CHAPTER 8: ENSEMBLE GCM SIMULATION OF SOUTH-EAST ASIAN DEFORESTATION: A DESCRIPTION OF THE EXPERIMENTAL DESIGN

8.1: Introduction

In this chapter, the general circulation model experiments run to test the sensitivity of the Southeast Asian climate and regional atmospheric circulation to deforestation are described and justified. The model used is the Unified Model, described in Chapter 6, in atmosphere-only mode with fixed sea surface temperature specified from climatology. The experiment is an ensemble experiment to overcome dependence on initial conditions with ten cases for each experiment defined by atmospheric initial conditions drawn from the European Centre for Medium-Range Weather Forecasts (ECMWF) observational data for the appropriate start-date. The following sections justify this experimental design.

The forcing for the deforestation experiments is single region, i.e. land surface characteristics are altered for the Southeast Asian region alone. A three-region deforestation scheme, as used in previous GCM experiments (see Chapter 5), could give rise to a misleading answer in the estimation of the actual contribution of each deforested region. Instead, a single-region deforestation GCM approach is more appropriate in this study enabling a clear definition of the contribution of Southeast Asia deforestation.

8.2: Simulation Procedure

8.2.1: Choice of model configuration

As noted in the introduction to this study, a subsidiary aim of the project was to assess the use of a workstation (specifically, a DEC Alpha) as a platform for undertaking general circulation model experiments. The main effect of the limited computing power of this platform on the experimental design is that the Unified Model could only be run in atmosphere-only mode. In this mode, for example, a single two-month simulation took around three days of real time to run. The sea surface temperature field was prescribed from climatology with the surface temperature field forced to return to the climatological mean for the appropriate time of the year every five days. This is similar to the kind of experiment run on more powerful platforms some five years ago [e.g. Mylne and Rowntree (1992) and Lean and Rowntree (1993)] so cannot be considered too great a limitation.

In terms of the experimental results, it can be expected that ocean-atmosphere interaction will be constrained and the effects of deforestation on the local temperature field will be underestimated as the sea surface temperature cannot respond to the change in the overlying air. Southeast Asia is strongly influenced by the surrounding ocean so this factor cannot be neglected. However, the short nature of the experimental runs (two months, see next section) means that the neglected ocean response would not be considerable over that period. Regardless, it should be borne in mind that the response of the local atmosphere and atmospheric circulation may not be complete without the ocean reaction. On the positive side, the atmosphere only formulation does enable a clearer assessment of the role of deforestation in influencing the overlying atmosphere without the complicating factor of ocean feedback effects.

8.2.2: The ensemble approach

There are many uncertainties and errors (both random and systematic) that limit the predictibility of any GCM used to study weather and climate on regional scales. Understanding the fundamental limit to predictibility is essential to the use and interpretation of any model. As indicated by Anthes *et al.* (1989), there are two types of model error, or uncertainty, which limit predictibility. The two types are: (i) errors in the model's numerical and physical approximations; and (ii) errors arising from uncertainty in the model's initial conditions. Even with a perfect model with no errors in the physics and calculations, there still remains an inherent uncertainty in any one simulation due to errors in the initial conditions. As described by Lorenz (1969), the nonlinearity of atmospheric motions and the presence of various instabilities in the atmosphere also limit the predictibility of atmospheric processes. This is another source of model error.

A promising way to deal with this problem in studying regional-scale climatology using a GCM is to undertake an ensemble approach. The ensemble approach may reduce the noise in the model results and can potentially improve predictability. A model with good climatological performance will produce an ensemble of simulations whose structure and statistical behaviour are similar to those of the atmosphere and climate. The ensemble approach, therefore, is the key for this study. The strategy for using GCMs to study climatological problems of the type just mentioned is somewhat different from that using GCMs as normally used to study global climate. Rather than integrating the GCM for many months or years to obtain long term statistics, a set, ensemble, of short-range (i.e., 60 days) simulations is undertaken using initial conditions for different times. The averages and other statistics computed from this ensemble would define the climate. Although large uncertainties might exist with any one simulation because of the small-scale nature of the phenomena resolved and the limits to predictibility, the statistics of the ensemble could be meaningful in understanding regional climate. The strategy of running GCMs over the ensemble of cases would be very useful since the errors of any single simulation might be large, but if they are random, they will cancel over a large number of cases so that the model will have little or no bias errors. The use of short-range simulations for an ensemble of cases can also be an advantage if computer power is limited. For example, a less powerful computer (e.g. workstation) could be utilised more effectively to run a number of short integrations. Currently, an integration for many months or years to obtain long-term statistics is only effectively run with an expensive supercomputer.

In this study, ten 60-day simulations each for the "unperturbed" (control) and for the "perturbed" (deforestation) were performed with initial conditions drawn from the 1986/87 to 1995/96 winter monsoons. The simulated period is from the first of December to the end of January. Similarly, another ten 60-day control and deforestation simulations are also undertaken to represent the 1986 to 1995 summer monsoons, simulated from the first day of June to the end of July. For each simulation, the initial boundary conditions were interpolated from observational analyses from the European Centre for Medium-Range Weather Forecasts

(ECMWF). The initial data are valid at 00GMT on the first day of December and June, the starting date and time of every experimental run. Details of the perturbation procedure will be described in the following section. In order to reduce the spin-up effect, as mentioned earlier, from the total 60-day period integrated, only diagnostic outputs after the 30^{th} day were saved to serve as a member of the ensemble.

It should be noted that the use of initial conditions for a particular date does not mean that an attempt is being made to simulate conditions during that specific season. The oceanic boundary conditions are fixed at the long-term climatology values, as noted, and only atmospheric initial conditions are used. The use of actual data is simply a convenient way of perturbing the initial conditions to limit the problem of dependence on these data.

****[RESERVED SPACE FOR A COMPLETE DESCRIPTION OF "THE ENSEMBLE APPROACH]

8.2.3: Model spin-up and simulation duration

The results at the beginning of any integration period of a GCM are largely influenced by the model spin-up as it reaches initial equilibrium. To reduce the spin-up effect, model integration is usually begun at a certain period ahead of the intended starting time at which the output will be saved and used for forecasting or diagnostic purposes.

The spin-up time for a global GCM is usually defined as the integration time required for globally-averaged precipitation to be approximately equal to globally-averaged evaporation. To determine the spin-up time of the GCM, the evolution of both precipitation and evaporation time series from the start of integration should be noted. During the spin-up period, both precipitation and evaporation time series will be out of phase and, therefore, not in good correlation although both are generally growing during this period. At the end of spin-up time, however, both precipitation and evaporation time series curves start to show synchronised characteristics and tend to equalise even though not completely in balance. Visually, both time series plots will show better correlation after the completion of the spin-up phase and throughout the rest of integration period.

In order to determine the spin-up time of the model used in this study, four 90-day control simulations were undertaken, two for the winter months (December to February) and another two for the summer months (June to August). Figure 8.1 and 8.2 show time series plots of precipitation and surface evaporation [Panels (a) and (b)] and their difference [i.e. Precipitation - Evaporation in Panels (c) and (d)] for the winter simulations and the summer simulations respectively. Using the criterion mentioned above, it could be seen that the model requires at least ten days for the spin-up period. The "difference" time series suggest that the simulations are certainly stable after about a 30 days integration period. Therefore, it was decided, rather conservatively, to allow 30 days of spin-up for all the simulations undertaken in this study.

The diagnostic output was saved for the months of January and July to represent the winter and summer monsoon, respectively. It is important to note that the integration period, inclusive of the spin-up, was set by computational considerations as noted earlier. Of course, short simulations of a GCM will not reach equilibrium for all subsystems of the climate system. However, important climatic domains which are directly relevant to this study such as the "free atmosphere" and "boundary layer", as well as "vegetation" can achieve equilibrium within the 30 days spin-up period chosen for this study. As noted by McGuffie and Henderson-Sellers (1997), the free atmosphere needs only 11 days to re-equilibriate following perturbation, the boundary layer requires only 24 hours to do so, and vegetation 11 days.

8.2.4: Other settings

The UK Meteorological Office Unified Model (Version 4.0) employed in this study in atmospheric global mode uses the primitive equations to describe and predict the atmospheric variables at 19 levels between the surface and the top of the atmosphere (see Chapter 6). The horizontal resolution at each level is 2.5^o x 3.75^o. In the vertical, the 19 sigma layers (sigma is equal to the pressure divided by its surface value) are spaced unevenly to allow for enhanced resolution near the surface and in the upper troposphere. For every experiment undertaken in this study, a 30-minute timestep was used to correspond with the above horizontal and vertical resolution and ensure stability. All other settings used are as the standard experiments specified by the UK Meteorological Office for climate simulations (Met. Office, 1996).

8.3: The Control and Perturbation Experiments

Geographical variations of land surface and soil types are represented within the model. The vegetation and soil types are based on Wilson and Henderson-Sellers (1985) 1^o x 1^o resolution global archive of land cover and soils data interpolated onto the 2.5^o x 3.75^o resolution by Warrilow and Buckley (1989). For the UM, Jones (1995) describes details of derivation from the Wilson and Henderson-Sellers data sets. The vegetation and soil parameters required at the start of a simulation are as listed below:

a. Vegetation Parameters:

i. Root depth
ii. Snow-free surface albedo
iii. Surface resistance to evaporation
iv. Roughness length
v. Surface (canopy) capacity
vi. Vegetation fraction
vii. Infiltration factor
viii. Deep snow surface albedo

b. Soil Parameters:

i. Soil-moisture concentration at wilting point
ii. Soil moisture concentration at critical point
iii. Soil moisture concentration at field capacity
iv. Soil moisture concentration at saturation
v. Eagleson's exponent in soil-conductivity relationship
vi. Thermal conductivity
vii. Saturated soil conductivity
viii. Thermal capacity (volumetric heat capacity)
ix. Saturated soil water suction.

In the control simulations, the vegetation parameters and soil parameters are specified globally, as detailed by Jones (1995), and input to the model without any change to their values. Figure 8.3 to 8.9 show part of the globe covering South and Southeast Asia for each of the vegetation parameters involved. For the area shown in the figures, each grid-box over the land area is given with a value of each vegetation parameter that is altered. The top panel (a) of each figure gives the values for the control experiments.

In order to simulate the gross effects of total removal of the Southeast Asian forest, all the global boundary conditions are kept constant except over Southeast Asia where the vegetation parameter values are changed to correspond to deforestation conditions. The locations of the area where deforestation is simulated are also represented graphically in Figure 8.3 to 8.9 in the lower panel (b). The inner box in Panel (b) of each figure delineates the study area, covering most of the Southeast Asia with the 38 grid-boxes over land areas are deforested. These 38 grid boxes, therefore, define deforested region which is referred in the next chapters. In all cases, the ecotype is changed from tropical moist forest to scrub grassland. For each of those figures, the grid-boxes over land areas within the inner box drawn in Panel (b) are given with the parameter values for deforestation conditions.

The character of the landscape after deforestation in Southeast Asia is, actually, difficult to specify. Previous experiments (cf. Chapters 4 and 5) have used a degraded scrub-grassland, as selected for this study, but it must be noted that this is a course approximation. The actual land cover in the tropical deforested regions is composed of a mosaic of crops, bare soil, grassland, and secondary vegetation of various ages. The cultural and economic incentives to clear forest depend on location, leading to different forest replacement in different regions. For example, cleared land in Southeast Asia is commonly used for swidden agriculture, in contrast to deforested areas of the Amazonia where pasture is a more common replacement following deforestation (Giambelluca *et al*., 1996). Parameter values for a variety of deforested areas of the Amazon and Southeast Asia including secondary vegetation at various stages of development, therefore, must be different from each other. At the time of this study, there is no extensive measurement to represent more realistically the vegetation parameter values for deforestation within the GCM grids for Southeast Asia. The latest parameter values for tropical deforestation that are available at this time are from measurements taken during the recent Anglo-Brazilian Amazonian Climate Observation Study (ABRACOS). These new data were used by Lean and Rowntree (1997) in their most recent deforestation experiment for the Amazonia. In view of the nature of the current sensitivity study, when all the tropical forest in Southeast Asia was replaced by tropical grassland, it is acceptable that the same perturbation values used by Lean and Rowntree (1993) are used and this facilitates comparison of experimental results. Action, however, is taken in order to avoid inappropriate conversion from an extreme land-surface property to a less extreme grassland in some of the grid-boxes. This is done by maintaining parameter values which are already exceeding the grassland property in certain grid-boxes (such as for urban or sandy areas), keeping them identical to the control values.

Table 8.1: Deforestation condition of the vegetation parameters used in this study.

Vegetation parameter	New value
Root depth (m)	0.619
Snow-free albedo	0.188
Surface resistance to evaporation(s m^{-1})	82.2
Surface roughness length (m)	0.04
Surface (canopy) capacity (mm)	0.633
Vegetation fraction	0.84
Infiltration factor	1.93
Deep snow surface albedo	No change (same as the control)

For this study, relevant values of each deforested vegetation parameter in Figure 8.3 to 8.9, as given by the points in the inner boxes of the Panel (b), are also listed in Table 8.1. According to Lean and Rowntree (1993), these values are based on previous measurements and studies reported in the literature. For the deforested experiment, the vegetation parameters were derived from a number of sources: root depth from Eagleson (1970), Thompson *et al*. (1981) and Halldin *et al*. (1984); albedos from Wilson and Henderson-Sellers (1985); surface resistance from Shuttleworth *et al*. (1984) and Monteith (1976); roughness length from Brutsaert (1982), Thompson *et al*. (1981) and Eagleson (1970); canopy capacities from Dolman (Lean and Rowntree, personal communication); and the infiltration factor from Warrilow *et al.* (1986).

In summary, comparing the deforested with the control simulation (cf. Figures 8.3 to 8.9): root depth (i.e., the average depth of soil from which moisture is available to plant roots) is reduced; albedo is increased as grassland reflects more solar radiation than forests which are very efficient absorbers and scatterer of short-wave radiation; surface resistance to evaporation is less for grassland under freely available soil moisture; roughness length is reduced as forest offers a significant resistance to the wind in the lower layers of the atmosphere; canopy capacity is reduced for grassland compared to the forest when the retention of water become less on smaller leaves of grass or crop trees; vegetation fraction, in reality, should become less as forest is converted to grassland; and the infiltration factor is larger for forests as tree-root systems and forest litter enhance the infiltration rate.

All soil characteristics are kept similar or constant for both the control and deforestation experiments, except for infiltration capacity which is dependent on the infiltration factor (cf. Section 7.3, Chapter 7). Hence, we assume that the soil properties are the same for both the control and deforestation cases. In reality, however, this is not the case since changes in vegetation parameters should be accompanied by some changes also in the soil properties. Nobre *et al*. (1991) and Dickinson and Henderson-Sellers (1988) treated deforestation scenarios by assuming that soil texture becomes finer. As the texture becomes finer, two counterbalancing mechanisms operate (Warrilow *et al*., 1986): the field capacity increases and so allows more water to remain in the soil, whereas the saturated conductivity decreases and so promotes more surface runoff. However, at the time of this study, there is not enough data available from point measurement of soil parameters to derive similarly changed values in Southeast Asia. At present, only a few point measurements of soil parameters for tropical moist forests are available from the ABRACOS field project. Even Lean and Rowntree (1997) in their recent deforestation experiment for the Amazonia did not utilise the soil deforestation data from the ABRACOS. They caution in their report that it was inappropriate to assume the values obtained from the ABRACOS field campaigns from only a few point measurements as representative of the whole Amazon basin. The deficiency in soil specification, however, is said not to affect the model simulation too much since the variations in soils are apparent on much smaller scales than vegetation characteristics (Lean and Rowntree, 1997).

In both the control and perturbation experiments, all the parameters are derived at the start of a model simulation and are then assumed to be fixed throughout the run; thus diurnal or seasonal variations in the parameters are ignored.

8.4: Method of Diagnostic Analysis

8.4.1: Process analysis and large-scale dynamics

The analysis methods focus attention on the processes underlying any local climate change and the role of large-scale dynamics associated with the winter and summer monsoons of Southeast Asia, with particular respect to the hypotheses stated in Chapter 5. Analysis of the grid-element-scale to regional-scale impacts is undertaken in two ways: firstly, with reference to spatially-averaged results from the region that encompasses the deforested region and study area; secondly, with reference to the spatial grid-box results for the area of South and Southeast Asia combined (70^oE to150^oE and 20^oS to 40^oN).

The standard Student's *t*-test (paired samples) is used to access statistical significance assuming independence of values at each grid point and with mean values from the ensemble taken as independent samples. The test is a parametric test of the null hypothesis that two univariate random variable (i.e. control and deforestation) have equal means. Under the null hypothesis, the test statistic has a *t* distribution. The hypothesis testing applied here is for the mean of a population of differences using "paired samples *t*-test" if the observed differences are independent of one another. Applied to this study, the control and deforestation output generates pairs of samples that are independent of each other. Hence, the *t*-test is used to test the difference of the means between the pair of samples. The null hypothesis is that the control and perturbation (deforestation) would have the same mean of the variable concerned, or that the mean difference is zero, as the ensemble control and deforestation results in this study produce pairs of samples that are independent of each other. Our main interest in this hypothesis testing, therefore, is in the difference between pairs with the assumption that the differences must be approximately normally distributed. A rejection of this hypothesis at a level of 5% means that there is only a 5% probability that the difference in mean is a result of chance, or that there is a 95% probability that the difference is real. In this study, statistical significance at the 5% level is used when testing the changes following deforestation.

In most cases, the diagnostic parameters used are self-explanatory and are introduced in the next two chapters. Here, however, we cover in greater detail the main atmospheric circulation diagnostic variables that are used. It is essential that the divergence and exact stream flow of the atmosphere over the study area are specified as accurately as possible. An effective way to describe those parameters is by looking at two fields, the "velocity potential" and "stream function", both derived mathematically from the u- and v-component of the actual winds obtained from the model output. Both are functions of scalar quantities with dimension L^2T^{-1}. In the following subsections, brief descriptions of these two fields are given in turn. Finally, previous studies with either a GCM or a simpler model suggest that large-scale deforestation weakens regional atmospheric circulations (e.g. Zhang *et al*., 1996b; Eltahir *et al*., 1996). The relative reduction in boundary layer entropy compared to the surroundings causes the weakening of the circulation. The significance of the boundary layer entropy is discussed here before its use as a diagnostic tool in Chapter 10.

8.4.2: The velocity potential

Velocity potential is a parameter which can conveniently describe divergence patterns of the atmospheric flow. This parameter is defined in terms of equipotential lines of the divergent wind vector acting normal to it. Note that when we consider a velocity potential, the flow must be irrotational; that is, the vorticity must be zero.

If the vorticity vanishes (i.e. the flow is irrotational) throughout a region (or, practically, over the globe) then mathematically we can write:

$$\nabla \times \mathbf{v} = 0 \tag{8.1a}$$

or,

$$\nabla \times (\nabla \phi) = 0 \tag{8.1b}$$

Hence, ϕ, the velocity potential, can be defined so that

$$\mathbf{v} = -\nabla \phi \tag{8.2a}$$

or,

$$\nabla \cdot \mathbf{v} = \nabla^2 \phi \tag{8.2b}$$

Equation 8.2a implies that **v** is normal to the equipotential lines and is directed from high to low potential. Thus, **v** is referred to as the divergent component of the actual u- or v-wind component which can be resolved from the actual wind to present the potential field known as the velocity potential, ϕ, which always exists in the irrotational fluid motion of the atmosphere. The velocity potential is particularly useful in portraying the pattern of divergence at any level of the atmosphere, after resolving the actual u- and v-component of wind to their divergent counterparts.

8.4.3: The stream function

Stream function is another parameter that is used to conveniently describe the horizontal air current at a level of non-divergence. When divergence vanishes throughout a region (or, practically, over the globe), this condition is called solenoidal or incompressible motion. Mathematically, we can write:

$$\nabla \cdot \mathbf{v} = 0 \qquad (8.3a)$$

or,

$$\nabla \cdot (\nabla \times \Psi) = 0 \qquad (8.3b)$$

Hence, ≅ can be defined such that

$$\mathbf{v} = \nabla \times \Psi \qquad (8.4a)$$

Then using (9.4a), for a level of non-divergence in a horizontal air current, a stream function ( ) may be defined such that:

$$\mathbf{v} = \mathbf{k} \times \nabla \psi \qquad (8.4b)$$

where **v** is the wind velocity vector, **k** is the vertical unit vector and ∇ the gradient of the stream function. The wind velocity vector is normal to, and to the left of ∇ (in the northern hemisphere), that is the wind blows along the isopleths of with low values to the left. The isopleths of are, therefore, called the true "streamlines" which represent the exact flow of air. Thus, **v** is referred to as non-divergent component of the u- or v-wind component which can be resolved from the actual wind to present the stream function . The stream function is useful in portraying the pattern of pure stream flow at any level of the atmosphere after resolving the actual u- and v-component of wind to their non-divergent counterparts.

8.4.4. Tropical deforestation, net surface radiation and boundary layer entropy

There is an important relationship between the net surface radiation and boundary layer entropy. The modification in the surface energy balance following deforestation affects the boundary layer entropy primarily through the change in the total flux of latent and sensible heat from the surface into the atmosphere. The sign and magnitude of the change in the boundary layer entropy depends on how the total flux of heat, including latent and sensible forms, may change after deforestation. The change from forest to short grass or bare soil increase the relative magnitude of the sensible heat flux compared to the latent heat flux resulting in a larger Bowen ratio (cf. Equation A.3c in Appendix A). The reduction in evaporation follows mainly from the smaller root depth associated with short grass in comparison to forest.

The less obvious reaction, however, is how deforestation changes the magnitude of the total flux of latent and sensible heat. This could be addressed by considering the energy balance at the boundary between the land surface and the atmosphere following a description given by Eltahir (1996). Before deforestation, the equilibrium state of the energy balance at any point in the boundary between the land surface and the atmosphere is described by

$$N - F = 0 \tag{8.5}$$

where N is the net surface radiation and F is the total flux of heat from the surface, including both the latent and sensible forms. Each of the terms N and F has two components,

$$\left.\begin{aligned} N &= N_s + N_t \\ F &= \lambda E + H \end{aligned}\right\} \tag{8.6}$$

where N_s is the net solar radiation equivalent to incident solar radiation minus reflected solar radiation; N_t is the net terrestrial (long-wave) radiation defined as downward flux minus upward flux of terrestrial radiation; E is evaporation; λ is the latent heat of vaporisation; and H is sensible heat flux.

Following deforestation, a new equilibrium of the surface energy balance is achieved which is described by

$$\lambda(E + \delta E) + (H + \delta H) = (N_s + \delta N_s) + (N_t + \delta N_t) \tag{8.7}$$

where Λ preceding any of the terms denotes a small change in that variable due to deforestation. Subtraction of (8.6) from (8.7) yields

$$\delta F = \lambda \delta E + \delta H = \delta N_s + \delta N_t = \delta N \tag{8.8}$$

Equation 8.8 suggests that the change in the total flux of heat F is exactly the same as the change in net surface radiation, including both components: solar and terrestrial radiation. The results from the ensemble simulations reported earlier (cf., Table 9.3) show that Southeast Asian deforestation modifies the surface energy balance by reducing the total net surface radiation, including terrestrial and solar forms. Most of the reduction in net radiation comes from the change in solar (SW) radiation. Following (8.8), the energy balance at the land-atmosphere boundary requires that any reduction in net surface radiation has to be balanced exactly by a similar reduction in the total flux of heat from the surface, including latent and sensible forms. Hence, large-scale deforestation should result in smaller flux of heat from the surface into the boundary layer. Given that is true, then the change in boundary layer entropy will follow the change in net radiation at the surface.

Although boundary layer entropy over any region is primarily controlled by the surface fluxes of latent and sensible heat, there are potential feedbacks due to the changes in three other processes: entrainment at the top of the boundary layer, convective downdrafts, and radiative cooling of the boundary layer air. According to Eltahir (1996), the three potential feedbacks however, are relatively small compared to the effects due to the change in the total flux of heat. These three processes are likely to introduce an additional reduction in boundary layer entropy by producing positive feedbacks. Following deforestation, the increase in the surface temperature would tend to increase the intensity of entrainment at the top of boundary layer resulting in further reduction of boundary layer entropy and causing a positive feedback. The same increase in boundary layer temperature enhances radiative cooling and results in an additional sink of entropy. A potential negative feedback that may increase boundary layer entropy is the reduction in convective downdrafts if the cloud amount is decreased.

The complete explanation of entropy is derived from the second law of thermodynamics. In terms of heat conduction and entropy, the second law of thermodynamics implies that heat flows from the warmer to the cooler regions of a system and that entropy flows in the same direction. For our purpose here, we shall merely define the entropy: the increase in entropy of a system is given by the ratio of the heat added in the system to the temperature of the system at which it is added. For a reversible adiabatic process, the entropy does not change; therefore, it is what we call an isentropic process. Following Fleagle and Businger (1980), a change in potential temperature, Π, is related to a change in specific entropy, *s*, by

$$ds \quad = \quad c_p \, d\ln\theta \qquad (8.9)$$

For an adiabatic process, the potential temperature, Π, can be used to represent the entropy, *s*. In a saturated adiabatic process, however, the equivalent potential temperature, Π_e, is an appropriate parameter to represent either saturated or unsaturated moist air, and for moist entropy. These two parameters (Π and Π_e) are related by

$$\theta_e = \theta \exp\left[\frac{Lw_s}{c_p\, T}\right] \tag{8.10a}$$

where L is latent heat of vaporisation , c_p is specific heat capacity at constant pressure, w_s is saturated mixing ratio, and T is temperature of air.

A convenient approximation may be obtained by expanding (8.10a) as a Taylor series and retaining only the first two terms with the results

$$\theta_e = \theta + \frac{Lw_s}{c_p\, T} \tag{8.10b}$$

The second term on the right of (8.10b) may be neglected for practical purposes, because $Lw_s/c_pT \sim 10^{-1}$ for w_s of 10 g kg^{-1} and is less for smaller w_s. Therefore, the use of the potential temperature, Π, is a sufficient approximation for the equivalent potential temperature, Π_e, to represent the moist entropy in the analysis used in this chapter. The use of either Π or Π_e, actually, will not affect the interpretation of our results in representing the moist entropy since this parameter is only used for visual interpretation in this study.

8.5: Summary

The general circulation model experiments run to test the sensitivity of the Southeast Asian climate and regional atmospheric circulation to deforestation has been described and justified. The Unified Model in atmosphere-only mode with fixed sea surface temperature specified from climatology has been employed. To overcome dependence on initial conditions, an ensemble approach has been undertaken with ten cases for each experiment which is defined by atmospheric initial conditions drawn from the European Centre for Medium-Range Weather Forecasts (ECMWF) observational data for the appropriate start-date. As three-region deforestation schemes used in previous GCM experiments can give rise to a misleading answer in the estimation of the actual contribution of each deforested region, therefore, the forcing for the deforestation experiments in this study has been single region, i.e. land surface characteristics are altered for the Southeast Asian region alone enabling a clear definition of the contribution of Southeast Asia deforestation.

Comparing the deforested with the control simulation all the vegetation parameters set-up have been changed as follows:

- root depth (i.e., the average depth of soil from which moisture is available to plant roots) is reduced;
- albedo is increased as grassland reflects more solar radiation than forests which are very efficient absorbers and scatterer of short-wave radiation;
- surface resistance to evaporation is less for grassland under freely available soil moisture;
- roughness length is reduced as forest offers a significant resistance to the wind in the lower layers of the atmosphere;
- canopy capacity is reduced for grassland compared to the forest when the retention of water become less on smaller leaves of grass or crop trees;
- vegetation fraction, in reality, should become less as forest is converted to grassland;
- and the infiltration factor is larger for forests as tree-root systems and forest litter enhance the infiltration rate.