CHAPTER 9: THE IMPACTS OF SOUTHEAST ASIAN DEFORESTATION: (I) BASIC RESULTS FROM THE ENSEMBLE GCM SIMULATIONS

9.1: Introduction

The main purpose of this chapter is to describe and discuss the basic results from the ensemble GCM simulations of the effects of Southeast Asian deforestation. At the beginning, in Section 9.2, an assessment will be presented of the ensemble control simulation by giving a comparison between the simulated climatological variables with observations (re-analysis data). The variability of the ensemble members is also presented in Section 9.3, to serve as a guide in the interpretation of the results. The impacts of Southeast Asian deforestation on local and regional climates are then presented in Section 9.4, followed by discussion of the causes of the simulated climatic changes in Section 9.5. Emphasis is given to changes in the radiative energy budget and hydrological processes. Finally, in Section 9.6, conclusions are given together with a brief discussion of points leading to further investigation of changes in the broader atmospheric circulation which will be reported in the next chapter. The emphasis in this chapter is on the immediate effects of deforestation.

9.2: Assessment of the Ensemble Control Simulation

In this section, a comparison will be made for some important climatological variables between real-world observations (in the form of re-analysis data) and the ensemble of ten control simulations (called "the ensemble" hereafter) for the Southeast Asian region. The ensemble control data for January and July is derived from the ten 60-day simulations for winter (December - January) and summer (June - July), as described in Chapter 8. For the purpose of this assessment, as in Chapter 5, both the NCEP/NCAR Reanalysis Data (Kalnay *et al*., 1996) and the Legates and Willmott (1990a,b) global climatology of mean monthly precipitation and surface air temperature were employed. For convenience, these data are referred to as "observations". Regridding was applied to these data to make their resolution similar to the model data (see discussion in Chapter 5).

A description of the climate of Southeast Asia with reference to conditions during the winter and summer monsoons (as represented by January and July respectively) has already been given in detail in Chapter 5 and will not be repeated here. Rather, particular attention is given here to identifying any bias produced by the ensemble so that necessary precautions can be taken when interpreting the results of the deforestation ensemble simulations.

9.2.1: Precipitation

Figure 9.1 shows the precipitation for January and July from the observations (after Legates and Willmott, 1990a) and the composite derived from the ensemble. In January, the ensemble precipitation distribution seem to be fairly satisfactory, agreeing well with the observations for much of the area of the West Pacific and the Indian Ocean. The mean position of the Inter-Tropical Convergence Zone (ITCZ) is situated to the south of the equator at this time of the year, resulting in regions of high precipitation positioned with an E-W orientation. The ensemble captures this feature well, though the magnitudes are slightly underestimated especially in the region north of Australia. The precipitation maxima centred over Malaysia-Indonesia (covering eastern Borneo, southern Peninsular Malaysia, southern Sumatra and Java) associated with the cyclonic lifting of low-level winds are poorly simulated as the magnitude is much underestimated in the ensemble. The intense precipitation located over the southeast of Philippines is also inadequately simulated in the ensemble.

In July, the mean position of the precipitation features in the ensemble, overall, is fairly close to the observations despite the region of maxima associated with the summer monsoon over north-eastern Bay of Bengal being displaced to the west. The strength of the maxima over this monsoon region is also underestimated and its NW-SE orientation is also displaced to a NE-SW position, shifting towards the middle of Bay of Bengal. The precipitation maxima located over the north of the Philippines and the equatorial western Pacific are not captured by the ensemble. Though the mean distribution of precipitation agrees with the observations over the western Pacific, the ensemble maximum is less intense than the observations suggest. The ensemble precipitation over the Malaysia-Indonesia region is less than 200 mm, agreeing well with the observations. There is no indication of positive bias of precipitation over the east coast of the Peninsular Malaysia as portrayed in the NCAR CCM analysed in Chapter 5. This suggests that the so-called rain shadow effect of the high mountain ranges of Sumatra is better represented by the current model, probably due to the increase in the model horizontal resolution compared to the NCAR CCM.

9.2.2: Surface temperature

As shown in Figure 9.2, the ensemble surface temperatures in January and July agree quite well with the observed surface temperatures from Legates and Willmott (1990b). The ensemble, however, produces too strong a temperature gradient from north to south as the continental interior is cooler than observed in both seasons. The northern part of continental Southeast Asia is most affected by this temperature gradient. Over the ocean, the ensemble shows good agreement with the observations in both seasons. This is as expected since the ensemble surface temperature here is controlled by the prescribed sea surface temperature derived from the monthly climatology. The inaccuracy in the temperature gradients is a source of concern because of the link with the regional monsoon circulations.

9.2.3: Sea level pressure and low-level winds

In January, the ensemble produces pressure and gradient level (850 hPa) wind patterns that are in good agreement with the observations (Figure 9.3a and c). Both indicate a strong pressure gradient from north to south resulting in a constant northeasterly trade wind over most of the Southeast Asia region. On crossing the equator, the northeasterly trades turn to becoming northwesterlies and merge with the southeasterly trades forming convergence zones (the ITCZ), which are well reproduced in both the wind vector and pressure plots of the ensemble. There is a strong northeasterly "jet" located over South China Sea resulting from cold surges of the Asian continent high pressure system, which is also very well captured by the ensemble, though the strength seems slightly greater than in the observations. In fact, overall, the 850 hPa wind strength over the whole of Southeast Asia is overestimated by the ensemble. This could be more clearly seen when the wind is resolved into its zonal and meridional component (u- and v-component). The ensemble produces a stronger band of easterlies (see Figure 9.4a and c), extending from the Pacific Ocean, South China Sea and Indian Ocean compared to the observed. This is reflected in the strength of northeasterly winds over Southeast Asia and the overestimated "jet" located over the South China Sea. As mentioned in Subsection *9.2.2* above, the inaccuracy in the north-south temperature gradients is a source of concern because of the link with the regional monsoon circulations. The overestimated northerly wind over the South China Sea, therefore, could be linked with the overestimated temperature gradient. The v-component of wind (see Figure 9.5a and c), however, is simulated well by the ensemble, though some underestimation occurs over the Indian Ocean and the Bay of Bengal.

In July, the Southeast Asia region is dominated by the southwesterly surface wind corresponding to a south-to-north pressure gradient (Figure 9.3b and d). Dry southeasterlies originating over the Australian region turn to southwesterlies upon crossing the equator and are reproduced well in the ensemble. The wind strength over the Malaysia-Indonesia region, however, seems to be underestimated by the ensemble. This underestimation can easily be seen when looking at the u-component of wind (Figure 9.4b and d). Obviously, the ensemble positive u-component (westerlies) is weaker than that observed to the north of the equator. On the other hand, the ensemble negative u-component (easterlies) is stronger than that observed to the south of the equator. These biases combined, therefore, produce resultant southwesterly winds over this region which are weaker than observed. The positive v-component (southerlies) over the Bay of Bengal (Figure 9.5b and d) is also overestimated by the ensemble, hence, could possibly exaggerate the summer monsoon's strength over this area.

9.2.4: Divergent winds and velocity potential

Velocity potential and the divergent wind together can effectively describe the motion in the lower and upper troposphere (see Chapter 8). When presenting these fields together for the whole globe, they illustrate the global-scale divergent flow of the atmospheric circulation at various levels. For the purpose of this assessment, however, these fields are only presented for the study area covering South and Southeast Asia so that a comparison between the patterns derived from the ensemble and observations (NCEP/NCAR Reanalysis data in this case) for this region can be made in greater depth. The velocity potential and its associated divergent wind vector at 850hPa and 200hPa are shown in Figure 9.6 and 9.7, respectively. In each figure, the January and July mean climatology obtained from the observations (a and b) and the fields derived from the ensemble (c and d) are presented.

a. Patterns at low level

In January, as illustrated by the observations (Figure 9.6a), the low level regional atmospheric circulation is represented by convergent flow over the entire western Pacific Ocean, maritime Southeast Asia and the adjacent north-eastern Indian Ocean, with the centre located near the dateline around 10^{o}S. The axis of maximum convergence corresponds to the mean position of the ITCZ during this time of the year. Meanwhile, low level divergence dominates most of continental Southeast Asia and the whole of the Indian sub-continent, with the centre positioned over the western edge of Tibetan plateau. Though the ensemble (Figure 9.6c) produces a similar general pattern as the observed data, important features, particularly both the centre of convergence and divergence and zero-line separating the regions of divergence and convergence, seem to be displaced westward. At the same time, the strength of the centre of convergence and the velocity potential gradient are both overestimated by the ensemble, and the associated divergent wind is also stronger than observed. This implies stronger northeasterly monsoonal winds and eventually a more intense ITCZ in the ensemble.

In July, the observations (Figure 9.6b) indicate convergent flow covering the whole of the region with an axis of maxima stretching from western China to the central Pacific, corresponding to the monsoon trough. The major centre is located over the western Pacific northwest of the Philippines. The ensemble (Figure 9.6d) seems to overestimate the strength of the centre and the maximum axis. At the same time, the orientation of the axis of maxima is somewhat shifted southwestward with its centre located near the Indian sub-continent and Bay of Bengal. This implies that a stronger summer monsoon is simulated by the ensemble which could possibly exaggerate the intensity of this weather phenomenon over the Indian sub-continent and Bay of Bengal, consistent with the low-level winds discussed earlier.

b Patterns at upper level

In January (Figure 9.7a), the regional atmospheric circulation at 200 hPa is represented by divergent flow over the entire Pacific Ocean and Southeast Asian region centred near the equatorial region, which is the opposite to the flow at low level. Meanwhile, continental Asia is dominated by upper level convergence corresponding to divergence at low levels. The opposite features in the upper-level atmosphere and the low-level atmosphere below illustrate the existence of a meridional vertical circulation, namely the Hadley cell between the equatorial Pacific and the continental Asia. The ensemble (Figure 9.7c) successfully simulates these overall features, though the intensity of the divergence centre or axis of maxima are slightly overestimated and the zero-line separating the region of divergence and convergence is also shifted southward.

In July (Figure 9.7b), the maximum divergence occurs over the subtropics of the western Pacific Ocean, corresponding with the low-level convergence. This is also a manifestation of the Hadley cell, as the ascending branch is located in this area and the descending branch is positioned in the Southern Hemisphere during this time of the year. The ensemble (Figure 9.7d) simulates the overall features fairly well when the axis of maximum divergence and the major centre over the western Pacific is satisfactorily located, though the intensity is somewhat reduced. Although another secondary centre is produced by the ensemble along the axis of maxima located over Bay of Bengal; its magnitude appears to be consistent with the observed.

9.2.5: Summary

Overall, the climate over Southeast Asia is satisfactorily simulated in the ensemble control experiment. The ensemble captures satisfactorily the observed patterns of precipitation, surface temperature, sea level pressure and low-level winds as well as divergent winds and velocity potential. The model, however, indicates some biases (underestimation as well as overestimation) as far as the magnitude of some of the above variables. Major differences between the ensemble and the observations should be taken into consideration when assessing the impact of deforestation on climate over the deforested region. Major differences are listed as follows: -

- The control ensemble's precipitation over the deforested region is underestimated in January, though it agrees fairly well with the observations in July.
- The control ensemble produces a too strong north-south temperature gradient, as the continental interior is cooler than observed in both January and July.
- The low level (850 hPa) wind strength over the entire Southeast Asia region in January is overestimated by the control ensemble, whereas the strength in July is underestimated.
- The strength of the centre of convergence and the velocity potential gradient are both overestimated and shifted westward by the control ensemble in both January and July.

It seems that there is a link between the stronger temperature gradient and the wind discrepancies. In January, for example, the stronger temperature gradient with cooler continental interior implies that high-pressure system is exaggerated by the control ensemble, overestimating the winter monsoon wind strength over Southeast Asia region. Though a stronger temperature gradient is also simulated by the control ensemble, the strengthening of the monsoon seems to be confined within the Bay of Bengal/Indian subcontinent area without affecting much of the Southeast Asia deforested region.

The validation of the Unified Model control simulations undertaken in this study is consistent with previous assessments (e.g. Lean and Warrilow, 1989; Lean and Rowntree, 1997).

9.3: Variability of the Ensemble Members

Before we move on to analyse the impacts of deforestation, it is important that we have an idea of how members of the ensemble vary from each other. The purpose is to identify areas with large variability among the ensemble members where "spurious" signals may appear. We choose only two variables for this inspection: one is vertical velocity and the other is precipitation. The vertical velocity is chosen in view of its function in the maintenance of the circulation, both zonally and meridionally. The circulation of the atmosphere over the tropics is characterised by the meridional Hadley Circulation and the zonal Walker Circulation. Precipitation is chosen not only because of the high variable nature of this parameter, but it is acknowledged as the most important meteorological variable in the tropics.

9.3.1: Standard deviation of vertical velocity

Figures 9.8a and 9.8b show the meridionally averaged (15^{o}S to 15^{o}N) standard deviation of monthly average vertical velocity from the 10-case ensemble control simulations for January and July respectively, corresponding for the Walker Circulation. In January (Figure 9.8a), the highest variability of vertical velocity occurs around 130^{o}E over maritime Southeast Asia. The second highest variability occurs around 140^{o}W over equatorial and central Pacific Ocean. In July (Figure 9.8b), the highest variability seems to relocate itself, splitting over two major areas: around 50^{o}E over Arabian Sea near Somalia and around 90^{o}E over the equatorial Indian Ocean /Bay of Bengal region. The third highest variability in July occurs around 80^{o}W near the coast of Peru, Ecuador and Colombia.

Figures 9.8c and 9.8d illustrate the zonally averaged (90^{o}E to 150^{o}E) standard deviation of monthly average vertical velocity from the 10-case ensemble control simulations for January and July respectively, corresponding for the Hadley Circulation. In January (Figure 9.8c), the strongest variability appears around 15^{o}S over the region to the north of Australia, followed by the area around 30^{o}N over northern China. In July (Figure 9.8d), apart from the highest variability near the Antarctica, the next strongest variability occurs around 15^{o}N covering part of the Bay of Bengal and Bangladesh, extending east to the southern China region.

The standard deviation of vertical velocity, both for the meridional and zonal cross-section in either January or July, clearly indicates that the variability of this variable is remarkably strong over the area most affected by the monsoons. This means that over the monsoonal areas, each member of the ensemble has the greatest disagreement with each other in simulating the vertical velocity. Since the ensemble members correspond to different years, we can also say that the inter-annual variability of the vertical velocity is greatest over the monsoon-affected areas.

9.3.2: Standard deviation of precipitation

The standard deviation of monthly precipitation from the ensemble simulation is derived and presented both for the control (Figures 9.9a and 9.9b) and deforestation (Figures 9.9c and 9.9d) experiments for January and July. It is obvious that the precipitation variability among the ensemble members is also greatest in the area most affected by the monsoon. In January, the area of maximum standard deviation is closely related with the axis of ITCZ, as would be expected, which is portrayed in both the control (Figure 9.9a) and deforestation (Figure 9.9c) cases. Similarly, in July, the maximum standard deviation corresponds to the monsoon trough, a prominent feature especially over the Bay of Bengal and the South China coastal area. As each of the ensemble members comes from a different year, we can also say that inter-annual variability of precipitation is highest over the monsoon-affected areas.

Though the major point of maximum standard deviation over the monsoon-affected areas is maintained, the patterns, as well as the magnitude of the standard deviation, differ between the control and deforestation experiments for both seasons. This indicates that precipitation variability is also affected by the imposed perturbation and implies that caution is necessary when interpreting results from the deforestation experiments particularly when comparing the changes before and after deforestation. A proper statistical technique for testing significant changes is a necessary requirement rather than just visual interpretation. In this study, we rely on the standard Student's *t*-test for testing for significant changes between the control and deforestation results.

9.4: Impacts of Southeast Asian Deforestation on the Local and Regional Climates

This section examines the simulated impacts of Southeast Asian deforestation on the local climate system. For convenience, the term "the deforested region" refers to the 38 grid-boxes over land areas within Southeast Asia region where vegetation parameters are changed to correspond with deforestation (see Figures 8.3 to 8.9). The term "the study area" refers to a rectangular area (bounded by 90^{o}E to 150^{o}E and 10^{o}S to 20^{o}N) of 180 grid-boxes covering both land points (the deforested region) and sea points (see Figures 8.3 to 8.9).

9.4.1: Spatially averaged regional climatic change

Table 9.1 summarises the monthly averaged climatic changes over the deforested region. Interestingly, for both January and July, all the variables have increased values following deforestation. The values given are area averages for the study area (from 10^{o}S to 20^{o}N and from 90^{o}E to 150^{o}E). However, the surface temperature is averaged only from the 38 land-points in the deforested region. Only the temperature differences are statistically significant at the 5% level according to the *t*-test. As will be shown later, in the case of the other variable, the pattern of change over the study area is varied with some areas exhibiting an increase and others a decrease and it is more appropriate to test for significant at the grid-point level.

In the previous GCM deforestation experiments by McGuffie *et al.* (1995) and Zhang *et al.* (1996a), the authors report a small decrease in surface temperature over Southeast Asia region in most months following the three-region deforestation. Our ensemble simulation of the effects of single region deforestation, however, shows an increase of temperature for both January and July by 0.6^{o}K and 0.5^{o}K, respectively. As mentioned above, the surface temperature differences computed from our results are based on the 38 land-points in the region. This is to exclude the null-change of surface temperature at the sea-points in the region solely controlled by the prescribed sea surface temperature from the monthly climatology.

In the control simulations over the study area, it is shown that the precipitation is mostly convective (~99%). Following deforestation, the total precipitation is increased by 4% in January and 3% in July; the convective precipitation accounting for 93% and 98% of the change in January and July, respectively. A positive change in total precipitation also does not agree with the results from Henderson-Sellers *et al.* (1993), McGuffie *et al.* (1995) and Zhang *et al.* (1996a) when all the three region deforestation experiments showed a reduction of total precipitation almost throughout the year for Southeast Asia region, though few points were statistically significant.

Moisture convergence or divergence is one of the key indicators used to define the impacts of tropical deforestation. It is defined as the difference between total precipitation and total evaporation (P - E). The large positive values for both January and July as simulated by the ensemble control simulation (cf. Table 9.1) indicate that there is strong moisture convergence over the region in both months. The strength of moisture convergence in July, however, is more than three times greater than in January. This is due to the difference in the total precipitation and evaporation between the two months. The total precipitation is higher in July than in January and, conversely, the total evaporation is lower in July than in January. The positive differences in moisture convergence [i.e., –(P - E)] indicate that the "already strong" moisture convergence in both months is further increased by deforestation. The change in moisture convergence is higher in January than in July. In January, a weaker moisture convergence is simulated by the ensemble control; both the precipitation and evaporation increase following deforestation, with the magnitude of the precipitation increment greater than that of evaporation. In July, a stronger moisture convergence is simulated by the ensemble; since there is an increase in precipitation and a smaller increase in evaporation and the difference (P - E) is larger.

The three-region deforestation studies [Henderson-Sellers *et al.* (1993); McGuffie *et al.* (1995); and Zhang *et al.* (1996a)] indicate that evaporation decreases all year round for Southeast Asia. This model's ensemble results (see Table 9.1), however, simulate a slight increase of the total evaporation for Southeast Asia in January (+0.3%) and July (+0.2%). In terms of evaporation, it would be useful to look at the change over the land area (deforested region) alone in addition to the change over the whole region (study area). We should not forget that out of the 180 grid-points that cover the whole region under consideration, only 38 are land-points affected by land-surface modification.

Table 9.2 summarises the changes in the local surface hydrological processes based on the 38 land-points only following deforestation. In this case, only the evaporation changes are significant at the 5% level, although the July precipitation change almost reaches this level. Again, statistical significance is considered in greater detail when discussing the spatial maps.

There is a reduction in the total evaporation (from both the canopy and soil). Consequently, there is a notable reduction in the total evaporation on the land (i.e., -13% in January and -5% in July), though only small changes (i.e. +0.3% in January and +0.2% in July) are indicated when evaporation is averaged for the whole region (Table 9.1). The increase in evaporation for the whole region in January is, therefore, influenced by the increase in evaporation over the sea which opposes the reduction over the land. The most likely reason for the increase in evaporation over the sea in January is a strengthening of the surface wind associated with the winter monsoon over this region (discussed further later). The strengthening of winds, however, seems not to influence the reduction in evaporation from the canopy and the soil in January. In July, although the evaporation from the land (not shown) is reduced, the whole region (Table 9.1) indicate an increase in evaporation following deforestation.

The partition presented in Table 9.2 indicates clearly that the imposed land surface changes could effectively account for the change in the hydrological processes. One important conclusion from this analysis is that the effect of increased surface albedo is cancelled by decreased evaporation over the land area, thus explaining the small changes in surface temperature over deforested region. Further discussion of these processes in connection to the surface and atmospheric energy budgets will be given later in Section 9.5.

9.4.2: Spatial pattern of local to regional-scale changes

An assessment of the local to regional climatic changes over Southeast Asia in January and July, the two representative months for the winter and summer monsoon, is given in Figures

9.10 to 9.14. These figures show selected fields as differences with superimposed areas where the Student's *t*-statistic for a grid-point difference is significant at the 5% level.

b. Changes in ground surface temperature

Ground surface temperature over the deforested region is significantly increased for both January and July (Figure 9.10a and b). Point-to-point changes can be clearly seen from the associated grid-point values for both months in Figure 9.10c and d. In January (Figure 9.10c), the greatest increase in ground surface temperature occurs over South Thailand where the change is over +2^{o}K. In July (Figure 9.10d), the largest increase is over South New Guinea where the change is also above +2^{o}K. Most of the grid-points in the deforested region, however, show ground surface temperature changes varying from slightly above zero to 1^{o}K for both months, resulting in the monthly averaged climatic changes for the deforested grid-points as presented earlier as +0.6^{o}K and +0.5^{o}K for January and July, respectively.

The changes in ground surface temperature outside the deforested region within neighbouring areas in January (Figure 9.10a), overall, are not statistically significant, though a small area over the north-east of the India indicates a significant increase in temperature slightly over +0.5^{o}K. In July (Figure 9.10b), the changes over the northern part of Australia indicate a significant decrease in ground surface temperature, with a peak change of over -2^{o}K, implying a colder winter over this region.

b. Changes in precipitation

In January, the changes over the study area are mixed between an increase and decrease in precipitation, but only a few changes are statistically significant. There are two areas in the deforested region, however, where the changes are significant at 5% level. The major changes are the increase of between 60mm to 100mm per month in the area to the south of Sumatra and Borneo covering most of Java and extending west to a region over the Indian Ocean around 10^{o}S. Most likely, the increase of precipitation is connected with the intensification of the ITCZ over the same area, which will be discussed later. The second major area is over the north of Philippines where the increase is over 80 mm per month. Another major change is a significant decrease over the west Pacific Ocean to the east of Philippines.

In July, mixed changes are also observed over the study area with a few areas statistically significant (Figure 9.11b). Over small parts of Peninsular Malaysia and Sumatra, the total monthly precipitation decreases between 40mm to 60mm per month and the change is significant at the 5% level. In the areas over South Borneo and the Southeast Philippines, however, there is a significant increase of total July precipitation by between 20mm to 60mm. The major changes in precipitation in July occur in the middle of the Bay of Bengal when there is an increase of between 100mm to 140mm per month, which is statistically significant at 5% level. This increase in precipitation is most likely connected with the intensification of the summer monsoon over this area, which will be discussed later.

c. Changes in surface evaporation

As expected (cf. Table 9.1), total surface evaporation decreases over most land areas affected by deforestation in both months, though only few changes are statistically significant. Although fewer are statistically significant, most likely due to larger inter-annual variability, the magnitude of change over oceanic areas seems greater than over the land areas.

In January, the largest reduction in surface evaporation (-30mm to -40mm) occurs over Indo-China and Thailand (Figure 9.12a). There is also a smaller reduction in evaporation (-10mm to -20mm) over the west coast of Peninsular Malaysia and adjacent Sumatra which is statistically significant.

In July, the changes seem to be less marked than in January. The majority of the deforested grid-points seem to experience a negative change in surface evaporation but few changes are statistically significant (Figure 9.12b). Small areas in eastern Borneo, Celebes and New Guinea experience a significant decrease in surface evaporation of about 10mm per month. The most significant change in surface evaporation during this month appears to be outside the deforested region over the ocean. One such area is to the north of Australia where the increase in surface evaporation exceeds 40mm per month. Another area is in the southern Bay of Bengal just to the east of Indian subcontinent where the surface evaporation increases significantly by at least 20mm per month. Again, the most likely reason for this change is due to the increase in atmospheric upwelling associated with the strengthening of the monsoon winds.

d. Changes in the 1000 hPa relative humidity

Despite the high amount of moisture provided by the continuous "fuelling" from the adjacent Pacific Ocean, Indian Ocean and South China Sea, changes in relative humidity over the study area following deforestation are quite marked.

In January, the significant reduction of more than 10% in relative humidity over southern Thailand/Indochina and near Myanmar can be directly associated with the land surface change over the area (Figure 9.13a). This reduction is positively correlated with both the reduction of surface evaporation and the increase of ground surface temperature over this area. Another area with a significant change in relative humidity in January is located along the 5^{o}S latitude, extending from the Indian Ocean to the north of Australia, associated with the common location of the ITCZ during this time of the year.

In July, the changes in relative humidity over southern Thailand/Indochina and near Myanmar seem to be less marked than in January. However, the majority of the deforested grid-points seem to experience a negative change in relative humidity with many changes are statistically significant (Figure 9.13b). Significant reductions in relative humidity can be seen over the land grid-points in North Borneo, south Sumatra, Celebes and New Guinea, and the reduction is positively correlated with both the reduction of surface evaporation and the increase of ground surface temperature over this area.

9.5: Causes of the Climatic Changes: Overview of the Processes

In this section, an attempt is made to account for the changes in regional climate resulting from deforestation by considering three particular aspects of the simulation: changes in surface net radiation; the radiative energy budget; and hydrological processes. In the following subsections, these three changes are discussed in turn.

9.5.1: Changes in the surface net radiation

Here, based on the ensemble results, the extent to which changes in surface characteristics following deforestation have contributed to changes in the radiation balance at the surface is assessed. Maps of the changes in surface radiation for the net down surface short-wave and long-wave radiation as well as the resultant net down surface radiation are presented. This analysis shows how significant the extent of changes in the surface radiation following deforestation are in the study area. As will be seen, many of the grid-point differences are significant at the 5% level.

a. Net down surface short-wave (SW) radiation

Figure 9.14a and b show the pattern of changes in net down surface short-wave (SW) radiation following deforestation in January and July, respectively. Though only a few areas indicate changes which are statistically significant, the imposed increase in surface albedo after deforestation seems to have produced reduction in the net down SW radiation absorbed at the surface over the majority of areas in the deforested region in both January and July.

The most interesting result here is that the largest changes in net down SW radiation in January do not show right over the deforested land area in Southeast Asia. Rather, the areas with the largest, statistically significant changes are all located over the neighbouring oceans: first, over the Indian Ocean (-40 Wm^{-2}) near the mean position of ITCZ during this time of the year; second, over the north of South China Sea near the coast of China (-25Wm^{-2})

associated with the northeast (winter) monsoon; and third, over the western Pacific Ocean (+20 Wm^{-2}). Large changes in short-wave radiation absorbed at the surface in July (Figure 9.14b) also occur remotely though they are not as marked. The most likely reason for this behaviour is the cloud radiative forcing and the balance between clear- and cloudy-sky conditions.

b. Net down surface long-wave (LW) radiation

Figures 9.15a and 9.15b show the pattern of change in net down surface long-wave (LW) radiation following deforestation in January and July respectively. As with SW radiation, only a few areas indicate changes which are statistically significant at 5% level. However, it is shown that in general the net down LW radiation received at the surface over the land area in the deforested region for both January and July is reduced. This reduction is closely related to the increase of ground surface temperatures at land areas following deforestation. The higher surface temperature will enhance the increase in net LW radiation lost from the surface, and, therefore, reduce the net down surface LW radiation. In addition, because of feedbacks from cloud and moisture radiative forcing, remote changes of the LW radiation outside the deforested region are also observed.

c. Net down surface radiation (resultant)

Figures 9.16a and b show the pattern of changes in net down surface (SW + LW) radiation following deforestation in January and July, respectively. The changes in the resultant net down surface radiation are closely related to the changes in the net down surface SW radiation. The reason for this is that, in general, the change in the net down surface SW radiation is much larger than the change in the net down surface LW radiation. A discussion of this net radiative change is given in more detail in the next subsection.

9.5.2: Changes in the radiative energy budget

Analysis of the radiative energy budget provides a basis for understanding how factors such as atmospheric moisture are related to the surface and the atmospheric energy budget. Previous studies (e.g. Gutowski *et al.*, 1991; Boer, 1993; Zhang *et al.*, 1996a) have shown that both the surface energy budget and the atmospheric energy budget determine how the atmospheric circulation behaves over a certain region. The changes in these energy budgets could alter both the atmospheric circulation and the water cycle.

9.5.2.1: Surface radiative balance

Table 9.3 summarises the net radiative balance at the surface averaged over the study area. This is to objectively determine the monthly (January and July) surface radiative energy budget for the control and deforestation over the region and, therefore, to assess how the relevant processes behave following deforestation. The net down radiation over the land (solid) surface as well as over the open sea within the study area are given in order to see how the radiative balance differs between the two different surfaces and how they each contribute to the overall surface radiative balance in the deforested region.

The net down solar fluxes at the surface in both January and July are reduced following deforestation. Two factors are causing the reduction: first, the amount of SW flux reflected from the surface is increased due to the imposed surface albedo increment; and second, the amount of the SW flux reaching the surface is reduced as the total cloud amount increase (see Table 9.3). Both months are characterised by a similar increment in surface albedo following deforestation. The reason for the larger reduction in the surface solar flux in January is due to the total cloud which increases more in January than in July. (The higher increment of both convective and large-scale precipitation in January than in July following deforestation also corresponds to the relative change in cloud amount.)

The net down solar flux on the land (solid) surface is also reduced following deforestation in both January and July, with the reduction in January more than in July. As above, this larger reduction in January is caused by the imposed increase in land-surface albedo and reinforced by the increase in total cloud amount over land. In July, however, there is a competing effect between the change in land-surface albedo and the change in cloud amount. A reduction in the cloud amount in this month should cause an increase in the downward component of solar radiation reaching the land-surface. However, the net down solar radiation absorbed at the surface is still reduced following deforestation because the effect of the imposed increase in land-surface albedo dominates over the effect of the reduction in the cloud amount.

The increases in cloud amount over ocean in both months following deforestation are solely responsible for the reduction in net down solar flux on the open sea since there is no imposed change on the surface albedo at the ocean-points. Over the open ocean, the larger increase in the amount of cloud in January corresponds with the greater reduction in the net down solar radiation.

The negative values of the net down surface LW radiation in all cases indicates that the LW downward component is weaker than the upward component, implying an increase in net LW radiation lost from the surface. In both January and July, the net down LW radiation at the surface becomes more negative following deforestation. From theory, this may be caused by both or either of the following: (i) the amount of LW flux re-radiated from the surface is increased due to the increase in surface temperature; and/or (ii) the amount of LW radiation that is absorbed by the atmosphere and re-radiated back to the surface is reduced due to a decrease in cloud amount and atmospheric moisture in the lower layers of the atmosphere. A competing effect may occur between the above two processes. In the following paragraphs, how the change in the temperature and cloud amount affecting the LW radiation budget over the whole study area, the land surface and the sea surface is discussed separately.

Over the whole study area, the amount of cloud is effectively increased following deforestation in both months. With more cloudiness, therefore, more LW radiation is re-radiated back to the surface from the atmosphere. At the same time, the surface temperature also increases which leads to an increase in the LW flux re-radiated out from the surface. As the net down surface LW radiation in both months becomes more negative following deforestation, it means that the increase in the upward component of the LW radiation is larger than the increase in the downward component. The effect from the surface temperature increase, therefore, dominates over the effect from the increase in the cloud amount. As the sea surface temperature is not changed between the control and deforestation cases, the dominant effect on the regional surface LW radiation budget is contributed by the land-surface temperature change.

The contribution from cloud to the net LW radiation budget over the land surface following deforestation is different between January and July (Table 9.3), though both months experience an increase in land surface temperature (Table 9.1). In January, the effect from the increase in ground surface temperature is more dominant than the effect of the increase in the total cloud amount. The net LW radiation lost from the land surface is, therefore, increased in January. In July, on the other hand, the increase in the net LW radiation lost from the land surface is caused by both the increase in the upward component of the LW radiation (as the land surface temperature increases) and the reduction of the downward component of the LW radiation (as the cloud amount is reduced) on the other.

Differing from the land surface situation, the net LW radiation lost from the open sea becomes less in both months following deforestation. The LW flux emitted from the sea surface should not have changed as the sea surface temperature is kept constant in each month for both the control and deforestation cases. The increase in cloud amount, therefore, must cause the downward component of the LW radiation over the sea surface to increase while the upward component is kept constant.

Overall, the monthly average surface radiative energy budget indicates that the net radiation absorbed at the surface for the whole region, land surface and open sea is less in both January and July following deforestation (Table 9.3). At the surface, a reduction in the net radiative energy absorbed by the surface will lead to a reduction in the energy contributed together by the latent heat flux and sensible heat flux. The reduction, therefore, should diminish the turbulent exchanges of heat and moisture between the surface and the atmosphere, and thus reduce cloud formation and precipitation locally. This effect has been demonstrated well in the case of the Amazonian deforestation by previous GCM simulations (e.g. Henderson-Sellers *et al*., 1993; McGuffie *et al*. 1995; Zhang *et al*., 1996a). In the context of Southeast

Asian deforestation, however, the effect does not reduce the amount of cloud and precipitation over the whole region. Rather these quantities, overall, increase following deforestation (cf. Table 9.1 and 9.3). Over the ocean, as can be deduced from the two tables, cloud and precipitation increase in the two months. Over the land, the amount of cloud increases in January but decreases in July (though precipitation over the land decrease for both months), and evaporation reduces in both months following deforestation. It seems that the reduction of net solar radiation absorbed by the surface only affects land areas in the classic manner during summer (July). Other processes must complicate the picture in January. All the above pattern of changes, between the two seasons as well as between the land and ocean areas, signals the existence of external factors from outside of the deforested region which might influence the processes and changes in the deforested region itself. This possibility is investigated later.

9.5.2.2: Atmospheric radiative balance

Table 9.4 shows the regional monthly averages (January and July) of the atmospheric radiative energy budget from the control and deforestation experiments as well as the difference between the deforestation and the control results over the Southeast Asian study area. Seasonally, the incoming solar radiation at the top of the atmosphere over Southeast Asia for the both control and deforestation cases is higher in summer (July) than in winter (January). At the same time, the net solar radiation at the top of the atmosphere and the outgoing SW radiation at the top of the atmosphere are also greater in July than in January. Following deforestation, there is an increase in the outgoing SW radiation at the top of the atmosphere for both January and July, with the increment in January more than in July. The increment of the outgoing SW radiation at the top of the atmosphere corresponds to a decrease in the net radiation input at the top of the atmosphere, with the decrease in January more than in July. The change in the radiative budget at the top of the atmosphere is caused by an increase in albedo as cloud amounts increase. The higher increment of the outgoing SW radiation (or higher reduction of the net solar radiation at the top of the atmosphere) in January corresponds to the greater increment of the cloud amount during this winter monsoon month, as confirmed by the simulated increase of the total cloud amount which is higher in January (+2.2 %) than in July (+0.4 %)

The outgoing LW radiation at the top of the atmosphere is higher in January than in July. The principal factor that controls the amount of outgoing LW radiation at the top of the atmosphere is the amount and type of cloud as well as the moisture content in the atmospheric layers. Lower and warmer clouds such as "bad weather cumulus" and "cumulonimbus", with an abundance of moisture in the atmosphere, re-radiate less LW radiation energy to outer space because of the enhanced greenhouse effect in the lower atmospheric layers. Water vapour is the most efficient greenhouse gas in the lower atmosphere. With more clouds and water vapour, more LW radiation is re-emitted downward to the surface causing a warming effect in the lower levels of the atmosphere. From the indicated amount of outgoing LW radiation at the top of the atmosphere, we may deduce that more clouds of the above type are available in July compared to January in the study area. In the tropics, the majority of clouds formed are of the convective type which contribute the bulk of the overall cloud amount in this region. This corresponds to the total cloud amount in the region as simulated by the ensemble simulations (Table 9.4). Following deforestation, there is a decrease in the outgoing LW radiation at the top of the atmosphere for both months, with the reduction in January more than in July. The reduction in the outgoing LW radiation at the top of the atmosphere also corresponds with the increase of the cloud amount for the two months, again the increase in January is more than in July.

In summary, then:

(i) The positive changes of the outgoing SW radiation at the top of the atmosphere (or the negative changes of the net solar radiation at the top of the atmosphere) and the negative changes of the outgoing LW radiation at the top of the atmosphere in both months are controlled by the positive changes in the cloud amount.

(ii) The increase in the outgoing LW radiation from the land surface and the decrease in the outgoing LW radiation at the top of the atmosphere combine to give a net increase in the radiative energy absorbed by the atmosphere in spite of the increase in surface albedo. The net radiative energy for heating the atmosphere, therefore, is increased by the changes in the cloud radiative forcing and the LW radiation

Finally, it is noted that the increase in the radiative energy absorbed by the atmosphere could lead to an increase of net energy available for supporting the regional atmospheric circulation. The atmosphere over the region becomes more energetic following deforestation. For example, the increase in precipitation over the whole region (Table 9.1) following deforestation, as simulated by the ensemble suggest a strengthening of the monsoon circulations over the region in both seasons.

9.5.3: Changes of hydrological processes

According to Entekhabi *et al*. (1992) and Brubaker *et al*. (1993), when considering a large continental region, the water recycling over a certain area is determined by: (i) how much precipitated water can be returned to the atmosphere as surface evapotranspiration; and (ii) to what extent the surface-evaporated water can be held in the atmosphere over this area and then become precipitation to sustain the regional hydrological cycle.

Figure 9.17 shows a simplified model of atmospheric moisture fluxes over a land region (modified from Brubaker *et al*., 1993). The horizontal arrows indicate the advective flux of water vapour into and out of the atmospheric control volume; W is the amount of water vapour contained in the air as it moves through the control volume; E is the net evapotranspiration from the underlying land surface; P is the net precipitation onto the land surface. The two sources of precipitation are indicated by the two branches that join to form the larger arrow labelled P: P_m is precipitation of local (evaporative) origin and P_a is that of advective origin. The arrow labelled E splits into two branches, indicating that a certain fraction of the locally evaporated or transpired water is not returned to the land surface as precipitation but joins the atmospheric vapour reservoir and is advected out of the control volume. By definition, F^+ contains only advective moisture, and F^- contains the advected moisture that remains after P_a is removed, as well as moisture of local origin. R_s^+ and R_d^+ contain the surface and the deep ground (sub-surface) water runoff into the control volume, respectively, and R_s^- and R_d^- are the surface and the deep ground (sub-surface) water runoff that remain after E is removed.

For the Amazon Basin, the conceptual model illustrated in Figure 9.17 (Brubaker *et al.*, 1993) can be successfully simulated in the GCM (e.g. McGuffie *et al.*, 1995; Zhang *et al*. 1996a,b). The dominant convergent airflow over this region results in a major proportion of the surface-evaporated water remaining in the basin and only part of it is transported out of the basin. Zhang *et al*. (1996a) show the hydrological processes represented in the control experiment of CCM1-Oz (cf. Chapter 4). As reported, the annual averaged results over the whole Amazon Basin show that about 41% of the total precipitation is held and recycled into the atmosphere by rainforest transpiration and ground soil evaporation and 24% of the precipitation is intercepted by the tropical forest to be evaporated back to the atmosphere. Consequently, 65% of the total precipitation is held by the rainforest and returned into the atmosphere local to the rainfall occurrence, and the remaining 35% of the total precipitation is lost as surface and sub-surface runoff, which must be balanced by atmospheric transport. Zhang *et al*. (1996a) also successfully illustrated the changes in the hydrological cycle over the Amazon Basin after deforestation by giving percentage changes for each of the components.

It is difficult to investigate the hydrological cycle in similar way to the Amazon Basin in the context of the Southeast Asia region. Unlike Amazonia, the majority of land areas in Southeast Asia comprise of a small portion of continental areas, and peninsular and islands which are largely interspersed by sea-areas as well as surrounded by the adjacent tropical ocean (recall that only 38 grid-points in the study area are land out of the total of 180 grid-points). A model that only couples the water balance of continental landmasses and the overlying atmosphere obviously cannot represent the hydrological cycle in Southeast Asia. The main reason for the failure is that the total water lost as surface and sub-surface runoff, which must be balanced by atmospheric transport, cannot be represented in the conceptual model as illustrated in Figure 9.17. Over the open ocean, the total water lost as surface and sub-surface runoff is not calculated, resulting in a discontinuity of runoff over the region.

While a direct comparison cannot be drawn, it is useful to compare various characteristics of the hydrological cycles over the two regions using Figure 9.17. For the Amazon Basin, the term P_m, which represents precipitation of local (evaporative) origin is more dominant than the term P_a, which represents precipitation of advective origin. In another words, the terms F^+ and F^- that indicate the advective flux of water vapour into and out of the atmospheric control volume in the basin are not as dominant a factor as the other terms in controlling the changes in the hydrological processes in this region. For Southeast Asia, on the other hand, the term P_a is more important in maintaining the hydrological cycle and controlling changes in the hydrological processes following deforestation. The advective flux terms are more important than the other terms given in controlling changes in the hydrological processes. The different sign of changes in the moisture convergence between January and July (Table 9.2) for the 38 land-points within the deforested region, for example, is the results of the changes in the strength of those advective flux terms. The importance of P_a in the context of deforestation in Southeast Asia can be illustrated by the following two examples, for the whole region (study area) and for the land area (deforested region).

(i) Over the whole region or study area (Table 9.1): Total evaporation and precipitation are increased in both January and July following deforestation. A competing effect, however, may occur in both months when both evaporation and precipitation are increased. The increase in precipitation may be due to a slight reduction in the term P_a with magnitude less than the increase in evaporation. Alternatively, P_a is either unchanged or may also increase following deforestation. There is more precipitable water contributed by the advective moisture flux (F^+) and, therefore, the increase in precipitation is largely contributed by the term P_a in the hydrological process.

(ii) Over the land area or deforested region (Table 9.2): Although the total evaporation is reduced in January following deforestation, yet total precipitation still increases since the precipitation is largely contributed by the term P_a. A competing effect, however, may occur in July when both evaporation and precipitation are reduced following deforestation. The decrease in precipitation in July may be due to a slight increase in the term P_a with magnitude less than the decrease in evaporation, or P_a is either unchanged or also decreases following deforestation.

Clearly, the alteration in regional water recycling and hydrological processes as a result of the changes in the surface vegetation is not the only factors that control climatic change in the context of Southeast Asia deforestation. Rather, the changes in the advective flux of water vapour into and out of the deforested region are equally important.

9.6: Conclusions

The control ensemble results have been evaluated by comparison with the mean observed climatology defined by the NCEP/NCAR Reanalysis Data or the observed surface dataset from Legates and Willmott. It has been shown that the regional scale climates of Southeast Asia are satisfactorily simulated by the ensemble control experiment. In both January and July, the regional patterns of the ITCZ and monsoon trough dominates and these are adequately represented by the model used here. At the grid-point scale, however, the comparisons reveal various discrepancies with positive as well as negative biases in certain important surface variables. The most likely reason for the discrepancies is the unresolved sub-gridscale processes. The stronger north-south temperature gradient with cooler continental interior in January results in an overestimated winter monsoonal wind over the study area. This is a major discrepancy in the control ensemble.

Replacing tropical rain forest with grassland leads to three primary changes in the land surface properties: (i) an increase of surface albedo directly causes an alteration in the surface net radiation and/or indirectly influences the cloud feedback mechanism; (ii) reductions in the leaf and stem area lead to a decrease in the water holding capacity of the vegetation, hence, reduce evaporation of the intercepted precipitation and decrease transpiration; and (iii) a shorter surface roughness length leads to a weaker surface friction due to a smoother grassland, hence, strengthens surface winds. Overall, the regional climatic impacts over deforested Southeast Asia and neighbouring regions indicate there are significant responses from the imposed land surface modification. However, in the context of the Southeast Asian single region deforestation undertaken in this study, the results show rather different climatic changes compared to previous multi-region deforestation studies. The surface temperature and precipitation as well as cloudiness are all increased over the region following deforestation.

Figure 9.18 shows a schematic illustration of processes occurring following Southeast Asian deforestation and summarises the finding discussed in this chapter. The most prominent changing features after removing Southeast Asian forest is the increase in moisture convergence over the study area as the result of increasing in cloudiness and precipitation, overall, over the study area. The most likely cause for the increase in cloud amount and precipitation is the enhanced monsoonal flow over the study area. It is suggested that the existing monsoon flow can be strengthened either by the direct effect of the decrease in surface roughness length or indirect effect of the increase in surface temperature, both over the deforested region. The increase in surface temperature over deforested region is the result of the decrease in surface evaporation (direct effect) and the decrease in surface net radiation (indirect effect) when the surface albedo is increased. At the same time, the increase in surface evaporation over the study area leads to a net increase in the atmospheric energy budget. Thus the regional atmospheric circulation is strengthened and more water vapour is delivered into the study area (increase in moisture convergence). For deforestation in the Amazon Basin, as illustrated in Figure 4.1 (Chapter 4) and description by Zhang *et al*. (1996a,b), the processes are different which produce a different climatic change scenario for this region.

By mass continuity or conservation of energy, any change of the air column above the deforested region, either in terms of mass or energy properties, should cause changes beyond the areas of disturbance. Hence, it is very likely that the deforestation could lead to a disturbance in some aspects of the general circulation especially the Walker and Hadley cells. These aspects of circulation will be examined in the next chapter in order to extend our investigation on the effect of deforestation on the large-scale monsoon circulation.