CHAPTER 5: REGIONAL CLIMATE SENSITIVITY TO TROPICAL DEFORESTATION

5.1: Introduction

The atmospheric circulation over Southeast Asia region is largely controlled by the two Asian monsoons, called the winter and summer monsoons. The two monsoon seasons present a useful framework for any analysis of the impacts of tropical deforestation in Southeast Asia. The changes in regional climate in this region, therefore, can be described in terms of the monsoon circulation changes.

This chapter provides an overview of the Southeast Asia regional climate sensitivity to tropical deforestation. The regional climatic changes are directly caused by changes in regional-scale circulation. The aim of this analysis, therefore, is to present an assessment of the sensitivity of the regional-scale circulation over Southeast Asia and the nearby areas to tropical deforestation. The assessment is mainly based on the results of a previous experiment using the NCAR Community Climate Model, incorporating the Biosphere-Atmosphere Transfer Scheme (BATS1e) undertaken by Zhang *et al*. (1996a, b). This coupled version is known as CCM1-Oz (refer Table 4.2, Chapter 4). The experiment consists of a 25-year "control" integration and an 11-year "perturbation" integration. In the perturbation experiment, tropical forests for the three regions, the Amazon Basin, Southeast Asia, and tropical Africa, were all together converted into grassland. Based on the same experimental results used in this analysis, Zhang *et al.* (1996*a*) reported the impacts of tropical deforestation in term of process analysis of local climatic change. In their companion paper (Zhang *et al.*, 1996*b*), they also evaluated the regional to global-scale changes resulting from deforestation. In neither paper, though, is detailed consideration given to the contribution of Southeast Asian deforestation to the overall perturbation nor to affects on the Southeast Asia climate and atmospheric circulation. A re-analysis of the experimental results was, therefore, undertaken.

Based on a new analysis of Zhang *et al*.'s results [available on CD-Rom with book by McGuffie and Henderson-Sellers (1997)], the regional-scale circulation impacts following deforestation are determined in this chapter with reference to the change in the vertical velocity, divergence and vorticity fields for the two monsoon seasons. A modification to vertical motion, atmospheric convergence or vorticity patterns might signal a disturbance to aspects of the general circulation, especially the Walker and Hadley cells which derive their energy from warm and moist air ascending in equatorial regions. The circulation response for the two monsoon seasons is discussed by referring, in most cases, to the changes in January and July. The conditions during January are considered representative for the whole of the winter monsoon season that extends approximately from November to March when this region is under the dominance of northeasterly winds. The month of July is considered typical of conditions during the summer monsoon season that normally lasts from about June to September when the whole region is under the dominance of southwesterly winds.

This discussion contributes to the development of the experimental design used in the modelling studies discussed later in this thesis.

5.2: Control Climate Simulation

Model evaluation of general circulation models (GCMs) is a necessary requirement because it highlights biases in the model and provides information with which climate change prognostications can be assessed. Before considering the results of Zhang *et al*.'s deforestation experiment, it is necessary to assess how accurate their control experiment simulating present-day climate conditions. For the purpose of the validation, both the National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) Reanalysis Data (Kalnay *et al.*, 1996) and the well-known Legates and Wilmott (1990a,b) global climatology of mean monthly precipitation and surface air temperature were employed. Regridding, however, was applied on these data to make their resolution similar to the model data to be validated.

The NCEP/NCAR Re-analysis Data for the period of 1976-96 (21 years) were used to derive the mean climatology of the appropriate variables used in the validation. These reanalysis data are produced by a frozen state-of-the-art global data assimilation system with the operational model at the NCEP using a complete observation database. The system has been designed with advanced quality control and monitoring components. The database used has also been enhanced with additional sources of observation not available in real time during operations. In preparing the observation data, the reanalysis involves the recovery of land surface, ship, rawinsonde, pibal, aircraft, satellite and other data, quality controlling and a dynamically consistent data assimilation system. This reanalysis has provided a good horizontal 2.5^{o} of latitude by 2.5^{o} of longitude grid and vertical resolution of 17-levels for various variables. In term of reliability, the output variables from these data are classified into different categories depending on the relative influence of the observational data and the model. To avoid variables that are solely derived from the model fields forced only by the data assimilation, the variables used for this validation exercise are only from a category that indicates a direct influences, partly or fully, from the observational data: pressure vertical velocity near the surface, u- and v-wind at 1000 hPa level, and mean sea level pressure.

One other comparable set of reanalysis data is available from the European Centre for Medium-Range Weather Forecast (ECMWF) named as the ECMWF Reanalysis (ERA). This set is generated using the ECMWF forecast model by a special version of the ECMWF operational data assimilation system. At the time selection decisions were made, this data set was considerably shorter than the NCEP/NCAR data set and, for this reason and for reasons of ease of access, the NCEP/NCAR data set was chosen. There has been no full comparison of the two sets in terms of accuracy and reliability. In both cases it should be noted that the use of an atmospheric model to fill gaps in the direct observational data means that certain aspects of these reanalysis data sets may not truly reflect real world behaviour (a possible example of this is given in Chapter 10).

The global climatology of mean monthly precipitation (Legates and Willmott, 1990a) and mean monthly surface air temperature (Legates and Willmott, 1990b) were also used for the model validation. As reported by Legates and Willmott, both types of data were originated from ten sources, screened for coding errors, and redundant station records were removed.

The precipitation climatology was compiled using a database obtained from traditional land-based gauge measurements and shipboard estimates consist of 24,635 spatially independent terrestrial station records and 2,223 oceanic grid-point records. The raw precipitation data have been corrected to remove systematic errors caused by wind, wetting on the interior walls of the gauge, and evaporation from the gauge. The temperature climatology was compiled using a database obtained from terrestrial observations of shelter-height air temperature and shipboard measurements consist of 17,986 independent terrestrial station records and 6,955 oceanic grid-point records. Both the global climatology data have been interpolated to a 0.5^o of latitude by 0.5^o of longitude grid using a spherically based interpolation procedure. With the extensive quality control imposed in their preparation, these observational data can be regarded as sufficiently reliable for the validation purpose.

Other than the above reasons, another reason for selecting the Legates and Willmott global mean monthly precipitation and temperature data set is due to their spatial coverage, which include both the land and ocean. Since the decision to select the Legates and Willmott precipitation data set was made, a new monthly blended precipitation data set for land and ocean for extended tropical region of 30^oN to 30^oS from 1974 to 1994 has become available and has recently been documented (Doherty *et al.*, 1999). This data set represents an improvement on previous precipitation data set due to the inclusion of ocean component which is derived from satellite outgoing longwave radiation in addition to the earlier land component of Hulme (1992). In future work, this new data set should be used. Unfortunately, its lack of availability as the first analyses were undertaken in the present study meant that it was not possible to use this new data set. The main use of these data is in testing the model's ability to simulate present-day conditions and, in this context, the difference between the model fields and the observed data is considered likely to overwhelm any difference between the data sets.

All data set had to be brought to a consistent spatial resolution for inter-comparison. The question is whether to treat the GCM output data in this study as grid points (point quantities) or grid boxes (areal quantities) since the two different treatments involve with different interpolation techniques that need to be used. Interpolation, therefore, is constrained by whether the original data are point or areal quantities. A grid-point approach would represent a sampling of the expected climate surface, as an estimate at the centre of a grid-box. On the other hand, a grid-box approach would represent a smoother climate surface because of a larger area represented by a GCM value in a box. As discussed by Skelly and Henderson-Sellers (1996), there is no definitive evidence to clearly support one approach over the other among users of GCM data and according to them one can adopt either approach at least until a clear evidence is defined. On this issue, however, Skelly and Henderson-Sellers indicate a strong point by referring to mechanics of a GCM itself although according to them the problem cannot be answered fully. According to them, though model's derivation, for example on convection, radiation, and precipitation and surface boundary parameterization such as albedo, topographic height, and roughness are implicitly areal in implementation, GCMs using either finite difference or spectral methods produce point values rather than areal quantities as output. In view of the above, a grid-point approach has been chosen in this study when interpreting the model output data. With the approach, interpolation to the both model output and observation data has been undertaken using bilinear interpolation procedure to achieve consistency in their spatial resolution. Hence, a comparison of the GCM simulated data with observed climatologies at large spatial scales could be undertaken.

5.2.1: Seasonal cycle of regional vertical velocity

The seasonal cycle of the area average of the vertical velocity near the surface could indicate how a favourable atmospheric instability changes with seasons. This seasonal change over Southeast Asia region is presented here for both the NCEP/NCAR Reanalysis Data and the Zhang *et al*. (1996a, b) control GCM simulation, purposely to identify any possible bias in the GCM. If a bias is found, a necessary precaution should be taken when interpreting the results of the deforestation experiment from the GCM.

Figure 5.1(a & b) represent the seasonal cycle of average vertical velocity for an area bounded by15^{o}S to 30^{o}N and 90^{o}E to 130^{o}E, covering most of the Southeast Asia, for both the NCEP/NCAR Reanalysis Data and the control experiment. In addition, an area average for the equatorial Southeast Asia region bounded by 15^{o}S to 15^{o}N and 90^{o}E to 130^{o}E is also given in Figure 5.1(c & d). Each is presented with three regional averages: first, for an area to

the east of 110^{o}E (the eastern region); second, to the west of 110^{o}E (the western region); and third, is the mean of the two regions that represents the whole region.

As shown by the NCEP/NCAR Reanalysis Data (Figure 5.1a), Southeast Asia is, overall, under the influence of an upward vertical velocity throughout the year. On average, the whole region seems to follow a similar seasonal pattern as the western region, suggesting that seasonal variability of the vertical velocity over Southeast Asia overall be largely controlled by the western region. The western region (or the whole region alike) has a relatively stronger upward motion during June to September (summer monsoon months) and a weaker strength during November to March (winter monsoon months). An opposite effect, however, is shown by the eastern region when the strength of an upward motion is slightly greater during the winter monsoon compared to the summer monsoon. The strength of upward motion for the eastern region is relatively lower than the western region for most of the months, except December. These conditions, therefore, clearly indicate that the summer monsoon produces a stronger influence than the winter monsoon for the western region, but the winter monsoon contributes a slightly higher influence than the summer monsoon for the eastern region.

The control simulation (Figure 5.1b) indicates that, in the model, the whole of Southeast Asia is not controlled by upward vertical motion throughout the year, but only for the summer monsoon months. Although the month-to-month change in upward vertical velocity during the summer monsoon period is greater in the control run compared to the NCEP/NCAR Reanalysis Data, the mean for the overall season, however, seems not to show much difference between them. During the winter monsoon (November-March) and both the inter-monsoon periods (April-May and September-October), the control simulation has a markedly different vertical velocity field compared to the NCEP/NCAR Reanalysis Data. The control simulation produces a downward motion over the whole of Southeast Asia region. In fact, the eastern region has a downward motion for the whole year and the magnitude of downward motion for the winter and inter-monsoon periods are greater than during the summer monsoon. This is most likely due to an underestimation of the strength of the ITCZ (inter-tropical convergence zone) or surface trough system, normally associated with a strong vertical motion above it. Seasonally, the ITCZ shifts from the north (around 25^{o}N) during the summer monsoon to the south near the equator during the winter monsoon. Overall, the control simulation seems to underestimate the upward motion over the eastern region during all the seasons and over the western region during the winter and inter-monsoon seasons.

The area average for both the NCEP/NCAR Reanalysis Data and the control simulation are also calculated for a smaller area, representing the equatorial Southeast Asia region. Figure 5.1c indicates this area for the NCEP/NCAR Reanalysis Data. On comparison with the whole region (Figure 5.1a), a large reduction of upward vertical velocity is shown for the western region during the summer monsoon. This indicates that the influence of the summer monsoon is very strong over the northwest region, consistent with climatology as the monsoon trough is usually very prominently located around northern India and Bangladesh, extending over Myanmar. The strength of the upward vertical velocity for the eastern region, however, does not change all year round, suggesting that the equatorial eastern region is largely influenced by active troughs or low-pressure disturbances throughout the year, which enhance upward motion.

For the control simulation, a comparison between the area average of the whole Southeast Asia region (Figure 5.1b) and the equatorial Southeast Asia region (Figure 5.1d) is also made. A reduction of the upward vertical velocity for the western region still occurs during the summer monsoon indicating that the northwest region is strongly influenced by this monsoon. This is in agreement with the NCEP/NCAR Reanalysis Data. Despite the overall underestimation of upward motion by the model, the seasonal changes of vertical velocity (Figure 5.1b, d) still provide a meaningful explanation of the seasonality, quite similar to the NCEP/NCAR Reanalysis Data. As shown by Figure 5.1d, for the equatorial Southeast Asia, the western region is under the influence of upward motion in July, which is associated with the active summer monsoon in this month. On the other hand, the eastern region experiences a minimum downward vertical velocity in January, indicating that the month is most unstable, which is associated with the winter monsoon. Overall, as indicated by the mean curve (Figure 5.1d), January and July are the two months when the greatest influence of ascending motion is experienced by the equatorial Southeast Asia region.

In term of the vertical velocity field, it seems that there is no apparent bias in the GCM to the summer monsoon over northwestern Southeast Asia. The strength of the summer monsoon over the said region seems to be correctly modelled by the GCM. The model, however, seems to underestimate the magnitude of upward motion for the eastern region during all the seasons and for the western region during winter and inter-monsoon periods. This underestimation of the strength of the upward vertical velocity, therefore, should be considered when interpreting results of a perturbation experiment by the same GCM.

5.2.2: Climate of Southeast Asia

The seasonal cycle of regional vertical velocity (cf. Figure 5.1) suggests that, in terms of seasonality, the climate of Southeast Asia can be described effectively by the two monsoon seasons, the winter and summer monsoon. In addition, as shown by the above figure, those two seasons can be represented by January and July respectively, the extremes of the cycle. Here, reference is first made to the January and July precipitation and surface temperature fields, to present an overview on the climate of Southeast Asia based on these two variables.

Figure 5.2 shows the precipitation for January and July from observation (after Legates and Willmott, 1990a) and the Zhang *et al*. (1996a, b) model control simulation. In January, it seems that peak precipitation amounts simulated by the model exceed the observed. The model, however, underestimates the precipitation maximum centred over southeast of Philippines near Mindanao. For July, on the other hand, peak precipitation amounts simulated by the model seems to be less than observed over the mainland areas, and the maximum appears to shift northward. The precipitation maximum located over India subcontinent, north of Philippines, equatorial western Pacific, and nearby Celebes (Indonesia) are not produced by the model. Instead, a maximum, which is considered an overestimation, is produced by the model near the east coast of the Malaysian Peninsula over the South China Sea. This positive bias might be the result of misrepresentation of the so-called rain shadow effect of the high mountain ranges of Sumatra by the model.

The simulated and observed surface temperatures in January and July are shown in Figure 5.3. The observed field is plotted using data from Legates and Willmott (1990b). Overall, there is an underestimation of this variable as simulated by the model compared to the

observed field. The model also produces too strong a temperature gradient from north to south, as the continental interior is cooler than observed in both seasons. This is probably the most important discrepancy; the results of temperature changes following deforestation over this region need to be interpreted with extra caution.

The precipitation and temperature over Southeast Asia are strongly related to the surface circulation patterns during the two monsoon seasons. A comparison, therefore, is made for the surface flow patterns between the NCEP/NCAR Reanalysis Data and the model control experiment. Wind vector fields together with resolved components, the zonal (u) and meridional (v) wind, could reveal a more detailed picture regarding the surface circulation over Southeast Asia for the two monsoon seasons. At the same time, an investigation is undertaken to check whether the model shows any bias in simulating the above fields.

(a) The winter monsoon (as represented by January):

During the winter monsoon, the northeasterly surface wind flow brings cold, dry air down over continental (mainland) Southeast Asia (i.e., from the Hong Kong area to Vietnam, Cambodia, Myanmar and Thailand). The so-called dry season is therefore experienced by these mainland areas. In contrast, the maritime (island) Southeast Asia is surrounded by a warm tropical ocean air mass moving from the northeast which causes convection almost continuously in this region (Houze *et al.*, 1981). This maritime area, comprises the Indonesian archipelago (Borneo, Sumatra, Java and other small islands), Malaysia (Peninsular and East Malaysia), and the island of New Guinea. In addition to the great influence of the ocean air mass, the large, mountainous, forested islands of this maritime region induce very strong convective activity, characterised by the wet winter monsoon for this region.

The above general condition is better explained by looking at the mean synoptic chart of surface pressure and wind for the region. There is a strong pressure gradient from north to south resulting in a northeasterly surface wind flow as shown by the NCEP/NCAR Reanalysis Data (Figure 5.4a). The wind vector and pressure plots from the analysis data

clearly define the area of the ITCZ. This zone is also quite well reproduced in the wind vector of the control simulation (Figure 5.4c), showing up as a clear demarcation between the northeasterly and southeasterly trade wind systems. The northeasterly "jet stream" flow located over South China Sea associated with cold surges from the Asian continent high-pressure system is also captured by the model, but seems to be slightly weaker in the model compared to the analysis.

A comparison is also made for the *u*- and *v*-component of wind fields between the mean of the NCEP/NCAR Reanalysis Data and the control simulation (Figure 5.5 and 5.6). As shown by Figure 5.5a, the influence of relatively strong easterly component winds extending from the western Pacific to the Bay of Bengal covers most of the Southeast Asia region to the north of equator. A reverse component of maximum westerlies, however, is located south of equator. Figure 5.6a shows a marked northerly component with the greatest strength located over the South China Sea region. The ITCZs are usually placed at the common region of the easterly-westerly and southerly-northerly interfaces, as indicated in Figures 5.5a and 5.6a.

All the above general features seem to be well produced by the control simulation (Figures 5.5c and 5.6c). Two large discrepancies, however, are detected: (i) the strength of easterly and northerly wind component over the South China Sea is underestimated by the model, making weaker convergence over the area to the south; and (ii) the area over the Bay of Bengal, Andaman Sea and northern part of the Indian Ocean has a stronger easterly component, but a much weaker northerly component, resulting in an overestimation of the easterlies and therefore increasing divergence (or decreasing convergence) to the east. This underestimation of the convergence is, therefore, consistent with the earlier finding in the regional vertical velocity analysis (cf. Sec. *5.2.1*), which indicates an underestimation of vertical velocity by the model for the whole of Southeast Asia during this winter monsoon season.

(b) The summer monsoon (as represented by July):

During the summer monsoon, the southwesterly surface winds bring warm and moist air from the tropical ocean over both the mainland and the maritime area. Overall, throughout the whole region, the interaction of the land, ocean, topography and large-scale atmospheric wind flows produce large-scale precipitation. The main air masses come from the Indian Ocean. In most of continental Southeast Asia, this is the main rainfall season of the year caused by one, or a combination of cyclonic disturbances, convection, orographic lifting and convergence (Nieuwolt, 1981). Any lifting of the air masses, by convection, convergence, and disturbances or by relief, will bring large amounts of rainfall. Southeast Asia, overall, receives its largest amounts of precipitation during the summer monsoon. The combined effects of steep topography and convective uplift can produce vast amounts of precipitation in very short time periods. Nieuwolt (1981) reported that Akyab, on the West Coast of Myanmar, receives a mean rainfall of 1400 mm in July, but actual amounts may be as high as 2000 mm per month. An exceptional condition, however, occurs over the Malaysian Peninsula and its vicinity where this part of Southeast Asia is affected by the rain shadow effect of the high mountain ranges of Sumatra. For this part of the region, the air mass boundary brings only limited amounts of moisture, and furthermore the convergence is of very low intensity. The Malaysian Peninsula, therefore, experiences a relatively dry period during this season.

Again, the above general condition is better explained by looking at mean synoptic situations for the region. Conditions at the surface in July as shown by the NCEP/NCAR Reanalysis Data (Figure 5.4b) include a southwesterly surface wind regime corresponding to a south to north pressure gradient. Dry southeasterlies, originating over the Australian continent, dominate the Indonesian surface weather condition (Sukanto, 1981). These surface flows are quite well captured by the model (Figure 5.4d).

A comparison is also undertaken between the mean analysis and the control simulation for the *u*- and *v*-component of wind fields over the region (Figure 5.5 and 5.6). Overall, the model also satisfactorily simulates the wind components over this region as the main patterns are successfully produced. Detailed investigation at sub-regional scale, however, reveals some differences between the model-produced fields and the observed fields. As shown by the observed zonal wind field (Figure 5.5b), there is an influence of a relatively strong westerly wind component, extending from the Bay of Bengal to the western Pacific. On the other hand, extending from the Pacific toward southern China is a reverse wind component of easterlies. The two opposing wind components produce an area of strong convergence at the interface associated with summer monsoon ITCZ. This convergent feature, however, is not produced by the control simulation (Figure 5.5d), one of the reason for the model underestimation of the upward motion over this eastern region.

Another discrepancy in the control simulation is an overestimation of the maximum westerly component over the Bay of Bengal, and furthermore, the maximum is extended too far to the east. This might be related to misrepresentation of the high mountain ranges at Sumatra and Malaysian Peninsula by the model. It makes the area just to the east of Malaysian Peninsula become more susceptible to convergence, hence, a stronger upward motion. This might be the reason for the unnecessary production of a rainfall peak by the control simulation over this region during the summer monsoon (cf. Figure 5.2d). This is, however, a small-scale effect and is not a dominant factor that can effectively increase the mean strength of upward motion for the eastern region overall. The mean upward vertical motion over the eastern region, therefore, is still underestimated by the model during summer monsoon season.

5.3: Impacts of Deforestation Over Southeast Asia: An Overview

Henderson-Sellers *et al*. (1993), McGuffie *et al*. (1995), Polcher and Laval (1994b) and Zhang *et al*. (1996a) (cf. Table 4.2, Chapter 4) reported that deforestation in Southeast Asia contributes only limited effects on local and regional climate. Impacts over Southeast Asia are shown to be quite different from those seen over the Amazon Basin: (i) a small decrease in surface temperatures; and (ii) possible effects on the monsoon system.

5.3.1: Annually averaged climatic changes

Table 5.1 summarises the annually averaged climatic changes, the differences between the control and deforested climate, as computed by Zhang *et al*. (1996a). The values given are area-average for the region (from 15^{o}S to 20^{o}N and from 95^{o}E to 150^{o}E). The surface air temperature decreases only slightly and total precipitation is reduced following deforestation.

Removal of the rain forest also leads to a reduction in evaporation. Overall, on an annual average basis, local hydrological processes are changed as shown by the reduction in moisture convergence. This means, the reduction in total evaporation is less than the reduction in total precipitation.

In the control experiment, the precipitation is mostly convective precipitation. About 42% of total precipitation in the region is contributed by evaporation, while the remainder (58%) is provided by the regional moisture convergence from the Pacific and Indian Ocean. Evaporation is, therefore, slightly less important than moisture transport into the region for local precipitation. This is because of the vast amount of rainfall received throughout the year, particularly in the wet season following strong forcing of rainfall. Most of the rainfall is caused by orographic uplift with continual "fuelling" from the adjacent tropical ocean. The areas of tropical forest, therefore, seem to playing a lesser role in producing the high rainfall of this area than in the South America.

Table 5.1: Annually averaged climatic changes over Southeast Asia Region following deforestation. [After Zhang *et al*.(1996a)]

Variable	Control	Differences	
Surface air temperature (K)	297.6	-0.2	
Total precipitation (P) (mm yr^{-1})	3169.0	-250.9	(-8%)
Total convective precipitation (mm yr^{-1})	2219.0	-235.6	(-11%)
Total large-scale precipitation (mm yr^{-1})	956.5	-15.4	(-2%)
Evaporation (E) (mm yr^{-1})	1345.0	-137.6	(-10%)
Moisture convergence (P-E) (mm yr^{-1})	1824.5	-113.3	(-6%)

Table 5.2: As in Table 5.1 but for the Amazon Basin. [After Zhang *et al*.(1996a)]

Variable	Control	Differences	
Surface air temperature (K)	298.7	+0.3	
Total precipitation (P) (mm yr^{-1})	1897.2	-402.0	(-21%)
Total convective precipitation (mm yr^{-1})	1164.0	-306.0	(-26%)
Total large-scale precipitation (mm yr^{-1})	733.2	-96.0	(-13%)
Evaporation (E) (mm yr^{-1})	1243.2	-222.0	(-18%)
Moisture convergence (P-E) (mm yr^{-1})	654.0	-181.2	(-28%)

Table 5.2 gives comparable figures for the Amazonia (average for the region from 15°S to 5°N and from 80°W to 50°W). As given by the control experiment, in contrast to the Southeast Asia region, total precipitation in the Amazon Basin is contributed more by evaporation (66%) than the regional moisture convergence from the adjacent Oceans (34%). According to this model, evaporation is, therefore, more important than moisture transport for local precipitation in Amazonia.

Overall, on an annual average basis, the local hydrological processes in the Southeast Asia show smaller changes compared to Amazonia. For Amazonia, in addition to the opposite change of the surface air temperature, which increases slightly, the reduction in precipitation, evaporation and moisture convergence are at higher percentage compared to the Southeast Asia. The greater reduction in evaporation for Amazonia clearly indicates that tropical forest over this region plays more roles in producing the rainfall of this area than in Southeast Asia region.

5.3.2: Seasonal variation of impacts of deforestation

Henderson-Sellers *et al.* (1993) described grid-element-scale impacts of "instantaneous" tropical deforestation for Southeast Asia (and other region) and showed only a statistically significant decrease of precipitation in the wet seasons months (June and July). Whereas, the result of McGuffie *et al.* (1995) showed that following deforestation, total precipitation is, marginally, increased in the wet season from April to August but decreased in other months. This illustrates that the experiments of Henderson-Sellers *et al.* and McGuffie *et al.* indicated a different impact of deforestation over Southeast Asia despite the fact that both used the same model (CCM1-Oz) and the same six-year integration time for the perturbation (deforestation) experiments. Both also used the same cloud parameterization and radiation updates and full seasonal and diurnal cycles (cf. Table 4.2, Chapter 4). The only differences between the two experiments are that McGuffie *et al*. improved the following: (i) increased the time of integration for the control simulation from a six-year used by Henderson-Sellers *et al*. to a fourteen-year; (ii) improved a formulation of the q-flux correction in the slab ocean model for ocean advection of energy (cf. Table 4.2 and 4.3 in Chapter 4); and (iii) added tropical Africa as another deforestation region (changed to degraded grassland vegetation) in addition to the Amazon Basin and Southeast Asia, as covered by Henderson-Sellers *et al*.

Figure 5.7 summarises seasonal variation of impacts of deforestation over Southeast Asia as computed by Zhang *et al*. (1996a). Also included are the results of Student's *t* tests represented by a *P* value (the possibility that changes are not statistically significant), which is arbitrarily cut off at 25% in the plots. Figure 5.7a shows that reductions of monthly total precipitation exist almost throughout the year but few are statistically significant. This is in agreement with Henderson-Sellers *et al*. (1993) and McGuffie *et al*. (1995). A statistically significant decrease in evapotranspiration following deforestation (Figure 5.7b) is shown throughout the year with larger reductions during the summer monsoon compared to the winter monsoon season. The change in moisture divergence [i.e., the difference between total evapotranspiration and total precipitation (E–P)] is shown in Figure 5.7c. There is strong moisture convergence throughout the year, which is weakened after deforestation, but only the change in October is statistically significant. For the surface net radiation, the differences between control and deforestation (Figure 5.7d) are statistically significant all year round but show little seasonality in the changes. For the sensible heat flux (Figure 5.7e), except for the two short periods April-May and September-October, the changes are statistically significant for all other months, with a striking seasonal variation (a decrease during winter monsoon but increase during summer monsoon). Lastly, for the surface temperature (Figure 5.7f), there are small decreases in most months, and the decreases during February to May (i.e., end of winter monsoon and inter-monsoon period) are larger than during the summer monsoon period. It seems that surface temperature is affected very little by deforestation during the summer monsoon months. A statistically significant change, however, is only present in April-May and November, suggesting that during inter-monsoon periods the land surface thermal conditions are most easily influenced by deforestation (Zhang *et al.*, 1996a). As a measure of vertically integrated diabatic heating and sensitive to a variety of climatic-forcing mechanisms, precipitation is probably the most appropriate variable to look at when studying climatic change at a local to regional scale. Total precipitation is, however, difficult to investigate in the context of deforestation in Southeast Asia region (Henderson-Sellers *et al.*, 1993). For example, on an area average basis, the total precipitation changes for January and July following deforestation are not statistically significant (cf. Figure 5.7a). As shown by

Figure 5.8, however, certain point values of the precipitation changes indicate an appreciable difference for precipitation received following deforestation in the two months. In both January and July, there are areas of significant changes. Such localised changes, averaged out in the area average analysis, are most likely associated with small-scale circulation changes induced by deforestation. These local scale changes, therefore, should not be ignored when interpreting climatic changes at the local to regional scale. As indicated in Figure 5.8a for January, a large reduction of precipitation occurs where the most active area of the ITCZ is located within the region during this winter monsoon season (cf. Figure 5.4). The two areas most affected by reduced precipitation are the areas centred just to the west of Sumatra and over northeast of Papua New Guinea (partly shown in Figure 5.4). The reduced point values, however, are almost compensated for by areas of increased precipitation, making the area average change in January not statistically significant.

Although an overall reduction of precipitation (not statistically significant) is given by the area average in July, point values indicate an effective local climatic change (see Figure 5.8b). For example, a large increase of precipitation is shown over two areas, centred at the north and southeast of Philippines. The increase of precipitation could be the result of an intensification of the monsoon trough or its shift from a normal position to these areas following deforestation. On the other hand, there is a major reduction of precipitation over the South China Sea centred near the East Coast of Malaysian Peninsula. This reduction, however, may occur because the model’s control run already indicates a positive bias as it overestimates precipitation over this area. The reduced point values are almost compensated by the increased point values and, therefore, make the area-average change in July not statistically significant.

5.4: Regional-scale Circulation Impacts of Tropical Deforestation

5.4.1: Walker and Hadley circulation changes

Monsoons have their origin outside the affected Southeast Asia region, and their circulation patterns are closely related to the overall general circulation of the atmosphere. The change in

land-surface characteristics because of deforestation over tropical forest regions may affect the atmospheric general circulation. One outcome anticipated from tropical deforestation is a reduction in the vertical ascent over the deforested region(s) caused by the imposed increase in surface albedo and the decrease in canopy extent and vegetation roughness length. By mass continuity, if there is a decrease in ascent above a deforested region, there will be changes beyond the areas of disturbance, either an increase in ascent or a decrease in descent elsewhere. This, in turn, might affect the components of either the longitudinal Walker or the latitudinal Hadley circulations in the area deforested.

Moreover, if large-scale land-surface disturbance caused by deforestation occurs coincidentally at the same time in more than one geographical location, then it might prompt circulation changes beyond the area of prescribed change. Then, it is possible that the effects of deforestation in more than one region will produce a sort of interaction, either dampening or reinforcing disturbances outside its own region. This possibility is assessed here by considering the changes in the vertical velocity over the three deforested regions. Latitudinal and longitudinal cross-sections are used to capture the cells of the Walker and Hadley circulation.

In the latitudinal cross-sections, the control experiment clearly shows the cells of the Walker circulation in all months. As noted earlier, to represent the two seasons only January and July cross-sections are shown here (Figure 5.9a and 5.10a). There is large-scale ascent over the western Pacific (~130^{o}E–180^{o}E), over Southeast Asia (~100^{o}E), over tropical Africa (~25^{o}E) and over the Amazon Basin (~50^{o}W). Apart from the western Pacific branch, the strong ascending branches of the Walker circulation are all located over tropical forest regions. The strength of ascent over all the above regions varies from month to month (not shown here). The most likely reason for this variation is the mean position and strength of ITCZ, which normally fluctuate from month to month. Ascent over Southeast Asia is significantly weaker in January than in July, suggesting that the ITCZ is stronger in July than in January over this region. On the other hand, ascent over the Amazon, tropical Africa and west Pacific is significantly weaker in July than in January, which is also likely due to the different strength of the ITCZ over these regions during these months.

The effects of deforestation are easier to assess by looking at the difference between the experiment and control vertical velocities. In January, the ascent over the three tropical forest regions is considerably diminished (Figure 5.9c). Over the equatorial western Pacific Ocean, adjacent to Southeast Asia region (between 130^{o}E-150^{o}E), however, the ascent appears to be strengthened to the west of about 140^{o}E but diminished to the east following deforestation. In July, the ascent over the Amazon Basin and tropical Africa is clearly weakened (Figure 5.10c). A different characteristic, however, is shown over the Southeast Asia region. A split change seems to occur in this region when to the west of about 100^{o}E the strength of the ascent is increased but to the east of this longitude the ascent is diminished. This might be due to a differential effect on maritime Southeast Asia, continental Southeast Asia and nearby Indian Ocean following deforestation. Similarly for the western Pacific, as also in January, a split change in the ascent can be detected as it is diminished to the east of 140^{o}E but strengthened to the west. These different characteristics of the effect on the ascent over Southeast Asia and western Pacific can also be interpreted as an eastward shift of ITCZ or monsoon trough in Southeast Asia region during both summer and winter monsoon following deforestation.

Figure 5.11 and 5.12 depict the Hadley circulation as captured by cross-sections through the atmosphere extending from 90^{o}E to 130^{o}E (i.e. including the Southeast Asia region). In the control simulation, the strongest ascent is clearly seen near the equator in January (Figure 5.11a) and at a broader range from 10^{o}N to 20^{o}N in July (Figure 5.12a). The strongest ascending limb near the equator in January is in consistent with the near equatorial trough associated with the northeast monsoon disturbances during this winter month. In July, however, the strongest ascending branch shifts northward with the establishment of the ITCZ associated with the summer monsoon. After deforestation, in January, the upwelling in the near equatorial trough and the descent in both subtropical regions immediately to the north and south are weakened (Figure 5.11c). In July, however, as indicated by the difference field (Figure 5.12c), the change of the upwelling in the ITCZ is somewhat mixed as some parts show weakening while others indicate strengthening. The zonal average taken from 90^{o}E to 130^{o}E to represent the change of Hadley circulation here could well be capturing both the changes in the Indian Ocean as well as the Southeast Asia sector. The differential effects

between the two areas, as suggested before for the Walker circulation, could probably cause the mixed character of the upwelling change in July.

In both January and July (Figures 5.11c and 5.12c), it seems that some effects extend outside the subtropical region (i.e. outside of 40^{o}N to 40^{o}S) following deforestation. This would suggest that deforestation might, particularly in July, influence the circulation at middle and high latitudes.

5.4.2: Southeast Asia-scale circulation changes

In this assessment, the differences of the surface vertical velocity, divergence and vorticity fields between the deforested experiment and the control are used to examine how Southeast Asia–scale monsoon circulations response to tropical deforestation. Again, both winter and summer monsoons are represented by conditions in January and July. Figures 5.13 and 5.14 illustrate the differences in vertical velocity near the surface for January and July, respectively. The control experiment (Figure 5.13a and 5.14a) indicates that the surface characteristics of the regional circulation of the atmosphere over this region are simulated satisfactorily; the potential areas of descent and ascent, as described earlier in the wind and pressure fields, are located correctly in both seasons.

Following deforestation, during the winter monsoon in January, an obvious change occurs over Southeast Asia. Again, the change is assessed through the difference field (Figure 5.13c). The area around the north of Thailand (centred at 100^{o}E and 20^{o}N) experiences major enhancement of downward motion, whilst the area to the south (centred at 100^{o}E and 10^{o}N) develops a suppression of the downward motion. The control experiment (Figure 5.13a) shows subsidence over both of the areas mentioned, which is in agreement with observation, when dry conditions normally prevail over the continental Southeast Asia during winter months. Figure 5.13c also indicates a slight diminishing of upward vertical velocity over the area of the monsoon or near equatorial trough, suggesting that this trough is weakening following deforestation. The character of changes indicated here could modify the circulation pattern over Southeast Asia during the northeast monsoon season.

In July, the area around the north of Thailand experiences an opposite effect following deforestation (Figure 5.14c), an enhancement of the upward vertical velocity. This, however, is balanced by the suppression of upward motion (associated with the monsoon trough) over the north of India. It is important to note that any change in the signal of the monsoon trough over the Indian subcontinent is connected to a circulation disturbance linked with conditions over the Indian Ocean. The circulation over this oceanic region is closely related to the penetration of the southern hemisphere Hadley cell during this month. These modifications may also be interpreted as an eastward shift of the mean position of the summer monsoon trough, in agreement with the change detected in the Walker circulation. In July, therefore, deforestation also prompts changes to the circulation pattern of the summer monsoon over Southeast Asia and the nearby Indian Ocean.

Similar to the vertical velocity, the control experiment also clearly indicates that the major area of convergence/positive vorticity and divergence/negative vorticity is located correctly in both seasons (Figure 5.15a and 5.16a). After deforestation, the changes of divergence and vorticity in both January and July are well established, in correspondence to the vertical velocity changes. In January, the major change indicated earlier by the vertical velocity is confirmed by the divergence and vorticity fields (Figure 5.15c). Divergence over the area to the north of Thailand is increased, while negative vorticity over the same area increases. On the other hand, divergence and negative vorticity over the southern area (centred over Cambodia/Gulf of Thailand) changes sign completely as the area is under the influence of convergence and positive vorticity following deforestation (comparing Figures 5.15a and 5.15b). The changes can be seen very clearly in Figure 5.15c. In addition, after deforestation we can also see a decrease in magnitude of convergence and a reduction in positive vorticity within the area of the monsoon or near equatorial trough. The monsoon trough seems to be weakened following deforestation. This change, therefore, is in agreement with the weakening of the ascending branch of Hadley circulation over this region in January.

In terms of divergence and vorticity, the change in July following deforestation (Figure 5.16c) is also in general agreement with the vertical velocity change. In certain areas, however, the divergence and vorticity fields capture the changes clearer than the vertical

velocity field. One remarkable difference, not clearly seen from the vertical velocity field but that can be detected from these two fields, is that deforestation causes a different and an opposite change between continental and maritime Southeast Asia. The continental area, in general, is influenced by the weakening of convergence and positive vorticity, whereas the maritime area is influenced oppositely. There is also an indication of an eastward shift of the strongest zone of the summer monsoon trough following deforestation, as can be seen from the difference field.

5.4.3: Main findings

Overall, the impact on the general circulation of the atmosphere following deforestation can be detected in the cells of the Walker circulation in both January and July. The results show diminished ascent in either the ITCZ or the monsoon trough over the three deforested areas. The circulation strength of the Hadley cell is also decreased over Southeast Asia despite different responses shown between January (northeast/winter monsoon) and July (southwest/summer monsoon).

It is also interesting to note that the area to the north and immediately to the south of Thailand undergoes similar changes in divergence but of the opposite sign in both seasons. Deforestation in this region, therefore, appears to be altering the character of these monsoons by affecting the extent to which they penetrate different parts of the region. Another example is the split change depicted between the west and east of about 100^{o}E in the summer monsoon, when the strength of the ascent changes in opposite direction following deforestation.

Analyses of the climatic changes evident in earlier simulations [e.g. Henderson-Sellers *et al.* (1993) and McGuffie *et al* (1995)] had already suggested that the impact of deforestation often extends beyond the deforested regions themselves. Based on the same experimental results used in this analysis, Zhang *et al.* (1996*a,b*) have supported this suggestion. Tropical deforestation is shown to affect an area adjacent and to the southwest of the Amazon Basin, the monsoon circulation in Southeast Asia and the ocean off the west coast of tropical Africa.

These effects have been suggested to be result of the regional atmospheric circulation changes over those regions. With special reference to Southeast Asia and the nearby areas, this analysis gives a similar impact scenario.

It is not, however, possible to confirm how strong the contribution of deforestation in Southeast Asia is to the changing signals over the Indian Ocean or/and west Pacific Ocean. It is also not clear whether the change over Southeast Asia is largely contributed by its own deforestation or deforestation in other region(s). As indicated in previous analyses, deforestation in more than one region is likely to produce interactions, either dampening or reinforcing disturbances outside its own region through global scale interaction. It is possible that the changes over an area distant from a deforested region are contributed to by a composite effect from more than one deforested region.

In the context of the Southeast Asia region, however, this GCM experiment based on the “three-region deforestation” has given us a valuable direction in building hypotheses that can be tested in this study. In general, it appears that deforestation can alter the character of monsoons, by affecting the extent to which the monsoon penetrates different parts of the region and producing a differential effect between continental and maritime areas, between north and south, and/or west and east part of the region. This analysis, therefore, has successfully defined hypotheses to be tested in this study's model experiment as follows: -

(i) that deforestation can, in general, modify the circulation patterns over this region during the summer and winter monsoon seasons, by shifting the monsoon trough of both seasons to the east; and

(ii) that deforestation can, more specifically, result in significant changes to the signals over the Indian Ocean and west Pacific Ocean by: (a) weakening both the summer monsoon trough over the Indian subcontinent and the winter monsoon trough over Southeast Asia region; and (b) strengthening both the effect of the summer and the winter monsoon over the western part of west Pacific Ocean, adjacent to Southeast Asia region.

Over certain regions, surface climatic changes may be linked to the circulation changes as indicated in the hypotheses. For example, an increase in upward motion associated with a trough indicates an enhancement of convective action, which would increase surface evaporation and produce an increase in precipitation. In addition, a shift of the monsoon trough or large-scale ascent from one region to another would effectively change moisture convergence fields, by enhancing or dampening in a certain region.

5.5: Summary

The NCAR Community Climate Model, incorporating the Biosphere-Atmosphere Transfer Scheme (BATS1e) is evaluated by comparisons between the control climate simulation and the mean observed climatology defined either by the NCEP/NCAR Reanalysis Data or the observed dataset from Legates and Wilmott. There is no apparent bias in the GCM as far as the summer monsoon over northwestern region of Southeast Asia is concerned. The model, however, underestimates the magnitude of upward motion for the eastern region during all the seasons and for the western region during the winter and inter-monsoon periods. There is an underestimation of vertical velocity given by the model in the eastern region during the winter monsoon resulting from an increase in divergence. The GCM used in this analysis, therefore, has various discrepancies in the control run. It gives positive as well as negative biases of the important surface variables such as the upward vertical velocity field and hence, precipitation over the Southeast Asia region. This is most likely due to unresolved subgrid-scale processes, related to, for example, topography.

Following deforestation, the mean changes in climate for the whole Southeast Asia region are not statistically significant. At a localised scale, however, the changes can be very large in magnitude. These localised changes, most likely, are the result of the regional-scale circulation changes. The analysis of the Southeast Asia-scale circulation changes shows that the impact on the general circulation of the atmosphere following deforestation can be detected in the cells of the Walker circulation in both January and July. The results show diminished ascent in either the ITCZ or the monsoon trough over the three deforested areas. The circulation strength of the Hadley cell is also decreased over Southeast Asia despite different responses shown between January (northeast/winter monsoon) and July (southwest/summer monsoon). In general, deforestation alters the character of the Southeast

Asia monsoons, by affecting the extent to which the monsoon penetrate different parts of the region and producing a differential effect between continental and maritime area, between north and south, and/or west and east part of the region.

The analysis successfully defines hypotheses for further examination. In general, the formulated hypotheses relate to changes and shifts in the strength of both winter and summer monsoon troughs. Over a certain region, surface climatic changes may be linked to the circulation changes as indicated in the hypotheses. For example, an increase in upward motion associated with a trough indicates an enhancement of convective action, which would increase surface evaporation and produce an increase in precipitation. In addition, a shift of monsoon trough or large-scale ascent from one region to another would effectively change moisture convergence fields, by enhancing or dampening at a certain region.

Lessons learned from this re-analysis of the Zhang *et al*. (1996a, b) experimental results have guided the design of the current model study (Chapter 8), including selection of diagnostic variables and analysis methodology (Chapter 9 and 10).