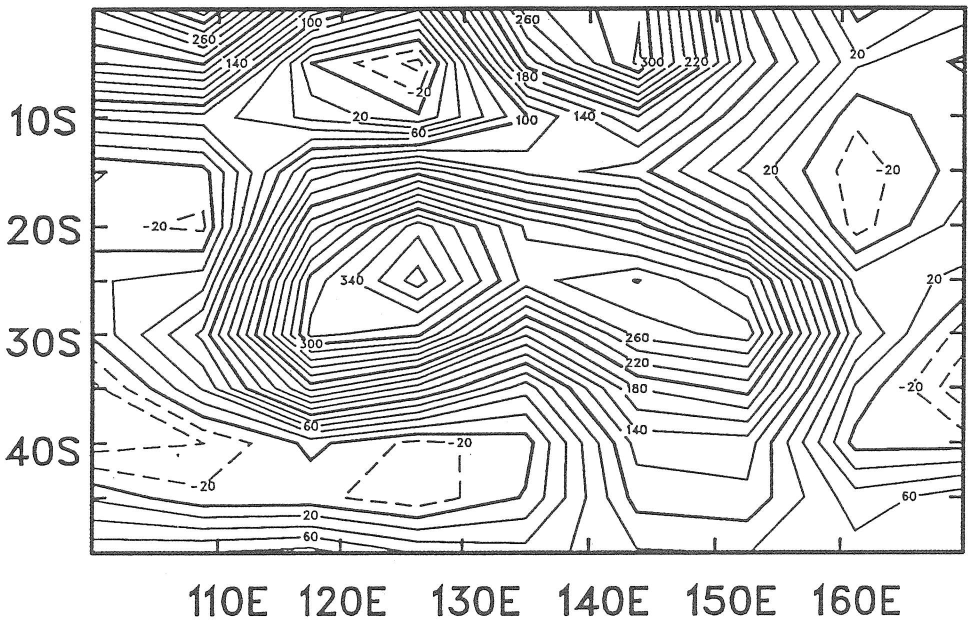
4. RESOLUTION

4.1 INTRODUCTION

Each individual grid square is an AGCM is extremely large. The highest resolution currently used in AGCMs is around 2.8° latitude x 2.8° longitude (UK Meteorological Office model, Slingo [29]). At the other extreme is the Goddard Institute for Space Sciences (GISS) model (Hansen *et al.* [2]) which is commonly integrated with a horizontal resolution of 8° latitude x 10° longitude. The area of each grid "element" is between approximately 10^4 and 10^6 km^2. In the following two sections the effects of resolution on large scale (continental) and regional scale precipitation patterns will be discussed.

4.2 CONTINENTAL SCALE PRECIPITATION

The consequences of the resolution problem in AGCMs is shown by Figure 4. Figure 4a shows the orography field from the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM) at R15 (spectral model rhomboidally truncated at wave number 15 corresponding to approximately 5° latitude x 7.5° longitude) for Australia. The details of the orography field are not important here, but note that Australia appears as a saddle dome, with maximum altitude west of the continental centre. Figure 4b shows the same field for T42 (spectral model triangular truncation at wave number 42, corresponding to approximately 3° x 3°). Although Figure 4b shows that the maximum altitude is still west of centre, the absolute heights are very different (maximum 380m at R15 compared to 500m at T42 resolution), but more significantly, note the representation of the coastal mountains along the east coast which was absent at R15 resolution. In Australia, the Great Dividing Range which parallels the east coast is extremely important in generating orographically induced precipitation. The NCAR CCM at R15 resolution could not simulate these processes since it does not represent the orography.



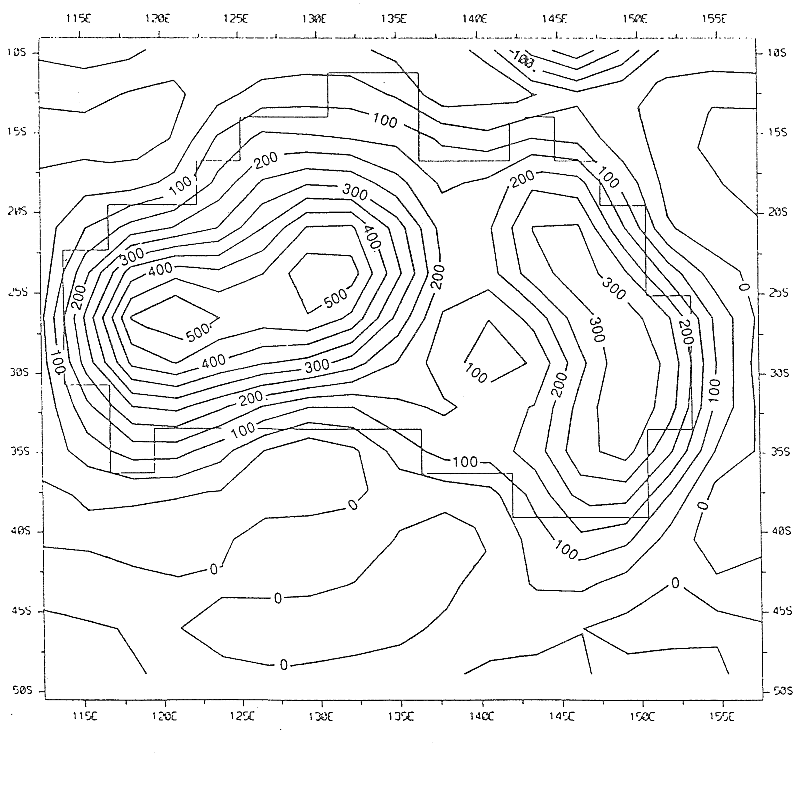
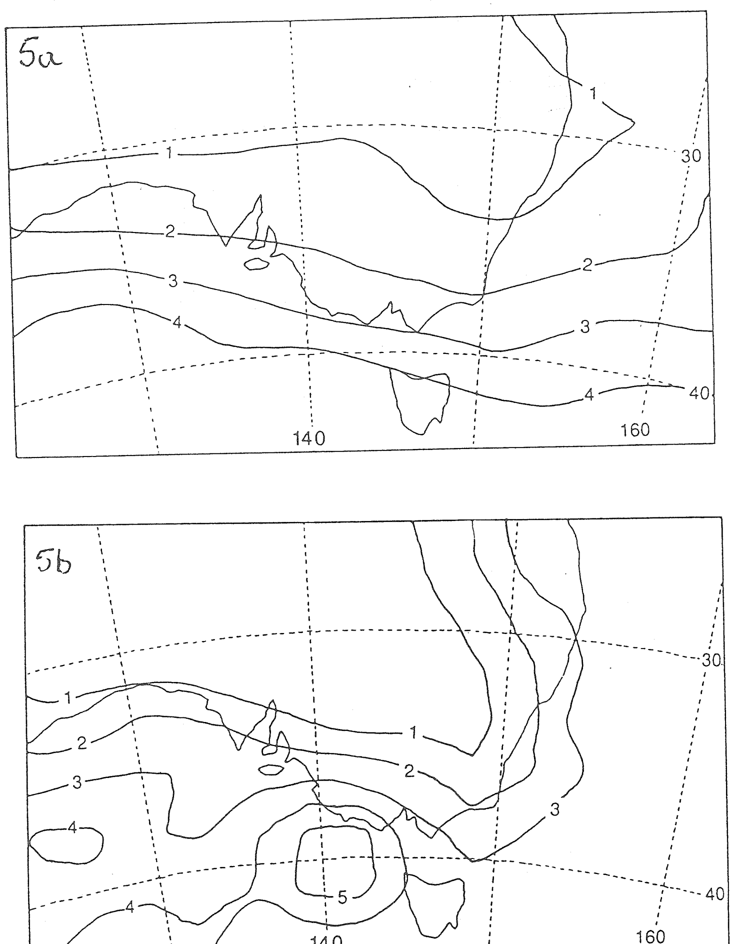


Figure 4 The representation of orography for Australia by the NCAR CCM at a) R15 resolution (contours every 20m) and b) T42 resolution (contours every 50m).

Figure 5 shows the precipitation patterns for a) the NCAR CCM (R15), b) the NCAR CCM (T42) and c) observed (from the Australian Bureau of Natural Resources [31]). The lack of realistic orography in the R15 simulation is clearly shown by the lack of precipitation along the east coast around 30°S. The isohyets in Figure 5a run parallel to latitude over the east coast in sharp contrast with the T42 simulation (Figure 5b) and the observed pattern (Figure 5c).



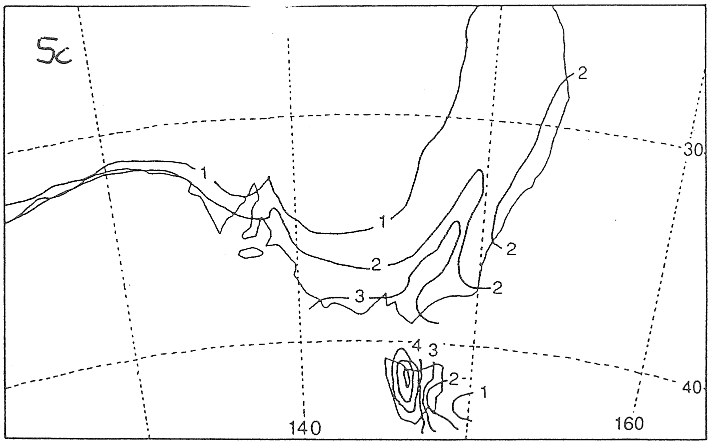


Figure 5 Simulation of July precipitation by the NCAR CCM at a) R15 resolution, b) T42 resolution an c) the observed pattern for July redrawn from the Bureau of Australian Resources [31]). Contours are in mm d^{-1}.

Figure 5a also shows that the NCAR CCM at R15 resolution overestimates precipitation inland from the coastal belt, in particular between 140° - 150°E. At T42 resolution, this problem is largely solved, with a rather more realistic pattern simulated. However, T42 resolution is not a solution to the resolution problem. The NCAR CCM still fails to concentrate precipitation close enough to the east coast, while it appears that the model overestimates precipitation close to the east coast north of 30°S.

Basically, improving resolution improves the orography field since it can be represented more realistically because it does not need to be smoothed as much. By improving the orography field, improvements in the precipitation patterns are also generated. However, increasing resolution is computationally very expensive and is not always possible. It would seem possible that some threshold exists above which some basic resolution dependent fields such as orography can be represented sufficiently well. The switch from R15 to T42 resolutions appears to cross this threshold in terms of orography, at least in so far as important sub–continental scale features become represented (e.g. the Great Dividing Range), leading to an improved simulation of some atmospheric quantities.

4.3 PRECIPITATION

Although increasing model resolution appears to improve the precipitation patterns over Australia, simple changes in resolution do not always produce such clear cut improvements. In this section, precipitation statistics derived from the NCAR CCM at two resolutions for grid boxes as close as possible to the Reserva Florestal Ducke (02° 57'S, 59° 57'W) near Manaus, Amazonas, Brazil will be discussed. This was the site used to observe the tropical forest microclimate (see Shuttleworth [32] and Shuttleworth *et al.* [33] for full details of the site and project). Two simulations at different horizontal resolutions have been performed using the NCAR CCM, in order to investigate the effects of model resolution on the simulation of precipitation.

First, a three year model simulation incorporating BATS, at R15 resolution was performed. The grid point in the model that was closest to the observation site was centred at 2.2°S, 60.0°E. A three year simulation incorporating a bucket type hydrological parameterization (e.g. Manabe [22]) at T42 resolution was also performed. The grid point in the model that was closest to the observation site was centred at 1.4°S, 59.1°E. In these high resolution results the average of four boxes was taken (these four boxes approximately correspond to the area of one R15 grid box) so that the difference in the areas represented by the AGCM results at the different resolutions were as small as possible. Figure 6 shows monthly averaged precipitation amounts from the NCAR CCM at T42 resolution, R15 resolution and observed amounts from Lloyd (unpublished manuscript).

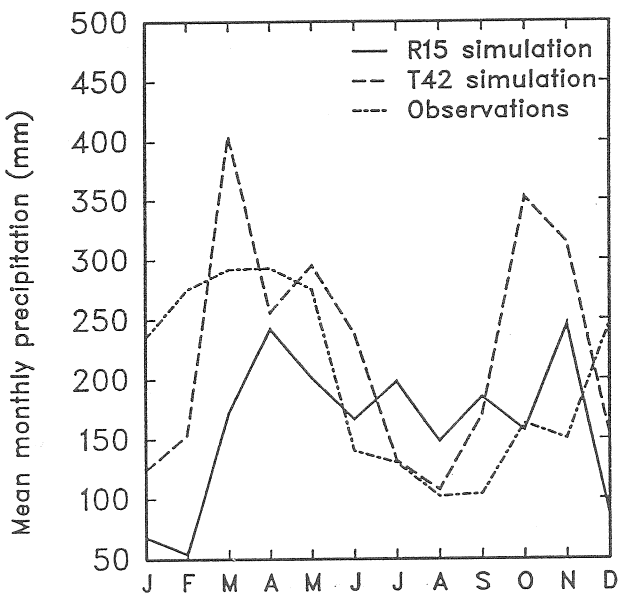


Figure 6 The observed mean monthly precipitation amount for the Reserva Florestal Ducke (02° 57'S, 59° 57'W) near Manaus, Amazonas, Brazil between 1971 and 1983 from Lloyd (unpublished manuscript). The results from a three year simulation with the NCAR CCM at T42 resolution and a three year simulation from the NCAR CCM at R15 resolution are also plotted. All values are expressed in mm month^{-1}.

Figure 6 shows that neither R15 nor T42 resolution produces particularly good simulations of Amazonian precipitation. At T42 resolution, the CCM overestimates precipitation early in the year, and overestimates it in September, October and November. The NCAR CCM at R15 resolution underestimates precipitation for the first five months and overestimates it in July, August and December.

Qualitatively, the T42 simulation is probably better than that generated at R15 resolution. Generally the model simulates the observed pattern quite well, although the October and November simulations are very poor. The NCAR CCM at R15 resolution misses the seasonal cycle almost completely. However, neither simulation is particularly good, hence although switching from R15 to T42 improves the spatial distribution of precipitation for Australia (due to improvements in the orography field) there are no similarly dramatic improvements in Amazonian precipitation amounts. It should be noted that, since the simulation at R15 incorporated BATS, while the T42 simulation did not, these results are not directly compatible. It is likely that improving resolution is important, but maintaining a realistic approach to modelling the land surface is also essential.

5. SUMMARY

The latest land surface schemes now incorporate 2-3 soil layers, 1-2 canopy layers and a series of parameterizations to represent those elements in the soil and canopy which are considered important. Land surface models are now close to reaching the limits of complexity possible within AGCMs for two reasons. First, computer time is limited and as the models improve it becomes harder to argue for further effort to be expended in this area. More critical are the lack of data. Those land surface models discussed above lack data for model development, for initialisation and for validation. There simply is not the data base to permit much more model development in terms of increasing complexity.

It can be argued, however, that extra complexity is unnecessary. Land surface models are becoming reasonably good considering the spatial scales to which they are applied. Section 3 of this paper illustrates that expending more effort on the land surface is pointless until the land surface-atmosphere coupling is improved. The sensitivity of the land surface hydrology to minor changes in the parameterization and to changes in the spatial distribution of precipitation is of considerable concern. The results discussed in Section 3 might invalidate much of the previous work on surface responses to land surface perturbations since the system is so sensitive. However, more work is needed within the AGCM environment before these results can be confirmed.

It seems that before land surface models are extended beyond their present levels of complexity, the representation of sub-grid-scale processes and resolution need to be improved as far as possible. However good the land surface model is, if the orography field is so smoothed that it is unrealistic, the AGCM can never simulate sub-continental scale climate patterns, or be used for climate change prediction.

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