CHAPTER 3: MODELLING THE EFFECTS OF TROPICAL DEFORESTATION ON CLIMATE: BACKGROUND

3.1: Introduction

As apparent in the preceding chapter, tropical forests are disappearing or deteriorating, due especially to population pressures and non-sustainable resource utilization. The chief concerns surrounding tropical deforestation are the consequent impacts on species diversity, atmospheric chemistry, and land surface-atmosphere interaction. Dickinson (1987) gives an overview of the linkages between tropical forests and the atmosphere, soils, and rivers in the Amazon region. According to Dickinson, the key links include exchange of important trace gases, heat, and moisture with the atmosphere. These linking processes make tropical forests an important component of our climate system. In this chapter, the major processes underlying the effect of deforestation on climate are discussed and modelling approaches are defined, prior to an overview of previous general circulation model (GCM) experiments in Chapter 4.

3.2: Tropical Deforestation and Climatic Change

3.2.1: The role in the global carbon cycle

The impacts of tropical deforestation on climate are usually viewed as the result of their role in the global carbon cycle (e.g., Houghton *et al.*, 1994, 1996; Watson *et al.*, 1996; Enquete-Kommission, 1995). Since tropical forests play an important part in the global carbon cycle, therefore, their conversion through land use change and practice can significantly alter carbon dioxide fluxes between the atmosphere and biosphere. Destruction of forests by man is reducing carbon pool stored on the earth. From an assessment on anthropogenic carbon budget, it has been estimated that resultant cumulative emissions from changing land use from 1850 to 1900 is 122 " 40 GtC (Houghton *et al.*, 1994). As also reported by Houghton *et al.*, the largest source of CO_2 emissions from changing land use is currently from the tropics. Emission from the tropics has generally been increasing since the 1950s and the estimated range for 1980 alone is from 0.4 to 2.5 GtC. With considerable uncertainty in estimated emissions from changing land use in the tropics, it was generally agreed that the average emission is 1.6 " 1.0 GtC/yr in the 1980s (Houghton *et al.*, 1994, 1996; Watson *et al.*, 1996). Alongside unprecedented consumption of fossil fuels and cement production, the resulting emissions from deforestation are also responsible for a significant percentage of the anthropogenic greenhouse effect. The contribution from deforestation is between 10 and 20 percent of this man-made greenhouse effect (Enquete-Kommission, 1995).

3.2.2: Changes in land surface properties

On the other hand, replacing tropical forests with other vegetation such as pasture, grassland or estate tree crops will lead to significant changes in land surface properties. One special property of tropical forests is that they have a very low surface albedo throughout the year. This condition either directly or indirectly influences the atmospheric radiation budget (e.g. Charney, 1975; Mylne and Rowntree, 1992). In a review of various radiation measurements for the Amazonian rain forest, Gash and Shuttleworth (1991) conclude that the albedo of tropical forests on average is between 0.11 and 0.13. Within this range, closed tropical forests can be assumed to reflect only about 13 percent (at most) of incident solar radiation, and at least 87 percent of solar radiation is absorbed. According to Gash and Shuttleworth, if forest is replaced by short grass, grazed by cattle, then albedoes will increase up to 0.25 or greater, and will directly result in a decrease in energy absorbed by the surface. Charney (1975) postulated a mechanism by which an increasing albedo resulted in the atmosphere above an arid area becoming a net radiative sink. This will then increase the transfer of radiation back to space at the top of the atmosphere, leading to enhance subsidence of the atmosphere and consequently reduce clouds and rainfall. As the result, more energy is absorbed by the surface. Although Charney (1975) refers to an arid area, processes following an extreme deforestation could also result in a similar mechanism. Charney's mechanism was supported by Charney *et al.* (1977) and Mintz (1984) in early GCM experiments. Later work using GCMs is discussed in Chapter 4.

3.2.3: The role in the hydrological cycle

Tropical forests also play an important role in the hydrological cycle (Mintz, 1982; Dickinson and Henderson-Sellers 1988; Henderson-Sellers *et al.*, 1993). Water is lost from trees to the atmosphere through interception loss (evaporation of the part of the rainfall that remains on leaves) and through normal plant transpiration. The leaf and stem areas of tropical forests are larger than those of any other vegetation, and the trees are tall. Reductions in leaf and stem area following deforestation lead to a decrease in water-holding capacity of the vegetation, leading to a decrease in evaporation of intercepted precipitation and transpiration. Other smaller vegetation like pasture and grassland are, however, much shorter and smoother than forest, so that surface roughness is dramatically reduced, resulting in reducing evaporation. A competing effect may occur, as shorter and smoother vegetation following deforestation also results in reducing surface friction, strengthening surface wind speed and therefore enhancing evaporation. Most model simulations, however, indicate an overall decrease in surface evaporation following deforestation (McGuffie and Henderson-Sellers, 1997). According to Salati (1987), in the Amazon region, for example, from the total incident rainfall, the quantity of water vaporized by interception is about 25%, plant transpiration is 50%, and water runoff (above or below ground) is 25%. Thus, in a dense forest, 75% of the precipitation is returned to the atmosphere by evapotranspiration. In less oceanically affected areas in particular, interception and transpiration by rain forests are the chief sources of moisture for cloud formation.

Meher-Homji (1991) contends that a reduction in forest cover in India, Southeast Asia, Africa, and Latin America caused a decrease in precipitation. By referring to their previous GCM experiments, McGuffie and Henderson-Sellers (1997) summarise the effect of deforestation, and point out that changes in surface evaporation will act as a connection between the changes in hydrological processes that determine regional water recycling. There is an interaction between changes in surface evapotranspiration and changes in total precipitation following deforestation. When surface evapotranspiration is reduced over deforested regions, then it will result in reduction in total precipitation. This reduction in the total precipitation further reduces the evapotranspiration. At the same time, the changes in surface evapotranspiration provide a connection between the hydrological processes and the changes in the land-surface and atmospheric energy budget (the sink of the surface energy budget and the source of the atmospheric energy budget). The weakening of the surface evapotranspiration causes a reduction in latent heat flux over the region that then leads to a net reduction in the atmospheric energy budget. Eventually, the regional atmospheric circulation is weakened and less water vapour is delivered into deforested regions. The scale of moisture convergence, cloudiness and convective activity changes may also contribute to a possibility of non-local climatic impacts following deforestation of a certain tropical region. McGuffie and Henderson-Sellers state that, following deforestation, the compensating effects between reduction in the surface net radiation (due to increased albedo) and reduction in the surface evapotranspiration results in a small increase in surface temperature in deforested regions.

In the context of the global atmospheric circulation, since strong ascending branches of the Walker and Hadley circulation are located over tropical forest regions, it has also been suggested that changing land-surface characteristics in regions of tropical forest may affect the general atmospheric circulation which could then extend the effect of deforestation to the global scale (e.g. Henderson-Sellers *et al.*, 1993; Polcher and Laval, 1994a, b; McGuffie *et al.*, 1995; Zhang *et al.*, 1996a, b). Thus, the atmospheric disturbance induced by tropical deforestation may also interact with the large-scale circulation.

3.3: The Land Surface in the Climate System

3.3.1: Surface-atmosphere interactions

In the climate system, the land surface acts as a lower boundary for approximately 30% of the earth atmosphere. It exchanges moisture, momentum and heat with the atmosphere. As noted by Dickinson (1992), there are significant differences between the land surface and the ocean, particularly regarding their roles in the climate system. The land surface has lower heat capacity than the oceans and therefore it provides much less energy storage and negligible horizontal energy transport, compared to the oceans. The land surface is, however, more variable and changeable than the oceans in many coupling processes with the atmosphere and it is much more responsive to net radiation from or through the atmosphere than are the oceans. Different from the oceans, land can be either wet, moist or dry, and the land system consists of wide-ranging of heterogeneity, with great difference in soil properties and vegetation cover distribution. Surface albedo therefore varies as vegetation and soils have large spectral variations.

The important interactions between land surface and the atmosphere involve the exchanges of radiation, sensible heat, latent heat and momentum. The bulk of the net incoming solar radiation is absorbed not by the atmosphere but by the underlying surface. Evaporation of moisture and the heating of the surface lead, however, to much of this energy being transferred to the atmosphere as latent and, to a lesser extent, sensible heat. The land-surface exchanges momentum with the atmosphere by a turbulent stress exerted on the lower boundary of the atmosphere. Surface wind will also be a factor in determining the contribution of turbulent exchange processes in the heat and water vapour balances at the surface-atmosphere interface.

3.3.2: Sensitivity of regional and global climate

Early GCM experiments were undertaken to determine the extent of the sensitivity of the regional and global climate to changes in the various fluxes caused by deforestation (e.g. Charney *et al.,* 1977; Shukla and Mintz, 1982; Mintz, 1984; Sud and Smith, 1985a, b). As commented by Sellers (1987), although sensitivity experiments from those early GCMs assumed only drastic and unrealistic changes in land surface parameters (e.g. albedo, roughness length and soil moisture), their results showed significant surface-atmosphere interactions.

Shukla and Mintz (1982), in their sensitivity experiments, compared two cases: first, the land surface was assumed to be evaporating freely; second, the land surface was assumed completely dry. Their results show both cases are considerably different and they conclude that the global fields of rainfall, temperature, and wind vector strongly depend on land surface evapotranspiration. The storage of heat and moisture in the soil is therefore a key factor in determining the spatial and temporal character of climate; it modulates the diurnal and seasonal periodicity of the atmospheric heat and moisture fluxes. Moisture mass added and stored in the soil is released back to the atmosphere with delays considerably longer than that characteristic of moisture residence in the atmosphere (Entekhabi and Eagleson, 1991). An interactive soil water reservoir is thus partly responsible for the persistence and long time scales associated with near surface fluctuations in temperature and relative humidity (Delworth and Manabe, 1989).

Over the regions where monsoonal circulations are prominent, any changes in land surface character that influence this circulation will also affect their climate. Sud and Smith (1985a) used a GCM experiment to investigate the influence of changes in land-surface fluxes over the Indian subcontinent. Following an increase in surface albedo and a reduction in surface roughness, as imposed in their experimental run, their results show a significant weakening of the Indian Monsoon circulation. The result also conforms with Charney's (1975) hypothesis, that higher surface albedo reduces rainfall. Furthermore, Sud and Smith showed that low surface roughness makes horizontal transport of moisture in the planetary boundary layer become more westerly. Then, cross-isobaric moisture convergence and therefore rainfall are both reduced over the northwestern India, while correspondingly increased over China. These results suggest that deforestation over the Indian subcontinent would have a significant influence on its July (monsoon) rainfall. According to them, the magnitude of their prescribed changes in surface albedo and surface roughness could plausibly be produced by major changes in tall natural vegetation over this subcontinent.

At a different extreme, particularly for desert or semi arid regions, many studies indicate the significant influence of land surface on climate (e.g., Charney, 1975; Charney *et al.*, 1977; Sud and Fennessy, 1982). According to Sud and Fennessy (1982), surface albedo change over a particular region causes a variability of mean monthly simulation in areas far away from the anomaly regions. In nature, formation of deserts leads to a reduction of surface roughness as the vegetation perishes and soil erosion ensues. Sud and Smith (1985b), in their GCM simulations and comparative analysis, showed that rainfall in the Sahara desert is reduced significantly following a reduction in surface roughness. The influence of low surface roughness of desert on the July rainfall is comparable to that of an increase in surface albedo. Also with a GCM experiment, Xue and Shukla (1993) showed that moisture flux convergence and rainfall are reduced in the prescribed desertification area, which led to changes in the surface characteristics, among others, including albedo, roughness length and soil moisture. Moisture flux convergence and rainfall were, however, increased to the immediate south of the test area. These findings are consistent with those observed during sub-Saharan dry years in which the axis of the maximum rainfall shifts to the south. The changes in atmospheric circulation patterns also agree with observations. Then, as an opposing perturbation, Xue and Shukla (1996) used several short-term

GCM integrations to explore the effect of large-scale afforestation in the sub-Saharan area on climate. According to their results, rainfall increases in the afforestation region but decreases to the south of that region. The perturbation imposed alters the surface energy balance and induces a circulation change over the perturbed region and adjacent areas.

In a series of studies, a research team at Ralph M. Parsons Laboratory, MIT (Eltahir and Bras, 1993; Eltahir, 1996; Eltahir and Gong, 1996; Zheng and Eltahir, 1998) used a linear or zonally symmetric model to describe the response of the tropical atmosphere to large-scale deforestation. Eltahir and Bras (1993) used a simple linear model of the tropical atmosphere in studying the effects of deforestation. They suggested that the impact of large-scale deforestation on the circulation of the tropical atmosphere consists of two components: the response to the positive change in surface temperature, and the response to the negative change in precipitation (latent heating). Owing to their different signs, the changes in predicted temperature and precipitation excite competing responses working in opposite directions. The two competing responses, or mechanisms, they suggest are: (1) a thermally direct converging circulation in the boundary layer driven by the increase in surface temperature and (2) a diverging circulation in the boundary layer due to the corresponding decrease in rainfall and latent heating. The study of Eltahir and Bras (1993) explains the evident sensitivity of GCMs results regarding the changes in runoff and circulation as due to the competition between the two different responses. The predicted change in the tropical circulation, according to them, determines the change, if any, in atmospheric moisture convergence, which is equivalent to the change in run-off. Their study, however, does not address the question of the ultimate sign of the overall change in the total atmospheric circulation following large-scale deforestation. Instead, the final result is left to the competition between the two mechanisms.

Eltahir (1996) states that the origin for the concept that deforestation excites two different and competing responses lies in the separation between the thermodynamics of atmospheric water vapour from the dynamics of the dry atmosphere. Eltahir's study, therefore, considers the unified dynamics of the moist tropical atmosphere to address the question of what happens to the total atmospheric circulation following deforestation. Eltahir suggests that in a moist atmosphere that satisfies a quasi-equilibrium between moist convection and radiative forcing the equilibrium temperature profile must be uniquely related to the boundary layer entropy. Under such conditions, large-scale deforestation reduces boundary layer entropy relative to the surroundings.

Zheng and Eltahir (1998) report on their study using a moist zonally symmetric atmospheric model coupled with a simple land surface scheme to investigate the role of the meridional distribution of vegetation in the dynamics of monsoons and rainfall over West Africa. Their results demonstrate that West African monsoons and, therefore, the rainfall distribution depend critically on the location of vegetation perturbations. Changes in vegetation cover (i.e., desertification) along the border between the Sahara desert and West Africa may have a minor impact on the simulated monsoon circulation. Coastal deforestation, however, may cause the collapse of the monsoon circulation and have a dramatic impact on regional rainfall. Zheng and Eltahir conclude that deforestation in West Africa is likely to be a significant contributor to the observed drought in this part of the world.

It is clear, then, that, although the precise effect is specific to each particular situation, the climate system is critically sensitive to the characteristics of the underlying land surface.

3.4: Parameterization of Land Surface Processes

Parameterization of land surface processes is necessary as modellers deal with large-scale climate modelling in order to incorporate the contribution of small-scale physical processes (Smagorinsky, 1982). The land surface interacts strongly with the atmosphere at all scales and therefore the interactions are complex. At the earth-atmosphere interface, the major transports of momentum, heat, and water vapour are generally simulated by macroscale motions that are explicitly resolved. For subgrid-scale diffusion, however, specification of some additional empirical parameters in a model are required in order to produce more realistic simulations. As the land surface is usually extremely heterogeneous over the resolvable scales considered in a model, surface parameterizations based on an assumption of "grid-scale homogeneity" may fail to represent the land-atmosphere interactions that occur at much smaller scales.

To this end, inadequate representations of land-atmosphere interactions have gradually been improved. Modellers have continued to develop more complex GCMs from a crude representation of significant processes on the Earth's land surface (as reviewed by Carson, 1982), to a more realistic formulation based on the solution of energy budget equations applied to soil surface and vegetation layers (e.g., Dickinson 1984; Sellers *et al.*, 1986; Sellers, 1987; Henderson-Sellers *et al.*, 1993). Future development of land surface parameterizations is perceived to be a challenging area of research for the modelling communities.

Historically, three broad strategies have been used to model land-surface processes in GCMs: first, the prescription of land surface parameters; second, the use of conceptual models; and finally, the use of biophysically based models. According to Sellers (1987), in earlier GCMs, particularly before the work of Dickinson (1984), all surface parameterizations had been confined only to the prescription of land surface parameters and the use of conceptual models. These two approaches are documented in Appendix A, based on Sellers (1987, 1991 and 1992).

The early modelling approaches, based on the prescription of land surface parameters and conceptual models, suffered from flaws that made them unrealistic and inaccurate (Sellers, 1987 and 1992). The major shortcomings are as follows:-

i. *Independent specification of surface attributes*: the transfers of radiation, momentum, and heat were specified as independent processes, that is, surface albedo, roughness length, and moisture availability were frequently specified as completely independent, unrelated characteristics of the land surface, rather than as complementary facets of the vegetation-soil system.

ii. *Unrealistic specification of albedo, roughness length, confusion of momentum and heat/vapour transfer pathways*: albedo fields were specified from data with no spatially integrating theory applied to check the reasonableness of the prescription. Roughness length z_o was sometimes used simply as one number for all land surfaces, similarly for heat, water, and momentum transfer.

iii. *Unrealistic specification of soil hydrology*: the ‘bucket’ model used is very far from reality, as in nature soil moisture capacities vary from virtually nil (desert, mountains) to several metres (tropical and some temperate forests).

iv. *Unrealistic description of the evapotranspiration process*: the use of the β-function (see Appendix A), range from zero to unity, fails to account for physically significant surface resistance, and the formulation systematically overestimates surface evaporation rates in the humid regions. According to Sellers (1992), this is probably the most damaging of all the flaws in the early models. It is not the way in which water vapour is transferred from the land to the atmosphere in the real world.

Based on the first two approaches, biophysical models were introduced and currently are being used to study more realistically the various feedback processes acting between the atmosphere and the vegetated surface. In the early conceptual models using bucket hydrology, vegetation was viewed as a passive sponge-like structure or as a pervious sheet separating the soil from the atmosphere. A more realistic model must consider plants as living organisms that regulate the passage of water and gas through their systems in their growth and survival. The link between surface and atmosphere in a model grid area is influenced by the physiology and morphology of the vegetation. For this reason, a biophysically realistic modelling approach was developed for land surface parameterizations, with the emphasis was on designing soil-vegetation-atmosphere transfer schemes (SVATs). This type of model specifies surface attributes of albedo, roughness and evaporation rate as mutually consistent surface properties, rather than use an independent specification of surface attributes.

3.5: Biophysically Based Models: An Overview

The models of Dickinson (1984) and Sellers *et al.* (1986) were the first attempts to incorporate biophysical (SVATs) principles into surface formulations appropriate for use within GCMs. In these first attempts, Dickinson (1984) introduced the Biosphere-Atmosphere Transfer Scheme (BATS) and Sellers *et al.* (1986) designed the Simple Biosphere Model (SiB). Sellers (1987, 1991) describe general strategies adopted in formulating the biophysical models of Dickinson (1984) and Sellers *et al.* (1986). These strategies are summarized as follows:

i. *Radiation absorption*: The chlorophyll molecule in a leaf of a green plant absorbs the visible (photosynthetically active radiation - PAR) wavelength interval of 0.4 - 0.72 μm in order to combine water and CO_2 into organic compounds needed for growth and maintenance. Green leaves are weakly absorbent (or moderately reflective) in the near-infrared region (0.72 - 4.0 μm) as they cannot drive photosynthesis with this lower energy radiation. Furthermore, the differing orientations of leaves within canopy and the effect of multiple reflection and reinterception of radiation between leaves make the canopy an effective radiation trap. In contrast, bare ground generally exhibits a gradual increase in reflectivity (from 0.1 - 0.3) with wavelength over the interval of 0.4 - 4.0 μm.

ii. *Biophysical control of evapotranspiration*: The photosynthetic reaction in the plant metabolism ensures CO_2 reaches the chlorophyll sites, and makes leaves maintain an open pathway between the atmosphere and their saturated tissues. This leads to an inevitable loss of water vapour over the same route and higher plants regulate the amount of gas exchange (and therefore water loss) by means of valve-like structures on the leaf surface called stomates. Since the radiant energy absorbed by the surface is mainly divided between sensible and latent heat, any decrease in the evapotranspiration rate will be approximately balanced by an increase in sensible heat loss. In addition, interception of precipitation by vegetative canopies that can store the equivalent of several mm of water on the leaf surface also significantly reduces the precipitation input into the soil. The intercepted water could be evaporated readily without reaching the ground.

iii. *Heat and momentum transfer*: In the planetary boundary layer, turbulent airflow enhances the transport of sensible and latent heat from the surface. A relatively rough, porous surface of vegetative canopies produces greater turbulence than short pastures/grassland or bare soil. Furthermore, vegetative canopies also exert a drag force that is significantly larger than grass or bare ground.

iv. *Insulation*: Under a dense vegetation canopy, the soil surface intercepts less radiation and is sheltered (aerodynamically). The energy available to the covered soil will be small and the component terms of the soil energy budget (evaporation, sensible heat flux and ground heat flux) will be correspondingly reduced.

v. *Soil moisture availability*: The amount of soil moisture available for evapotranspiration is dependent on the depth and density of the vegetation root systems.

A model which takes account of the above factors is constrained by the need for a solution that simulates the energy budget of all vegetation communities by a simple alteration of its parameters. In SVATs, vegetation is treated as a separate layer, scaling (usually linearly) from a size of normal leaves up to a grid square of sizes ranging from 50 x 50 km to 500 x 500 km. Usually only three land components (soil, snow, and vegetation) are treated explicitly, while lakes and other land covers are neglected (Henderson-Sellers *et al.*, 1993). Based on the above five factors, for surface calculation, the schemes have primarily involved five key elements: canopy conductance; aerodynamic resistance; albedo; water holding capacity; and runoff. For coupling with an atmospheric model, three elements are involved: precipitation; radiation; and the planetary boundary layer. BATS of Dickinson (1984) has since been used in a number of GCM simulation studies of the climatic impact of large-scale deforestation by coupling with the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM). The details of BATS coupled with CCM are described in Dickinson *et al.* (1986), and a new version (BATS1e) is provided in Dickinson *et al.* (1993). While differing in detail, much of the philosophy of Dickinson (1984) is also incorporated in the construction of SiB by Sellers *et al.* (1986).

3.6 Concluding Remarks

To date, there are many existing land surface schemes developed by different modelling groups. Henderson-Sellers *et al.* (1993) report the existing schemes that have been included in The Project for Intercomparison of Land-surface Parameterization Schemes (PILPS), that is an international project purposely planned to achieve greater understanding of the capabilities and potential applications of existing and new land surface schemes in atmospheric models. In addition to BATS and SiB, among other schemes in PILPS are: the U.K. Meteorological Office (UKMO) Model (Warrilow *et al*., 1986); the Goddard Institute for Space Studies (GISS) Model (Abramopoulos *et al.*, 1988); the Bare Essentials of Surface Transfer (BEST) (Pitman *et al.*, 1991); Simple SiB (SSiB) (Xue *et al.*, 1991); and the Canadian Land Surface Scheme (CLASS) (Verseghy, 1991). Despite some differences in formulation of individual processes, every scheme reveals commonality in aim, design and use. More discussion of such schemes, devoted to the land surface scheme of the model employed in this study, the UKMO Unified Model, will be given in Chapter 7. In the next chapter, the most sophisticated model simulations of deforestation are reviewed, drawing out points introduced in this chapter.