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The Climatic Effects of Deforestation in South and Southeast Asia

Rachindra Mawalagedara and Robert J. Oglesby
University of Nebraska, Lincoln
USA

1. Introduction

Deforestation is the removal of the existing natural vegetation cover, especially where the native cover is largely forest. The growth in the world population has increased the clearing of forests to obtain fuel and building material, to grow crops and to raise livestock. Over the past 300 years, 7-11 million km^2 of forest has been cleared (Foley et al. 2005). Deforestation can have a devastating impact on biodiversity as about 70% of land dwelling animals and plants are found in forests. Impacts such as land degradation in the absence of forest regrowth, soil erosion and sedimentation in rivers can have a negative impact on the environment. These impacts are discussed in greater detail in the other chapters of the book. Importantly, deforestation can also have strong effects on climate.

In the past it was assumed that the local climate determined the vegetation type in a region (Nobre et al. 1991) with the amount of incoming solar radiation, precipitation and soil type determining the vegetation cover of the region. But studies have shown that the atmosphere and the vegetation interact with each other, exchanging energy, moisture and momentum (Zeng et al. 1999) and are in a dynamic equilibrium (Nobre et al., 1991). Therefore any change in vegetation cover can potentially lead to a change in the climate. As deforestation is a pressing problem in most parts of the world it is important to understand the possible consequences of deforestation and the mechanisms by which the change in land cover can alter the climate.

The impacts of land cover changes on the atmosphere have been studied extensively using both observations and computer models (Suh and Lee, 2004; Lean and Warrilow, 1989; Kanae et al., 2001; Clark et al., 2001). Previous studies have shown that deforestation can change the surface albedo, surface roughness and the amount of evapotranspiration (evapotranspiration is the combined effect of evaporation from the surface and the transpiration from vegetation) (Gibbard et al., 2005; Oglesby et al., 2010; Hasler et al., 2007) thus, leading to a modification of the surface energy and moisture budgets.

In order to determine the full climatic impact of deforestation, it is necessary to understand the behavior of the surface energy and the moisture budgets, as deforestation interacts directly or indirectly with all the components of these budgets. The surface energy budget looks at all the possible sources and sinks of energy at the surface as well as any possible horizontal transport (fluxes) and storage of energy within the seasonally active layer just below the surface. Over land, incoming and reflected solar radiation (shortwave radiation),

incoming and outgoing longwave radiation, sensible heat flux, latent heat flux and ground storage are the important terms that need to be considered for the surface energy budget. Unlike in the oceans where transport of energy (by the oceans) is significant to the climate system, horizontal transport of energy in the ground is negligible. The moisture budget is related to the hydrologic cycle and takes into account precipitation, evaporation, surface runoff (horizontal transport of water) and storage. If the period considered is a year or longer, the storage of energy and moisture can be considered negligible. The changes in the terms of the two budgets can be used to determine the changes in the climate.

To understand how deforestation impacts the climate it is important to understand the behavior of the terms that make up the surface energy and moisture budgets. The primary source of energy that contributes to the surface energy budget is the sun. The energy that sustains the Earth and drives the global circulation is acquired from incoming solar radiation but is not absorbed directly by the atmosphere. Instead, most of the incoming solar radiation is first absorbed by the surface. The amount of solar radiation absorbed by the surface depends on the surface albedo, which determines the fraction of solar radiation reflected from the surface. The albedo can be expressed either as a percentage or a fraction, ranging from 100% (1.0) to 0% (0.0) with the former value indicating that all the incoming radiation is reflected (no absorption) and the latter indicating that no reflection of radiation takes place (all incoming radiation is absorbed). The surface albedo is related to the texture and the colour of the surface, with dark rough surfaces (low albedo) absorbing more energy.

Once the energy is absorbed by the surface, radiative and non-radiative processes transfer it from the surface to the atmosphere. Part of the absorbed energy warms the surface and is then emitted as longwave radiation from the surface. The magnitude of the emission is determined by the temperature and the emissivity of the surface (emissivity depends on the type of the surface: vegetated, bare soil etc.) and can be determined by the Stephen-Boltzmann Law. Part of the longwave radiation emitted by the surface is absorbed and reemitted by the atmosphere (can be calculated using the emissivity and the temperature of the atmosphere). A portion of the reemitted longwave radiation then acts as a source of energy for the surface. The remaining energy is partitioned between sensible and latent heat fluxes which are the non-radiative terms in the surface energy budget. The sensible heat flux heats the atmosphere in contact with the surface and is a less efficient method of heat transfer compared to the latent heat flux. It is difficult to measure the sensible heat flux directly. But it can be calculated easily if the latent heat flux and the Bowen ratio are known. The Bowen ratio is a measure of the water availability in a region and is the ratio between sensible heat flux and the latent heat flux. The latent heat flux can be calculated using the rate of evaporation.

The latent heat exchange, a significant process in the surface energy budget, is proportional to the amount of evaporation, and thereby provides a link to the surface moisture budget. The magnitude of the latent energy flux depends both on the amount of moisture available at the surface and the energy available for evaporation. In the tropics energy is not usually a limiting factor and hence the dependency is on the water availability. Therefore the energy that is not emitted as longwave radiation or stored in the ground is partitioned between sensible and latent heat and this partitioning is determined by the amount of water available at the surface. The latent heat flux also provides a measure of cooling at the surface due to

evapotranspiration (i.e. water absorbs energy from the surface and evaporates thus cooling the surface). Once the evaporated moisture condenses in the atmosphere the latent energy is released. The energy released contributes to convection and helps to drive the local circulation. Therefore the latent heat exchange is not only an important cooling mechanism but is an important measure of the energy available for regional circulation.

The surface moisture budget accounts for precipitation, evapotranspiration and surface runoff. The evapotranspiration term links this to the surface energy budget. Changes in availability and the partitioning of energy can have an impact on evapotranspiration and hence other terms in the moisture budget. The information from the surface moisture budget namely, precipitation and evaporation can be used to compute the atmospheric moisture convergence/divergence which gives the net amount of water vapor transported into or out of a region by the regional circulation. If the amount of precipitation is larger than the local evapotranspiration this indicates that the moisture transport into the region makes a significant contribution to precipitation. Thus the atmospheric moisture convergence can be used to determine the relative importance of an external moisture source to that of local evaporation.

As discussed above both surface albedo and emissivity are sensitive to the nature of the land surface and hence depend on the type of land-use (forest, water, urban etc). The magnitude of the latent heat flux also depends on the land-use category as the water available for evapotranspiration changes with land-use. For example a vegetated surface would have more moisture available for evapotranspiration than barren land, and, as described more fully later in the chapter, forested land will have higher values than grassland or shrubland. Therefore it is evident that any change in the land-use in a region will modify both the surface energy and the moisture budgets.

Deforestation alters the land surface properties and the interactions between the surface and the atmosphere. Two of the most important changes due to deforestation are the increase in surface albedo and the decrease in evapotranspiration. The significance of these changes is discussed in more detail under the methods section.

Deforestation results in two competing effects, warming due to the reduction in evapotranspiration and a cooling due to the increase surface albedo. Previous studies have shown that in most regions the magnitude of warming is much greater than that of cooling, resulting in warmer and drier conditions (Zhang et. al., 1996; Oglesby et al., 2010). But these impacts are further modulated or enhanced by the dominant circulation patterns and moisture sources of the considered region. For example the change in precipitation due to the decrease in evapotranspiration would be more dramatic in a region such as the Amazon basin where 50% of the moisture available for precipitation comes local evapotranspiration (Lean and Warrilow, 1989). But if the region is close to a water body such as an ocean or a lake and the local and/or regional circulations are favorable for moisture transport, the contribution from local evapotranspiration would not be as significant. Most coastal regions receive abundant moisture from the ocean, carried inland by onshore flow (i.e. sea breeze – local circulation) contributing to precipitation. This mechanism alone is not strong enough for moisture to be transported to inland regions far from the oceans. But a large scale circulation pattern such as the Asian monsoon can penetrate further inland supplying moisture to continental regions thus making the impact of reduced evapotranspiration on precipitation much smaller.

This study focused on South Asia, Southeast Asia and Sri Lanka, all regions where the Asian monsoon plays an important role, and all regions where deforestation is currently or potentially a major issue. In these regions, the monsoon flow brings moisture from the ocean over land and this together with moisture provided by evapotranspiration produce abundant rainfall. The wet season is mostly during the summer monsoon period, but some areas experience rainfall during the winter monsoon as well. Any changes in this established pattern of rainfall and associated climate could have devastating consequences as this is a heavily populated area where agriculture is of great importance.

The Hadley cell is one of the prominent features that make up the global circulation. The Hadley cell consists of two asymmetric cells extending between approximately 15° in the summer hemisphere and 30° in the winter hemisphere (Lu et al. 2007). This circulation is defined by a branch of rising air over the surface low pressure belt (Inter-Tropical Convergence Zone- ITCZ) in the tropics, a poleward flow in both hemispheres in the upper troposphere, a branch of subsiding air in the subtropics and equator-ward flow (in both hemispheres) at the surface that converges at the ITCZ (Mitas et al., 2005). This circulation transports both energy and angular momentum and gives rise to many of the climatic regions that are present today. For example regions that fall under the ITCZ receive abundant rainfall due to the strong convection associated with the rising branch of the Hadley cell and either receive rainfall throughout the year or have well defined wet/dry seasons. The subtropical deserts are in the regions where the subsiding branches of the Hadley cell are found as the sinking of air prevents the formation of clouds and hence precipitation. Any changes in the strength or the extension of the Hadley cell will therefore impact the climate within the region and have the ability to change the geographical boundaries of these climate zones. As the location of Southeast Asia is closely associated with the rising branch of the Hadley cell and the poleward transport of energy, deforestation can have consequences on both regional and global scales.

It is evident that tropical deforestation can have an impact on the climate by modifying the magnitudes and the spatial and the temporal patterns of temperature and precipitation, creating warmer and drier climatic conditions in most regions. These changes might then be modulated or enhanced by the dominant circulation patterns such as the Asian monsoon leading to additional changes in the climate. The local moisture sources of the considered region may also play a significant role in modifying the climatic effects of deforestation Therefore, it is important to understand the possible consequences of deforestation and the mechanisms by which the change in land cover can alter the monsoonal climate.

2. Methods

In order to understand the impacts of tropical deforestation on the climate in South and Southeast Asia, a widely-used regional climate model (WRF – Weather Research and Forecasting Model) was employed. The Weather Research and Forecasting (WRF) Model is a next generation mesoscale numerical weather prediction system that has been developed as a collaborative effort by the National Center for Atmospheric Research (NCAR), the National Oceanic and Atmospheric Administration (the National Centers for Environmental Prediction (NCEP) and the Forecast Systems Laboratory (FSL), the Air Force Weather Agency (AFWA), the Naval Research Laboratory, the University of Oklahoma, and the

Federal Aviation Administration (FAA). WRF can be used in a wide scope of spatial scales ranging from a few meters to thousands of kilometers and is suitable for both operational forecasting and atmospheric research (Skamarock et al., 2008; http://www.wrf-model.org/index.php).

The main focus of this study was to identify the impacts of deforestation on the monsoonal climate in South and Southeast Asia. Therefore a Regional Climate Model (RCM) was used for this study instead of a Global Circulation Model (GCM), so that it was possible to run the model at a high resolution. This allowed the model to include the regional features and predict the regional climate with more accuracy. The GCM which usually has a horizontal resolution of 100 – 250 km (McGuffie and Henderson-Sellers, 2005). can capture the features of large and synoptic scale atmospheric circulation, but is too coarse to include small scale features such as the effects of topography or land surface effects to simulate the climate on a regional scale accurately (Denis et al., 2002). Since GCM simulations are done for the entire globe it is not feasible to run it at very high resolution due to limitations in computing power. Therefore regional climate models (RCM) are used to study regional climate changes. The main difference is a RCM is focused on the region of interest and has a much higher resolution. For example WRF simulations can be done at a resolution of 4 km which allows many small scale features such as mountains, coastlines and, importantly for our purposes, land-use categories to be represented more accurately.

The model was forced at the lateral boundaries by NCEP/NCAR Reanalysis data (NNRP)[1] . Since the model is a regional model the simulations are done for a restricted region. Therefore, conditions at the boundaries of the specified domain need to be provided so that the model can properly simulate the climate within the domain. Reanalysis data is used to provide these boundary conditions as observations alone are not sufficient to describe the full state of the atmosphere due to missing or spatially non-uniform data. Reanalysis data solves this problem by combining actual observations with a global model of the atmosphere to produce a comprehensive data set that serves as a proxy for real observations, thus providing better boundary conditions for the regional model.

The model simulations focused on three specific domains: 1. South Asia (Southern India and Sri Lanka with a resolution of 12 km), 2. Southeast Asia (with a resolution of 12 km), 3. High resolution focus on Sri Lanka (with a resolution of 4 km). The WRF simulations were done for the years 1988, 1991 and 1993. These years represent a strong, weak and normal monsoon year with respect to South Asia. For each of these three years, a control run as well as two idealized runs (completely deforested and forested situations) were carried out and analyzed. The control runs are also compared to actual observations in order to identify model strengths and weaknesses, as well as any biases. In the deforested run all the land use categories (other than inland water) were replaced with grassland, which has a higher albedo than the tropical forests, but much less capability at extracting water from the soil. In the forested run evergreen broadleaf forest was used. These land-use changes, while extreme, provided the maximum possible range of impacts due to deforestation.

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[1]The data for this study are from the Research Data Archive (RDA) which is maintained by the Computational and Information Systems Laboratory (CISL) at the National Center for Atmospheric Research (NCAR). NCAR is sponsored by the National Science Foundation (NSF). The original data are available from the RDA (http://dss.ucar.edu) in dataset number ds090.2.

In a WRF simulation each grid point has a land-use category (grassland, cropland, evergreen broadleaf forest, water etc.) assigned to it based on the land-use data set being used for the model run. The properties (surface albedo, surface emissivity, moisture availability, surface roughness length) of each land-use category depend on the land surface model used in the WRF run. The land surface model is the component that takes care of the processes involving land-surface interactions. For the WRF runs, the 5-layer thermal diffusion scheme was selected as the land surface model. USGS (winter) data set was used to specify land-use categories and their properties. To simulate deforested conditions, all the land-use categories other than water were replaced with grassland. Grassland has a higher albedo (23%) than most other land-use types and therefore absorbs less energy. The specified moisture capacity (0.30) is, however, also low. Water bodies have the lowest surface albedo (8%) and a specified moisture availability of 1.0 (saturated surface). Forested conditions were simulated by replacing all land-use categories other than water by evergreen broadleaf forest which has a very low surface albedo (12%), but much higher moisture availability (0.5) compared to grassland, with the former allowing the surface to absorb more incoming energy and the latter supporting larger evaporation amounts.

Before analyzing the results of the WRF simulations it is important to understand the possible impacts of the land-use changes made to the deforested and deforested runs (as mentioned above), on the surface energy and the moisture budgets and how the changes will ultimately affect the regional climate.

Tropical forests have a low surface albedo throughout the year. This allows the forests to absorb a large part of the incoming radiation. Most land-use categories have surface albedos that are higher than that of a tropical rainforest. Therefore tropical deforestation leads to an increase in the surface albedo, allowing the surface to reflect more radiation. As a result the surface absorbs less radiation creating a cooling effect. This also reduces the amount of energy available for evapotranspiration.

On the other hand, the trees found in tropical forests have the capacity to draw water from the soil and thereby add a large amount of moisture to the atmosphere via transpiration. The large leaf and stem area allows the trees to intercept a significant amount of the rainfall. The intercepted water is then evaporated into the atmosphere. Evapotranspiration can be an important moisture source for local precipitation, especially in regions that are not in the proximity of a water body. Also evapotranspiration helps to lower the surface temperature. Deforestation leads to a decrease in evapotranspiration. This removes or reduces the capacity of the local moisture source and the cooling effect of evapotranspiration. This also alters the energy partitioning between latent and sensible heat fluxes at the surface. Due to the reduction in evapotranspiration and hence the latent energy flux, the energy transfer between the surface and the atmosphere would be achieved mostly through the exchange of sensible heat. As this is a less efficient method of heat transfer (compared to the cooling by latent heat transfer), this would lead to an increase in the surface temperature. The reduction in the latent energy flux means the energy available for convection is reduced. This can potentially lead to a weakening of the local circulation that in turn can have a negative impact on the moisture convergence in the area. The reduction of moisture convergence and evapotranspiration result in a reduction in precipitation. Therefore the reduction in evapotranspiration can result in a warming of the surface and a decrease in precipitation. The decrease in precipitation then acts as a positive feedback to further reduce

the evapotranspiration and enhance the warming effect. Thus, deforestation will result in a warmer and drier climate.

The surface energy and the moisture budgets were analyzed to see if the possible changes that are discussed above were present in the WRF output. Temperature, precipitation and evaporation anomalies (deforested – forested) were also calculated to identify the climatic impact of deforestation.

3. Results

In response to deforestation, for all three years, there is an increase in temperature over majority of the land areas whereas the changes over the oceans are more variable. Figure I show the changes in temperature (at 2 m) between the deforested and the forested runs for the year 1988. (For brevity, we show changes for one representative year.) Changes in the magnitudes in the spatial patterns of both the annual and the seasonal (JJA) temperatures are shown in the figure. The period of June, July and August (JJA) was selected to focus on the height of the summer monsoon season. Most regions show a clear warming, evident in both the annual and the seasonal values. The most prominent warming is seen along the west coast of India and Sri Lanka. There is a small region in the central highlands of Sri Lanka where the response is a cooling of temperature. Figure II shows the time series of temperature over land areas for 1988. The warming is evident in the monthly temperatures in all of the domains. The temperature at 2 m (for 1988) for the deforested and the forested situations and the corresponding changes are shown in table II.

The spatial pattern of precipitation over each of the three years shows a decrease over land whereas some regions over the oceans experience enhanced rainfall. The annual precipitation values show a larger reduction than just the values in the monsoon season. This indicates that while the amount of precipitation received during the summer monsoon is affected by deforestation so are the other mechanisms such as the winter monsoon and convection that provide rainfall during the rest of the year. Figure III shows the precipitation anomalies between the deforested and the forested simulations for the representative year 1988. Also the increase in Bowen ratio (over land) indicates that conditions become drier. The annual precipitation (for 1988) over South Asia decreases by 19% while the reduction over Sri Lanka is only 10% relative to the forested run. Southeast Asia experiences a 53% decrease in precipitation as a result of deforestation (All the percent changes provided in this chapter are calculated as the difference between the deforested and forested runs with respect to the forested run. i.e. [(D-F)/F]*100%).

Evaporation over land is much smaller after deforestation but the changes over the ocean in the domains over South Asia and Sri Lanka show an increase. (The increase over the ocean is likely due to the warmer temperatures due to deforestation. The amount of water vapor the atmosphere can hold strongly depends on the temperature with warm air being able to hold more moisture than cold air. As deforestation warms the atmosphere, the air over the adjacent ocean also warms, gaining the ability to hold more moisture. This implies that more evaporation can take place before the air is saturated thus enhancing evaporation over the ocean.) Overall, the annual evapotranspiration over all three domains is reduced. This can be seen in figure IV which shows the anomalies in evapotranspiration for 1988. The largest decrease is seen over Sri Lanka where the (area averaged) annual evaporation decreases by

465 mm (table I), which is 29% reduction compared to the forested run, over the year. The annual evapotranspiration over South Asia decreases by 420 mm (27%) whereas the decrease over Southeast Asia is 415 mm (25%). The reduction in evapotranspiration is a result of the lower transpiration, reduced interception of precipitation (due to reduced leaf and stem area) and smaller roughness length. The roughness length provides a measure of the surface friction and the exchange of moisture between the atmosphere and the surface, with smaller roughness lengths modulating the exchange.

The latent heat flux shows a decrease over all domains, consistent with the reduction in evaporation. As moisture becomes limited with the change in land-use, less energy is used for evapotranspiration. This means the evaporative cooling at the surface is reduced and the remaining energy goes into warming the surface. The sensible heat flux, which heats the air in contact with the surface, shows an increase due to the change in energy partitioning at the surface (i.e. in response to the decrease in latent energy flux). The changes in evaporation, latent heat flux and sensible heat flux support the warming observed in response to deforestation.

Precipitation is greater than the evapotranspiration over South Asia and Sri Lanka. This indicates that the moisture provided from a moisture source (i.e. moisture transported from the Indian Ocean into the area by the regional circulation) other than local evapotranspiration makes a significant contribution to the precipitation in the region, suggesting that the Indian monsoon plays a dominant role where precipitation is considered. Out of the two regions the moisture transport is more significant for Sri Lanka as precipitation over the region can be more than twice the amount of the local evapotranspiration.

The decrease in precipitation is much larger than that of evapotranspiration over Sri Lanka. The decrease in the local moisture recycling capacity alone cannot explain this strong reduction in precipitation. This indicates the moisture transported into the region has decreased, signaling a weakening or a change in the regional circulation leading to a reduced atmospheric moisture convergence into the area. This shows that while the moisture brought in with the monsoon flow may be a prominent factor in determining the amount of precipitation, deforestation can apparently weaken the moisture flux so that less water is available for precipitation in the region.

On the other hand, the reduction in precipitation over South Asia is less than that of evaporation. This indicates a stronger moisture convergence over the Indian subcontinent. As a result deforestation causes two competing effects on precipitation. The decrease in evapotranspiration has a negative impact on precipitation whereas the increased moisture convergence has a positive effect on it. But the magnitude of the latter is smaller than that of the evapotranspiration, therefore ultimately resulting in a decrease in precipitation.

The difference between precipitation and evapotranspiration is also a measure of surface runoff and storage. The changes in surface runoff can be very important for streamflow in the region. But further analysis of these terms are not possible because the land surface model used in the WRF runs does not have the capability to compute these terms separately.

Clouds reflect part of the incoming solar radiation back out to space, hence the amount of incoming shortwave radiation at the surface can be used as a proxy for cloud cover. The incoming shortwave radiation shows a reduction for all three years over all domains. This is a indication of reduced cloud cover. The decrease in evaporation (and hence latent heat flux) means that there is less moisture.
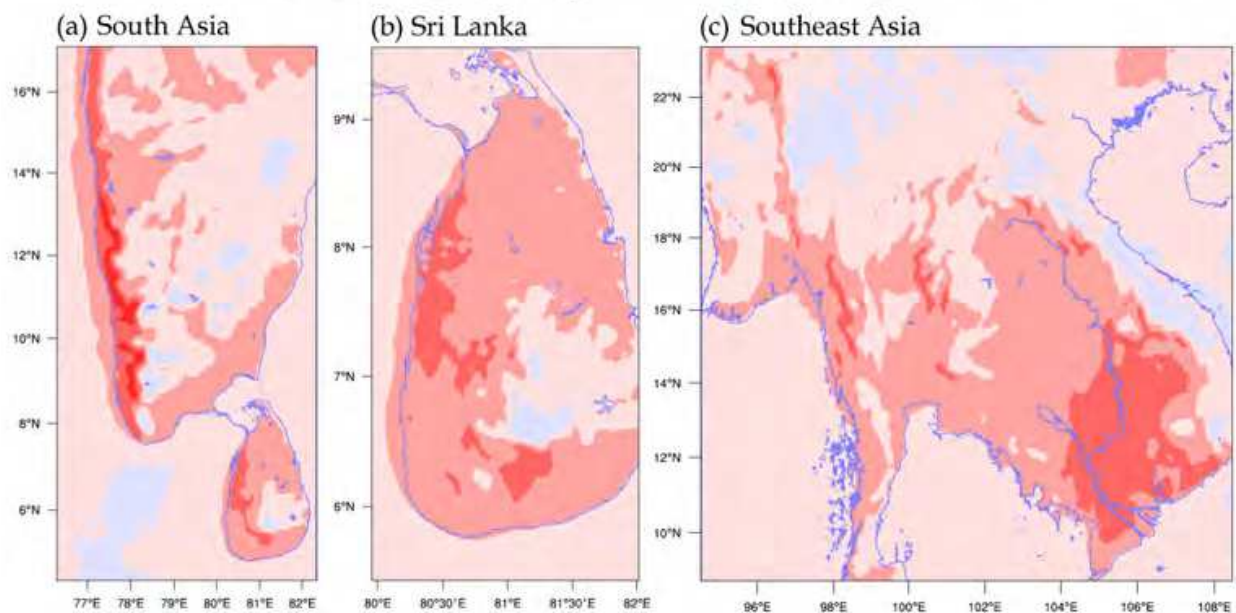
Term	South Asia			Sri Lanka			Southeast Asia		
	Deforest D	Forest F	Change D-F	Deforest D	Forest F	Change D-F	Deforest D	Forest F	Change D-F
SW_{down}	249.26	242.46	6.80	225.36	219.44	5.92	277.88	270.59	7.29
SW_{up}	57.33	29.10	28.23	51.83	26.33	25.50	63.91	32.47	31.44
SW_{net}	191.93	213.36	-21.43	173.53	193.11	-19.58	213.97	238.12	-24.15
LW_{down}	387.23	390.36	-3.13	414.85	415.26	-0.41	366.21	371.19	-4.98
LW_{up}	405.92	416.00	-10.08	418.60	428.21	-9.61	402.41	411.88	-9.47
LW_{net}	-18.69	-25.64	6.95	-3.75	-12.95	9.20	-36.20	-40.69	4.49
R_{net}	173.24	187.72	-14.48	169.78	180.16	-10.38	177.77	197.43	-19.66
LE	89.72	122.96	-33.24	88.77	125.58	-36.81	96.02	133.84	-37.82
SH	35.14	27.11	8.03	39.41	27.63	11.78	44.30	38.53	5.77
G_E	48.38	37.65	10.73	41.60	26.95	14.65	37.45	25.06	12.39
P (mm/yr)	1444.39	1775.57	-331.18	4700.86	5206.67	-505.81	172.93	370.34	-197.41
E (mm/yr)	1134.89	1555.33	-420.44	1122.83	1588.47	-465.64	1214.58	1629.90	-415.32
P-E (mm/yr)	309.50	220.24	89.26	3578.03	3618.20	-40.17	-1041.65	-1259.56	217.91
Bowen Ratio	0.39	0.22	0.17	0.44	0.22	0.22	0.46	0.29	0.17

Table 1. The terms of the surface energy and the moisture budgets for deforested (D) and forested (F) WRF simulations as well as the difference between the two simulations (D-F) over South Asia, Sri Lanka, and Southeast Asia . These are the areas averaged values for 1988 and are computed only over land areas. Precipitation and evaporation are yearly totals. The units are Wm^{-2} unless noted otherwise in the table. The Bowen ratio is unitless. Here SW – shortwave radiation, LW – longwave radiation, R_{net} – net radiation, LE – latent energy flux, SH – sensible heat flux, G_E – storage, P – precipitation, E – evaporation. The subscripts (up/down) give the direction of the radiation. SW_{net} and LW_{net} are calculated as the difference between downward and upward radiation. A positive net radiation value indicates that the radiation is aimed towards the surface (downwards).

Term	South Asia			Sri Lanka			Southeast Asia		
	Deforest D	Forest F	Change D-F	Deforest D	Forest F	Change D-F	Deforest D	Forest F	Change D-F
Temperature (at 2 m) (K)	296.56	296.31	0.25	298.82	298.47	0.35	295.72	295.47	0.25
Surface Skin Temperature (K)	296.88	296.35	0.53	299.22	298.54	0.68	296.13	295.53	0.6
Surface Emissivity	0.92	0.95	- 0.03	0.92	0.95	- 0.03	0.92	0.95	- 0.03
Surface Albedo	0.23	0.12	0.11	0.23	0.12	0.11	0.23	0.12	0.11

Table 2. The area averaged annual temperature at 2 m and surface skin temperature (surface temperature) are shown here. Temperature values are in Kelvins (K). Surface emissivity and surface albedo for both deforested and forested simulation are included here. All the values in the table are for 1988

Annual Temperature Anomalies

Seasonal Temperature Anomalies
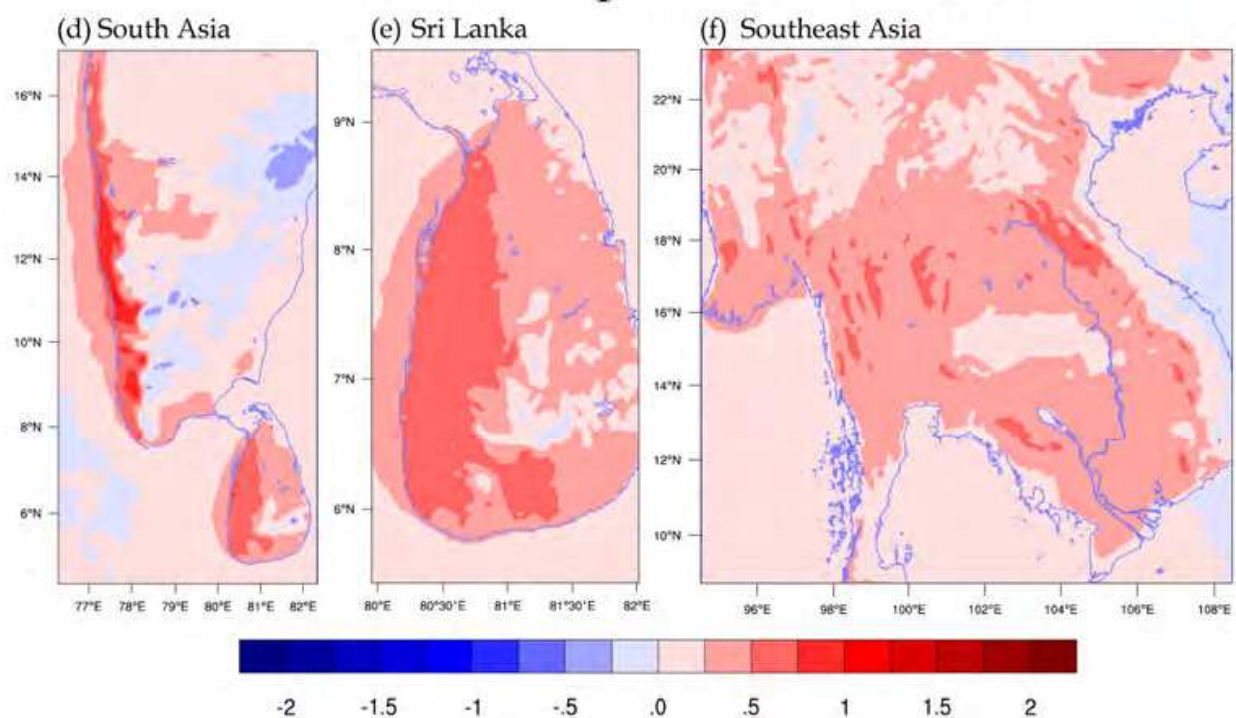
Fig. 1. The annual and the seasonal (June, July and August) temperature anomalies (i.e. deforest –forest) for South Asia (a,d), Sri Lanka (b,e) and Southeast Asia (c,f) for 1988. Temperature is the surface air temperature 2m above the surface and is expressed in Kelvins (K) with a contour interval of 0.25 K.
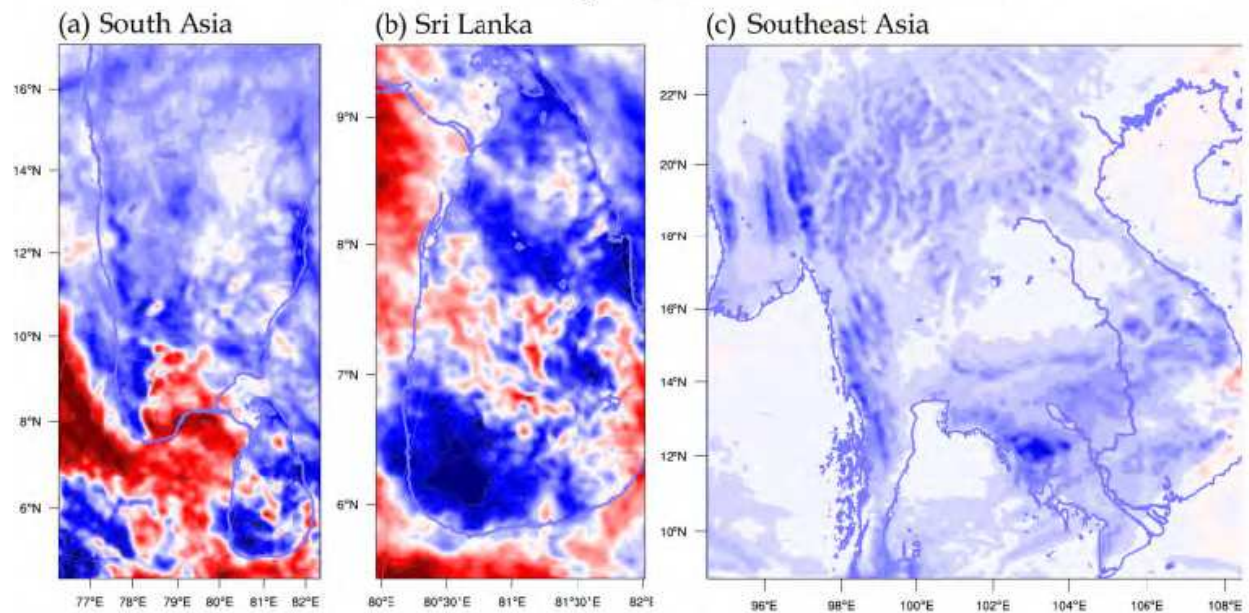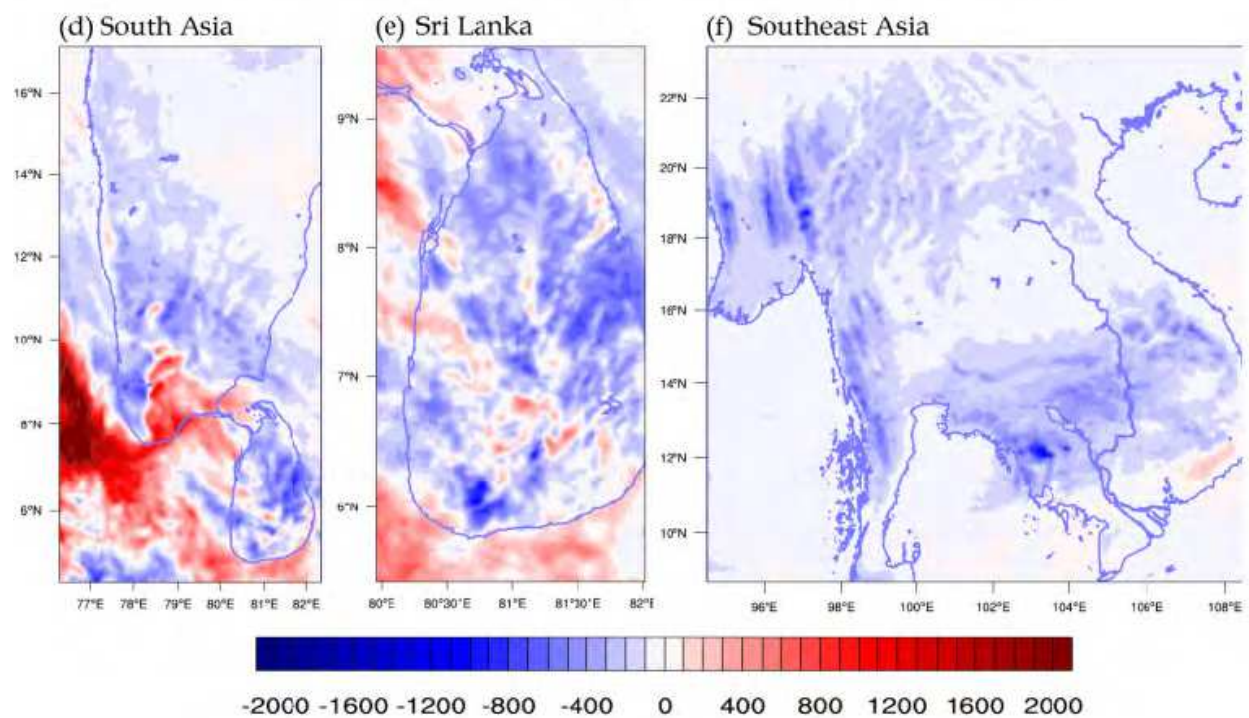
Monthly Temperatures
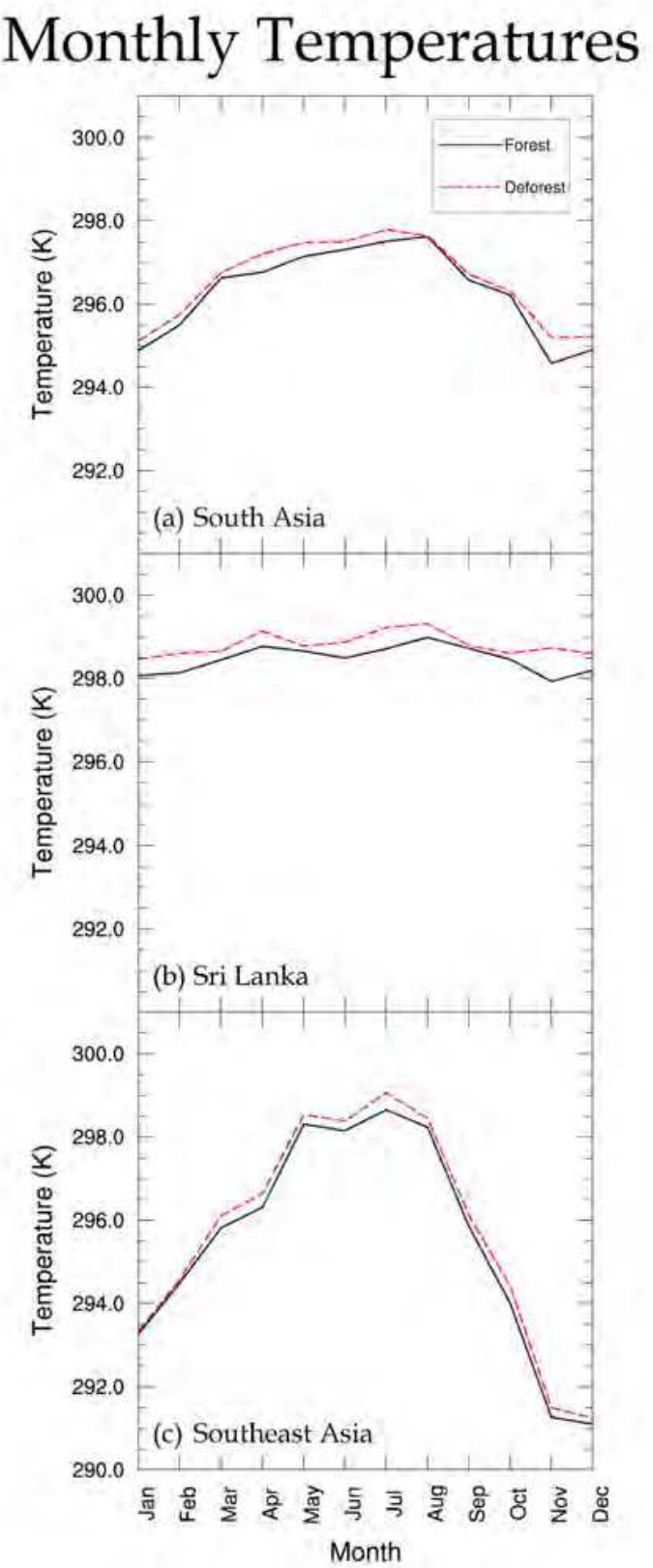
Fig. 2. The time series of temperature for 1988. The monthly temperature values over land regions (excluding oceans) for South Asia (a), Sri Lanka (b) and Southeast Asia (c) are shown here. Temperature is the air temperature 2m above the surface and is measured in Kelvin (K)

Fig. 3. The total annual and the seasonal (June, July and August) precipitation anomalies (i.e deforest –forest) for South Asia (a,d), Sri Lanka (b,e) and Southeast Asia (c,f) for 1988. Precipitation is in millimeters (mm) with a contour interval of 100 mm.
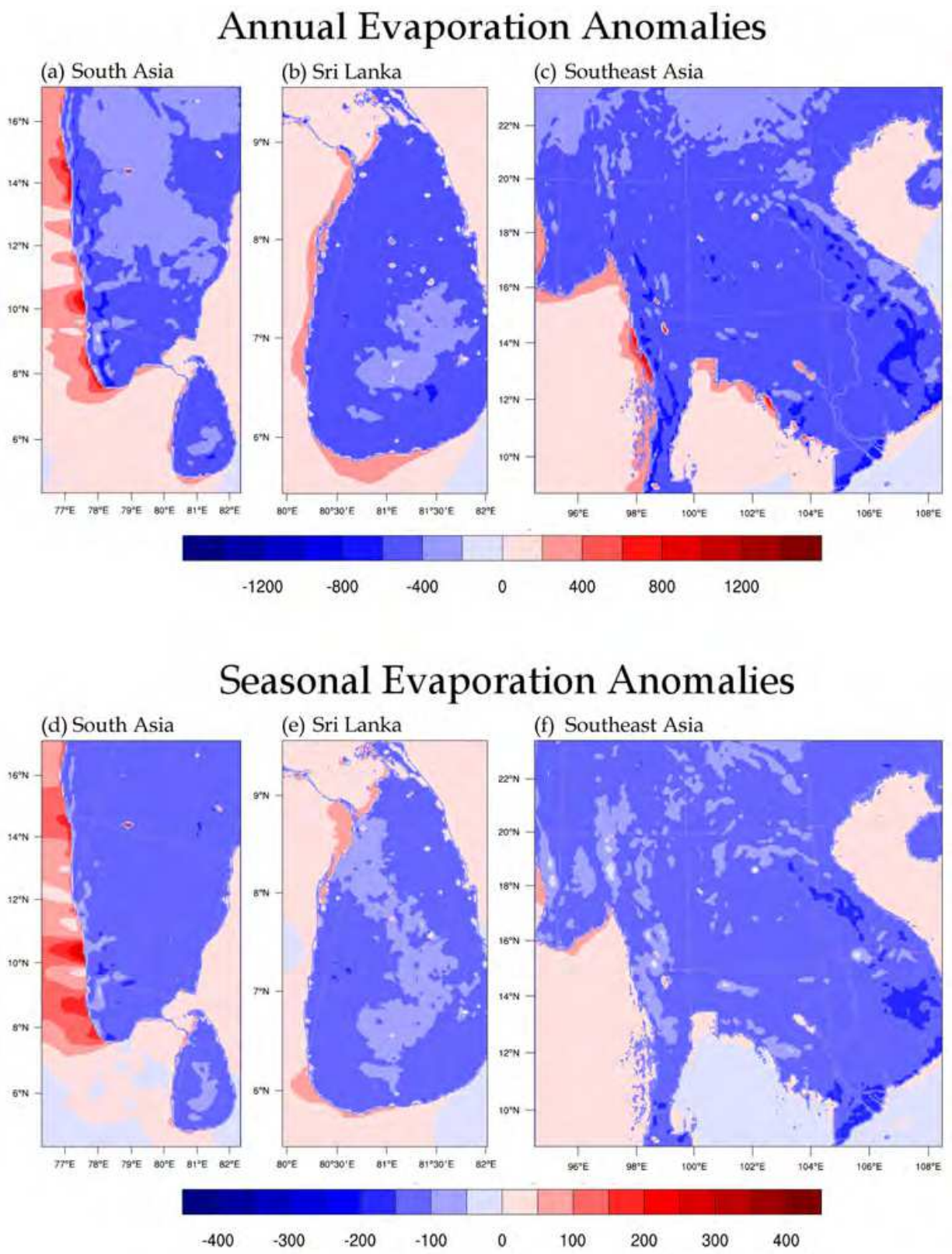
Fig. 4. The annual and the seasonal (June, July and August) evaporation anomalies (i.e deforest –forest) for South Asia (a,d), Sri Lanka (b,e) and Southeast Asia (c,f) for 1988. Evaporation is in millimeters (mm) with a contour interval of 200 mm for annual plot and 50 mm for the seasonal plot.

4. Discussion

Deforestation leads to warmer and drier climatic conditions. As a result of deforestation less moisture is available at the land surface, leading to a reduction in evapotranspiration. This in turn leads to an increase in temperature and a decrease in precipitation over land. These changes then work together to alter the moisture convergence in the region. These changes are seen not only during the monsoon season but through the entire year.

The monsoon is primarily driven by the pressure gradient created by the differential heating of the ocean and land. Land regions especially the Tibetan plateau warms up more than the Indian Ocean during the summer due to the small heat capacity of land. This creates a thermal low pressure over land and the resulting pressure gradient between ocean and land initiates the monsoon flow. Subsequently, the monsoon is sustained by the release of latent heat into the atmosphere. Therefore warmer temperatures should intensify the monsoon, increasing precipitation over land. But this is not seen in the results. The reasons for this are regionally specific. Over Sri Lanka it is likely that the warming of the land itself is reducing the amount of precipitation. The initial step in the formation of precipitation is the rising of parcels of warm moist air. The air being warm and moist by itself is not sufficient to cause rising motion. For air to rise the density of a parcel of air has to be lower than that of the surrounding air, that is the parcel temperature must be warmer than that of the surrounding (cooler) air. When deforestation warms the air over land, it reduces this difference, creating a more stable atmosphere which inhibits the rising of air and hence also weakens the atmospheric moisture convergence. Therefore even if the atmosphere otherwise holds plenty of moisture this process reduces the amount of precipitation that can form. Therefore the weakened moisture convergence together with the reduced evaporation causes the reduction in precipitation. The conditions are different in the Indian subcontinent where the summer monsoon is at its' strongest. The warming of the land region now results in a stronger moisture convergence indicating the possibility that the monsoon might become stronger over the region. It is possible that the warming over the Indian subcontinent (which has a smaller magnitude than the warming in Sri Lanka) is insufficient to stabilize the atmosphere. But as discussed in the results section the reduced evapotranspiration at any rate plays the dominant role in reducing the annual and the seasonal precipitation.

The changes in the climatic conditions due to deforestation can have a strong impact on society. Warmer and drier conditions can have far reaching consequences in many different ways. The elevated temperatures alone can cause life loss especially if the number of heat waves increases. India would be more susceptible to this as parts of the country experience high summer temperatures prior to the onset of the monsoon, a time during which people are vulnerable to the excess heat. The warm temperatures can also kill livestock and destroy stored goods.

The reduction in precipitation in combination with the warmer temperature can have a negative impact on agriculture in the region. These climatic changes can result in reduced crop yields, a shift in the life cycle of the crop and change the length of the growing season and the timing of the harvest. Also the regions that are the best suited for growing crops may migrate. As the warm and dry conditions act as positive feedbacks, deforestation can lead to persistent droughts. This may then result in desertification making some regions no longer viable as agricultural land and marginally habitable regions unsuitable for living. In

addition to this the changes in temperature and precipitation may change the habitats of animals and expand or alter the spread of diseases such as malaria.

Some regions depend largely on rainfall and/or stream flow for drinking water. If the decrease in precipitation is large enough, the lack of access to clean drinking water can become a major problem. In addition to this the water level in the rivers are important for transport along the rivers, generation of hydropower and irrigation of crops. If the regions that feed the river flow experience a decrease in precipitation it would lower the water levels regardless of what happens downstream causing many economic problems.

All these changes would lead to a shortage in food and a disruption in the well established livelihoods in the region leading to poverty and famine in extreme cases. They would also cause many health and safety issues to which the very young, the old and the poor would be the most sensitive.

Deforestation in South and Southeast Asia has an impact on the monsoonal climate. These climatic impacts then in turn will cause social, economical, environmental and health problems in the region. It is important to notice that Southeast Asia still has a significant amount of forest whereas the situation in south Asia is closer to the conditions of the deforested simulations. Therefore it is imperative to set in place, policies that would prevent or slow down deforestation in the region as well as to implement steps to deal with issues that have already arisen due to deforestation.

This study illustrates the maximum possible range of changes possible due to deforestation and shows that the signal due to deforestation can override even a strong monsoon. Therefore it is worthwhile to continue this study, focused on more realistic land-use changes and to examine the magnitude of the changes. Also future studies would include the changes to the Hadley cell and the occurrence of extreme events, especially floods, under both realistic and idealized situations.

5. Conclusion

Deforestation impacts the climate by modifying surface energy and moisture budgets. These modifications are mostly due to the decrease in evapotranspiration and increase in the surface albedo. The changes in surface albedo and evapotranspiration have competing effects on the temperature, but as seen from the results of the WRF runs the warming effect due to the reduced evapotranspiration dominates over the cooling effects of the reduced albedo.

Results show that majority of the land areas would become warmer and drier in response to deforestation, with precipitation, evapotranspiration and cloud cover all showing a decrease. Atmospheric moisture convergence shows regionally specific changes. These changes are seen in both the annual and monsoon seasonal values, suggesting that the changes that take place due to deforestation have the ability to override even a strong monsoon signal. The changes over the oceans are more variable, with an increase in evaporation seen in both seasonal and annual values in the domains over South Asia and Sri Lanka.

These changes due to deforestation can have far reaching social, economic and environmental impacts as well as cause serious health issues. Warm temperatures can cause illness and even heat related deaths. The decrease in precipitation can lead to the drying of

natural springs and reduced stream flow, cutting of access to clean drinking water in some regions. The habitats of plants and animals can change resulting in spread of diseases. Further the warm dry conditions can reduce or destroy crops, kill livestock and decrease the output from hydro powered electricity plants.

The climatic changes due to deforestation have the very real potential to impact the economic and the social structures of a country. As deforestation has been an ongoing problem, a climatic signal may already be present and the consequences of the changes already affecting us to some extent. If such a signal is present further deforestation may amplify these changes to a level that they would be clearly evident. Therefore it is important to understand the impacts of human activity on the climate, and set in place policies not only regarding deforestation but also implement steps to understand and deal with the consequences of a climatic signal that can result from current land-use practices.

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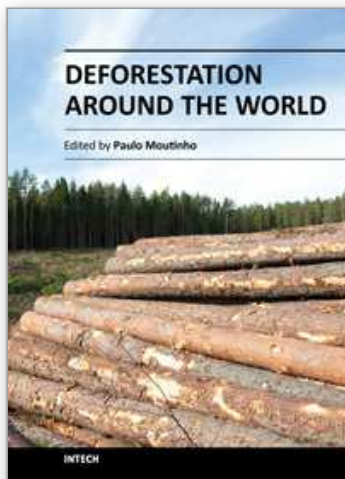
Deforestation Around the World

Edited by Dr. Paulo Moutinho

Deforestation and forest degradation represent a significant fraction of the annual worldwide human-induced emission of greenhouse gases to the atmosphere, the main source of biodiversity losses and the destruction of millions of people's homes. Despite local/regional causes, its consequences are global. This book provides a general view about deforestation dynamics around the world, incorporating analyses of its causes, impacts and actions to prevent it. Its 17 Chapters, organized in three sections, refer to deforestation impacts on climate, soil, biodiversity and human population, but also describe several initiatives to prevent it. A special emphasis is given to different remote-sensing and mapping techniques that could be used as a source for decision-makers and society to promote forest conservation and control deforestation.

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