Regional Impacts of Future Land-Cover Changes on the Amazon Basin Wet-Season Climate

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ABSTRACT

State-of-the-art socioeconomic scenarios of land-cover change in the Amazon basin for the years 2030 and 2050 are used together with the Regional Atmospheric Modeling System (RAMS) to simulate the hydrometeorological changes caused by deforestation in that region under diverse climatological conditions that include both El Niño and La Niña events. The basin-averaged rainfall progressively decreases with the increase of deforestation from 2000 to 2030, 2050, and so on, to total deforestation by the end of the twenty-first century. Furthermore, the spatial distribution of rainfall is significantly affected by both the land-cover type and topography. While the massively deforested region experiences an important decrease of precipitation, the areas at the edge of that region and at elevated regions receive more rainfall. Propagating squall lines over the massively deforested region dissipate before reaching the western part of the basin, causing a significant decrease of rainfall that could result in a catastrophic collapse of the ecosystem in that region. The basin experiences much stronger precipitation changes during El Niño events as deforestation increases. During these periods, deforestation in the western part of the basin induces a very significant decrease of precipitation. During wet years, however, deforestation has a minor overall impact on the basin climatology.

1. Introduction

The Amazon tropical forest covers a region of nearly six million square kilometers. It receives an average rainfall of 2500 mm yr$^{-1}$ (Richey et al. 1991), and generates an annual discharge of over a trillion cubic meters of water into the Atlantic Ocean (Vital and Steattegger 2000). Thus, it is an important component of the earth’s hydroclimate (Avissar and Werth 2005; Werth and Avissar 2002). Anthropogenic activity in the region results in a gross rate of deforestation on the order of 2$\times$10$^4$ km$^2$ yr$^{-1}$ (Achard et al. 2002), and the local, regional, and global hydroclimatic changes resulting from this replacement of natural forest by degraded vegetation has yet to be fully understood and quantified. Observations collected during the Large-Scale Biosphere–Atmosphere (LBA) experiment in Amazonia, together with state-of-the-art numerical models, are employed for this task (Avissar and Nobre 2002).

Several studies have been conducted with global climate models (GCMs) to estimate the impacts of Amazonian deforestation on the regional and global climate. These studies have shown that deforestation reduces local rainfall and increases surface air temperature (Dickinson and Kenneday 1992; Henderson-Sellers et al. 1993; Nobre et al. 1991; Shukla et al. 1990) and that massive deforestation affects the climate of other regions (Avissar and Werth 2005; Werth and Avissar 2002). However, GCMs have grid sizes of about 2°–5° latitude and longitude and, consequently, many key land–atmosphere interactions necessary for simulating the region hydroclimate properly are not resolved (Avissar et al. 2002; Ramos da Silva and Avissar 2006). Furthermore, experiments performed with such models generally assume a complete replacement of the entire Amazon rain forest with degraded vegetation, such as pasture. Moreover, current subgrid-scale parameterizations in GCMs represent convection and microphysical cloud processes based on the large-scale mean state of the atmosphere, which leads to errors in rainfall modeling (Molinar and Dudek 1992). With the increasing capability of computing systems, it is now feasible to study the effects of deforestation on the climate with
regional climate models set up with grid sizes on the order of 20–30 km. Obviously, at these scales, both weather and land-cover patchiness are better resolved. Another benefit is the availability of improved cumulus parameterizations designed to operate at this resolution (Kain 2004).

Mesoscale models show that surface heterogeneity caused by deforestation can induce local circulations and, as a result, can increase convection during the dry season (Baidya Roy and Avissar 2002; Silva Dias and Regnier 1996; Souza et al. 2000; Wang et al. 2000), and even during the wet season (Silva Dias et al. 2002). These circulations correlate well with clouds observed from satellites (Chagnon and Bras 2005; Cutrim et al. 1995; Durieux et al. 2003).

Given these findings, Avissar et al. (2002) have hypothesized possible changes in rainfall accumulation resulting from a progressive expansion of deforestation. Accordingly, the following three scenarios could be considered: (a) rainfall will decrease linearly with deforested area; (b) rainfall will first decrease rapidly and then more slowly as the deforested area expands; and (c) rainfall will first remain unaffected (or possibly even increase) as a result of initial deforestation, but will then fall rapidly as the deforested area crosses some threshold, estimated at 30%–50% of deforestation.

Recently, Soares-Filho et al. (2004) produced scenarios of Amazonian deforestation based on the planned paving of highways, river channeling, port improvements, and expansion of energy production derived from spatial logistic regression of historical patterns of land cover, econometric analysis, and policy interventions. Figure 1 presents the land-cover types derived from these scenarios for the years 2030 and 2050. They emphasize a pattern of deforestation that is quite different from the massive deforestation assumed in typical GCM simulations, and that can be resolved by regional climate models.

The goal of the study summarized in this paper was to evaluate the impacts of these land-cover change scenarios on the wet-season climate of the Amazon basin. For this purpose, we used the Regional Atmospheric
Modeling System (RAMS), a state-of-the-art regional climate model forced with actual meteorological conditions during a 4-yr period that includes an El Niño–La Niña sequence. For comparison, we also used the National Aeronautics and Space Administration (NASA) Goddard Institute for Space Studies (GISS) GCM.

2. Wet-season rainfall overview

During the wet season, rainfall in the Amazon basin is produced by various meteorological mechanisms with both small- and large-scale characteristics. Locally, the strong heating of the surface produces convection, which can lead to development of cumulus clouds, thunderstorms, and rainfall. During the LBA campaign it was found that the local initiation of convective clouds occurred preferentially over high elevations (Laurent et al. 2002). At the regional scale, synoptic-scale Rossby waves can propagate northward, causing baroclinic instability and producing precipitation (Garreaud and Wallace 1998). Squall lines, which extend and propagate at scales varying from the meso- to the macroscale, are usually generated along the northeast coast of South America and propagate inland up to the Andes. An example of such squall lines is depicted in Fig. 2 for a typical day during the rainy (wet) season. They can be as long as 2000–3000 km and are believed to be the major mechanism for rainfall accumulation during the rainy season (Greco et al. 1990). During the LBA–Wet Atmospheric Mesoscale Campaign (WetAMC), Rickenbach (2004) found that these squall lines move at a speed of about 10–15 m s\(^{-1}\) and are an important source for nocturnal rainfall. At larger scales, the position and strength of the intertropical convergence zone (ITCZ) leads to different rainfall regimes in the basin (Nobre and Shukla 1996). Also, El Niño conditions can disturb the Walker circulation and cause subsidence in the Amazon, affecting the rainfall spatial distribution (Richey et al. 1989).

Figure 3 displays the basin-averaged accumulated rainfall in January and February combined for the period between 1970 and 2000. On average, rainfall accumulation is 530 mm. During El Niño events, such as in 1998, one can notice a quite significant reduction of rainfall. In 1999 and 2000 the mean accumulations were very similar but, as illustrated in Fig. 4, the spatial distribution was quite different. Although rain falls almost everywhere in the basin during the rainy season, some regions receive more precipitation, for example, the northeast coast near the Amazon River mouth, and the western and south-central regions. During the 1998 El Niño, accumulation was considerably reduced in the northern and western parts of the basin (Fig. 4b). By contrast, during the La Niña event of 1999, higher accumulation was observed, mainly in the west (Fig. 4c). Thus, climatological regimes not only affect the mean precipitation in the basin, they also shift its spatial distribution.

3. Numerical experiments

To better understand the impacts of deforestation on the future of the Amazon basin hydrometeorology, the Regional Atmospheric Modeling System (RAMS) is used to simulate the months of January and February.
(i.e., a 60-day period starting on 1 January), using projected land surface data from the Soares-Filho et al. (2004) study for the years 2030 and 2050 (see Fig. 1). In addition, a total deforestation scenario is also simulated. Such a scenario is speculated to occur toward the end of the twenty-first century based on current rates of deforestation. While this speculation is clearly debatable, it is nevertheless an interesting case to consider for comparison with earlier GCM studies as well as to bracket the potential impacts of deforestation. A “reference” case that uses current (as of 2000) land-cover types is also simulated.

RAMS is a three-dimensional numerical model that is based on the full set of primitive dynamical equations that govern atmospheric motions (Pielke et al. 1992). These equations are supplemented with parameterizations for turbulent diffusion (Mellor and Yamada 1974); solar and terrestrial radiation (Chen and Cotton 1987); moist processes, including the formation and interaction of clouds and precipitating liquid and ice hydrometeors (Meyers et al. 1997; Walko et al. 1995); and the kinematic effects of terrain, cumulus convection (Kain 2004), and sensible and latent heat exchange between the atmosphere, multiple soil layers, a vegetation canopy, and surface water (Avissar and Pielke 1989; Walko et al. 2000).

RAMS is fundamentally a limited-area model, but may be configured to cover an area as large as South America for simulating meso- and macroscale atmospheric systems. There is no lower limit to the domain size or to its resolution; microscale phenomena, such as tornadoes and boundary layer eddies, as well as submicroscale turbulent flow over buildings and in a wind tunnel, have been simulated with this code. Two-way interactive grid nesting in RAMS allows local fine-mesh grids to resolve microscale atmospheric systems, such as thunderstorms, while simultaneously modeling the large-scale environment of the systems on a coarser grid. RAMS is capable of simulating the convective atmospheric boundary layer (Avissar et al. 1998), the mesoscale systems (Baidya Roy and Avissar 2002; Ramos da Silva and Avissar 2006; Silva Dias et al. 2002), and

![Average accumulated rainfall](http://journals.ametsoc.org/jcli/article-pdf/21/6/1153/3947630/2007jcli1304_1.pdf)
the climate of various regions (Gandu et al. 2004; Hasler et al. 2005; Liston and Pielke 2001; Takle et al. 1999) quite well.

For this study, RAMS was set up with a grid of 210 × 150 horizontal elements, each representing a 20 km × 20 km area, centered at 5°S, 60.3°W(Fig. 5). With such a grid size, the large squall lines that are believed to be responsible for more than 80% of the rainfall accumulation during the wet season (Greco et al. 1990) are at least partly resolved. The grid is stretched vertically, starting at 150 m above ground and progressively increasing up to a size of 1.5 km near the top of the model domain, which is set at a height of 23.3 km. The soil numerical grid consists of 12 vertical layers that are also stretched from the surface down to a depth of 4 m. Dawson (1993) and Nepstad et al. (1994) found that vegetation in the Amazon basin can uptake water from deep soil layers, thus requiring the grid selected here. The integration time step of the model is 30 s.

The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) and the National Oceanic and Atmospheric Administration (NOAA) weekly sea surface temperature (SST; Reynolds et al. 2002) from the years 1997, 1998, 1999, and 2000 are used to provide the atmospheric and oceanic initial and boundary conditions, respectively. These years are chosen because they offer a broad variability of climate regimes, including El Niño (in 1998, when the Amazon precipitation was much lower than normal) and La Niña (in 1999, when wetter conditions prevailed). We argue that using realistic atmospheric conditions that include the various climate regimes (i.e., El Niño, La Niña, and “normal” years) is an appropriate alternative to a GCM-based output of future (uncertain) climate conditions. This is particularly true given the difficulty of GCMs in predicting regional climate (as confirmed later in this paper) and, consequently, the appropriate forcing for a regional model. Yet, we realize that our approach is not flawless either. Indeed, while the weather sequence we selected for our experiments is quite variable, it fails to account for the frequency of El Niño, La Niña, and “normal” years under a changing climate and their cumulative effects on the hydroecology of the region.

Initial conditions for soil moisture are based on a profile measured in Rondônia on 6 February 1999 during the LBA-WetAMC (Alvalá et al. 2002). Accordingly, at the first level below the surface, the initial soil moisture is set to 39% of saturation, increasing linearly to 45% of saturation at the deepest layer. This profile is adopted for all of the numerical experiments and for the entire domain. This is because, during the wet season, rain is widespread over the Amazon and soil moisture is usually high in most of the basin. We also attempted to use the large-scale soil moisture produced by the NCEP–NCAR reanalysis (Kalnay et al. 1996). However, the low water content simulated with that model resulted in stressed vegetation in RAMS associated with an unrealistic stomatal closure. As a result, RAMS produced unrealistically dry hydroclimatic conditions with this initialization. A sandy–clay–loam soil texture was assumed in accordance with maps from the Food and Agriculture Organization (FAO).

To represent the Amazonian vegetation, the model uses several parameters estimated from in situ measurements (Gash and Nobre 1997), which were tested in recent numerical experiments (Gandu et al. 2004; Ramos da Silva and Avissar 2006).

In the version used for this study, the NASA GISS GCM has 12 vertical layers and a horizontal grid size of 5° longitude × 4° latitude. Heat and humidity are advected with a quadratic upstream scheme, and momentum is advected with a second-order scheme. The model has both shallow and deep convection and a second-order closure atmospheric boundary layer scheme for moisture and heat transfer. The model uses six soil layers and a hydrology scheme that accounts for soil moisture transfer and root extraction (Rosenzweig and Abramopoulos 1997), the latter of which depends on the vegetation specified within a grid element. The model has been used recently to study the impacts of total deforestation of tropical rain forest regions on hydroclimatic changes (Avissar and Werth 2005; Werth and Avissar 2002), and has proven to be quite reliable in simulating the hydroclimate of these regions (Avissar and Werth 2005).
4. Results

a. Reference case

Figure 6 presents the statistics for the January–February rainfall accumulation simulated for the reference case and for the observations. The three selected subregions (west, east, and south-central) illustrated in Fig. 5 for the years 1997, 1998, 1999, and 2000 are considered separately. Overall, the statistics of the central tendency (i.e., median) for the simulations compare well with those of the observations. Observations and the model agree that 1998 was the driest year, while 1997 and 2000 were wet years.

In the east, at the boundary between the continent and the Atlantic, simulated rainfall is underestimated. We first assumed this was due to the proximity of this region to the boundaries of the domain, but tests performed using a larger domain resulted in similar discrepancies. Another possible reason for such shortfall could be the horizontal grid resolution used in the model. This region consists of a complex land cover, including several islands, river–and ocean–land boundaries, and rain forest, deforested, and agricultural areas. A higher resolution may be required to better simulate this complex environment. Indeed, Ramos da Silva and Avisser (2006) showed that a grid cell not larger than 1 km would be required to properly simulate moist convection in the Amazon. Unfortunately, it is not yet practical with currently available state-of-the-art computing systems to produce seasonal, regional-scale simulations with a grid size of 1 km. Last, discrepancies between observations and the reference case could result from a scarcity of observations in the region used to create the half-degree gridded dataset used here. The more heterogeneous the precipitation pattern in the region, the larger these discrepancies are likely to be.

In the west, the model overestimates the year-to-year variability of precipitation, while in the south-central region, simulated rainfall compares very well with observations. A good simulation of the latter region is particularly important, because it is expected to experience most of the deforestation increase in the coming years (Fig. 1). The interquartile range also compare well with observations (Fig. 6).

Squall lines move along with the low-level flow westward and are a key mechanism in the water cycle of the basin. Figure 7 illustrates some of these squall lines simulated with RAMS, emphasizing the ability of this model to resolve these important meteorological events. Thus, in general, these results demonstrate that the model can reasonably simulate the regional hydroclimatology of the Amazon basin.

b. Year 2030

The comparison of the 4-yr (1997–2000) average rainfall accumulation between 2030 and that obtained with the current land cover (Fig. 8) indicates a large decrease in the deforested areas and an increase in the nondeforested areas of the southwestern (near the Andes) and eastern parts of the simulated domain, where topography is generally higher. Indeed, the overall spatial correlation coefficient shows that topography explains 11% of the variance in accumulated precipitation when the latter is simulated with current vegetation cover. This increases to 12% for the 2030 deforestation scenario, suggesting that when the forest is eliminated, topographic effects become relatively more important. As a result of this partial deforestation, the domain-averaged air temperature has barely increased,
wind speed has increased by about 20%, and latent heat flux has decreased by about 5% (Table 1). The sensible heat flux, however, has practically not changed and the Bowen ratio rises only slightly.

c. Year 2050

The expansion of deforestation in 2050 reduces the rainfall even more and over larger areas, mostly over the deforested regions (Fig. 8). Similar to the scenario of 2030, rainfall increases in some nondeforested areas, over the Andes, and in the eastern part of the basin. Average wind speed increases by $\approx 35\%$ and latent heat flux decreases by a few more percent as compared to that in 2030 (Table 1). There is a consistent decrease in the sum of latent and sensible heat fluxes, emphasizing the combined importance of the albedo increase and the increase of longwave radiation emitted by the earth’s surface due to deforestation. The nearly constant Bowen ratio shows that, for these conditions, the partition of energy between sensible and latent heat flux remains the same (Table 1). Somewhat unexpectedly, sensible heat flux decreases. Sensible heat fluxes simulated with RAMS show that the predominant changes are strongly associated with cloud cover distribution changes.

Figure 9 indicates that, as a result of deforestation, wind speed intensifies, especially over deforested regions. In these regions, tall forest (35 m) is replaced with short grass (0.25 m) and, therefore, causes a decrease in the roughness length. The northeastern part of the basin that is “exposed” to the ocean is the most affected by this drastic land-cover change. There, the average northeasterly wind, which was about 7 m s$^{-1}$ before deforestation, increased to about 11 m s$^{-1}$ following it.

d. Total deforestation

Complete deforestation intensifies the rainfall shift in the region by decreasing its amount in the western part of the basin even further and increasing it in the eastern part and over the Andes (Fig. 8). Under such extreme conditions, the rainfall distribution follows an east–west dipole pattern, decreasing downstream toward the western Amazon and increasing at high elevations in
the Andes and in the southeastern part of the simulated domain. Indeed, the overall spatial correlation coefficient shows that topography now explains 14% of the variance in accumulated precipitation. Thus, an important part of the rainfall occurs over steep topography, where vertical motion favors convection. The precipitation reduction in the western part of the basin suggests that deforestation may alter the propagation of squall lines into that region, as is illustrated in Fig. 10. Indeed, following deforestation the propagation is slowed down compared to the nondeforested case, considerably reducing an important source of water for the western Amazon basin. These westward-propagating tropical squall lines have sloping moist updrafts ahead of their motion (Emanuel 1994; Garstang and Fitzjarrald 1999) that cause instability and require moist air from the surface that is supplied by the rain forest evapotranspiration. Thus, a decrease in latent heat flux caused by deforestation inhibits this mechanism and, as a result, their propagation.

In their study of African deforestation, Zheng and Eltahir (1997) explained that the spatial gradient of surface heat fluxes generated by the deforestation affected large-scale circulations. The studies of Werth and Avisar (2002, 2005a,b) and Avisar and Werth (2005) confirmed that tropical deforestation in general (not only in Africa), had such an impact. To explore this mechanism in more detail, some of the RAMS-simulated fields at 5°S for the January–February 1997 period are presented in Fig. 11. Near the surface, the prevailing winds are easterly ($u < 0$, see also Fig. 9). After deforestation, these easterly winds strengthen ($\Delta u < 0$), partly due to the reduced roughness length as tall trees are replaced with pasture. However, maybe even more importantly, this strengthening can be explained by the increased horizontal pressure gradient between the ocean and the deforested (warmer) land. At an about 3-km height, however (near the bottom of the cloud layer), we notice that the easterly winds weaken ($\Delta u > 0$) as a westerly (positive) perturbation is imposed upon them. This occurs mostly in the eastern high-elevated regions, where convection increases.

The sum of sensible and latent heat flux is a measure of energy infused into the atmospheric boundary layer. One can see that deforestation causes a reduction in the zonal gradient of that energy (Fig. 11b): Initially (i.e., for the current landscape) the flux is greater in the west as compared to the east, but after deforestation, the flux is reduced in the west. Zheng and Eltahir (1998) emphasized that such a flux spatial gradient induces a thermally driven circulation, which here produces subsidence in the west and increases convection in the east. Note that the positive wind anomaly in Fig. 11a represents the lower branch of the return flow. Therefore, the sensible heat flux at this latitude rises over the drier west and falls over the wetter east (Fig. 11c), and the precipitation increases over the eastern part of the do-
This offsets the changes over the deforested region when the basin as a whole is considered.

Figure 12 indicates that sensible heat flux is positively correlated with the downward solar radiation reaching the surface, and is negatively correlated with cloud cover, but is not strongly correlated with the surface albedo. The spatial correlations between sensible heat flux and clouds remain high even after total deforestation, emphasizing the important cloud feedbacks with the surface energy balance.

Surface air temperatures increase over the western part of the basin (Fig. 13). This is associated with the decrease in rainfall that reduces soil moisture and, consequently, reduces the latent heat flux into the atmosphere from the land surface. In the east, cloud cover and rainfall increase, leading to a decrease in the downward solar radiation flux reaching the surface. This emphasizes the important regional feedbacks between clouds and the energy balance at the surface.

### Table 1. Spatially averaged (between 2°N and 14°S, and 45° and 72°W) hydrometeorological variables simulated with RAMS for January–February in the Amazon basin. Bold lines highlight the mean values for each variable and scenario.

<table>
<thead>
<tr>
<th>Year</th>
<th>Current land cover</th>
<th>2030 scenario</th>
<th>2050 scenario</th>
<th>Total deforestation</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Accumulated rainfall (mm)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1997</td>
<td>564</td>
<td>564</td>
<td>543</td>
<td>525</td>
</tr>
<tr>
<td>1998</td>
<td>300</td>
<td>259</td>
<td>245</td>
<td>256</td>
</tr>
<tr>
<td>1999</td>
<td>456</td>
<td>434</td>
<td>421</td>
<td>370</td>
</tr>
<tr>
<td>2000</td>
<td>561</td>
<td>544</td>
<td>532</td>
<td>494</td>
</tr>
<tr>
<td><strong>Mean</strong></td>
<td><strong>470</strong></td>
<td><strong>450</strong></td>
<td><strong>435</strong></td>
<td><strong>411</strong></td>
</tr>
<tr>
<td><strong>Air temperature (°C)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1997</td>
<td>25.8</td>
<td>25.9</td>
<td>25.9</td>
<td>26.4</td>
</tr>
<tr>
<td>1998</td>
<td>30.3</td>
<td>30.9</td>
<td>29.5</td>
<td>30.3</td>
</tr>
<tr>
<td>1999</td>
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<td>26.3</td>
<td>26.4</td>
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<tr>
<td>2000</td>
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<td>25.4</td>
<td>24.1</td>
<td>25.5</td>
</tr>
<tr>
<td><strong>Mean</strong></td>
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<td><strong>27.1</strong></td>
<td><strong>26.5</strong></td>
<td><strong>27.2</strong></td>
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<tr>
<td><strong>Latent heat flux (W m⁻²)</strong></td>
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<td></td>
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<tr>
<td>1997</td>
<td>164</td>
<td>160</td>
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<td>139</td>
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<tr>
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<td>2000</td>
<td>167</td>
<td>163</td>
<td>159</td>
<td>147</td>
</tr>
<tr>
<td><strong>Mean</strong></td>
<td><strong>149</strong></td>
<td><strong>142</strong></td>
<td><strong>137</strong></td>
<td><strong>124</strong></td>
</tr>
<tr>
<td><strong>Sensible heat flux (W m⁻²)</strong></td>
<td></td>
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<td></td>
<td></td>
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<td>1997</td>
<td>55</td>
<td>53</td>
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<tr>
<td>1998</td>
<td>110</td>
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<td>71</td>
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<td>2000</td>
<td>54</td>
<td>52</td>
<td>51</td>
<td>47</td>
</tr>
<tr>
<td><strong>Mean</strong></td>
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<td><strong>72</strong></td>
<td><strong>71</strong></td>
<td><strong>64</strong></td>
</tr>
<tr>
<td><strong>Sensible + latent heat flux (W m⁻²)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
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<td><strong>Mean</strong></td>
<td><strong>221</strong></td>
<td><strong>214</strong></td>
<td><strong>208</strong></td>
<td><strong>188</strong></td>
</tr>
</tbody>
</table>

### Fig. 9. January–February mean wind speed differences (m s⁻¹) obtained at a height of 72 m between the RAMS simulation with the land cover of 2000 and that of 2050. The streamlines represent the mean wind direction for 2000.
FIG. 10. Cloud-cover fraction averaged between 1° and 2°S for selected periods of propagating squall lines for the (left) reference case and (right) total deforestation.
FIG. 11. Simulated January–February 1997 averages at 5°S of (a) zonal wind profile (m s\(^{-1}\)) (contours are for current vegetation and shaded regions are anomaly caused by total deforestation); (b) sensible plus latent heat flux (W m\(^{-2}\)) for the reference case (open circles) and the total deforestation scenario (closed circles); and (c) sensible heat flux difference (W m\(^{-2}\)) between the reference case and total deforestation (open circles, left scale), and topography (m) (closed circles, right scale).

FIG. 12. Daily spatial correlation coefficients between sensible heat flux and downward solar radiation at the surface (open circles), albedo (crosses), and cloud cover fraction (triangles) for the total deforestation case (black) and reference case (gray).

FIG. 13. January–February surface air temperature difference (°C) between the reference case and total deforestation. A contour line is plotted between positive and negative anomalies.
results highlight the importance of using high-resolution models to study land-cover change impacts on the hydroclimate, because various intricate feedback mechanisms are typically involved in the relevant processes.

e. Interannual variability

Figure 14 and Table 1 indicate that, with current vegetation, the spatial variance in accumulated precipitation is influenced somewhat by the topography, with
the explained variance equal to 6%, 28%, 10%, and 6% for 1997, 1998, 1999, and 2000, respectively, suggesting that under very dry conditions (i.e., 1998) topography exerts a stronger influence on precipitation. Under the 2030 deforestation conditions (Fig. 14), rainfall increases at the edges of the deforested areas for the wet case of 1997, and it compensates for the rainfall decrease in the deforested regions, resulting in a negligible domain-averaged change. In contrast, during the dry 1998 El Niño, rainfall is reduced over a much larger area in response to the same deforestation, causing an average decrease of about 14% over the basin. During such dry periods the vegetation is under water stress and the surface latent heat flux is very low as compared to that obtained in the wet years. As a result of this partial deforestation, the domain-averaged air temperature has barely increased, wind speed has increased, and latent heat flux has slightly decreased for all years.

As also seen in Fig. 14, the expansion of deforestation in 2050 reduces the rainfall even more and over larger areas, mostly over the deforested regions. Under complete deforestation conditions, the rainfall distribution follows an east–west pattern, decreasing downstream toward the western Amazon and increasing at high elevations in the Andes and in the southeastern part of the simulated domain. At this point, there is not much annual variance because the mitigating mechanisms during wet years become less effective. The spatial variance in accumulated precipitation explained by the topography has risen to 14%, 17%, and 10% for 1997, 1999, and 2000, respectively. For 1998, it has dropped to 17% due to a decrease in precipitation over larger areas. Overall, however, rainfall occurs mostly over steep topography where vertical motion favors convection, and the surface vegetation now has a reduced effect. Winds intensify and latent heat flux decreases, similar to what has been simulated with GCMs (Nobre et al. 1991).

f. GCM simulations

The following six scenarios are simulated with the NASA GISS GCM: 0%, 20%, 40%, 60%, 80%, and 100% deforestation. Each one is run for 48 yr. In the 0% scenario, the model simulates the rain forest as mapped by Matthews (1983) for the period starting in 1960 and ending in 1979. In the other scenarios, a percentage of that rain forest is converted into a mixture of grassland and shrubs (i.e., within each grid element, the fraction of that element covered by rain forest is reduced, and that same amount added to grassland/shrub). In the total deforestation scenario, all of the Amazon rain forest is converted into grassland/shrub. Observed monthly mean SSTs derived from multiyear climatological records from the Hadley Centre were used for all simulations.

Figure 15 and Table 2 summarize some of the results obtained from the simulations produced with the NASA GISS GCM. As has been seen in other GCM

### Table 2. Same as Table 1, but simulated with the GISS GCM.

<table>
<thead>
<tr>
<th>Percent of deforestation</th>
<th>0%</th>
<th>20%</th>
<th>40%</th>
<th>60%</th>
<th>80%</th>
<th>100%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Accumulated precipitation (mm)</td>
<td>390.5</td>
<td>382.36</td>
<td>360.89</td>
<td>343.46</td>
<td>324.73</td>
<td>316.49</td>
</tr>
<tr>
<td>Air temperature (°C)</td>
<td>24.36</td>
<td>24.37</td>
<td>24.50</td>
<td>24.63</td>
<td>24.87</td>
<td>25.06</td>
</tr>
<tr>
<td>Latent heat flux (W m⁻²)</td>
<td>104.96</td>
<td>101.99</td>
<td>98.62</td>
<td>94.51</td>
<td>89.42</td>
<td>84.84</td>
</tr>
<tr>
<td>Sensible heat flux (W m⁻²)</td>
<td>41.74</td>
<td>43.28</td>
<td>45.11</td>
<td>47.83</td>
<td>51.15</td>
<td>53.71</td>
</tr>
<tr>
<td>Sensible + latent heat flux (W m⁻²)</td>
<td>146.7</td>
<td>145.3</td>
<td>143.7</td>
<td>142.3</td>
<td>140.6</td>
<td>138.6</td>
</tr>
<tr>
<td>Bowen ratio</td>
<td>0.40</td>
<td>0.42</td>
<td>0.46</td>
<td>0.51</td>
<td>0.57</td>
<td>0.63</td>
</tr>
<tr>
<td>Zonal wind (m s⁻¹)</td>
<td>-1.05</td>
<td>-1.09</td>
<td>-1.08</td>
<td>-1.14</td>
<td>-1.15</td>
<td>-1.2</td>
</tr>
<tr>
<td>Meridional wind (m s⁻¹)</td>
<td>-0.65</td>
<td>-0.65</td>
<td>-0.64</td>
<td>-0.66</td>
<td>-0.67</td>
<td>-0.67</td>
</tr>
<tr>
<td>Albedo (%)</td>
<td>12.0</td>
<td>13.35</td>
<td>14.70</td>
<td>16.05</td>
<td>17.41</td>
<td>18.77</td>
</tr>
<tr>
<td>Moist convergence (mm day⁻¹)</td>
<td>3.01</td>
<td>2.98</td>
<td>2.73</td>
<td>2.58</td>
<td>2.44</td>
<td>2.45</td>
</tr>
</tbody>
</table>
The impact of deforestation on precipitation illustrated in Fig. 16 depicts inflection points with both the 
GISS GCM and RAMS simulations. These are especially noticeable in the western part of the basin. We 
premise that this is because this region is the one mostly affected by the nonlinear processes involved in 
the Amazon basin land–atmosphere interactions. Indeed, in addition to the local effects of deforestation, as 
explained above, this region is also affected by teleconnected hydroclimatic processes occurring upwind. 
Furthermore, as we had estimated in the past (e.g., Chen and Avisser 1994; Avisser and Liu 1996; 
Baidya Roy and Avisser 2002; Avisser et al. 2002), mesoscale effects optimize at a scale roughly equivalent 
to the local Rossby deformation radius. This corresponds
to a deforestation of \(\sim 40\%\). But, of course, GCMs do not resolve mesoscale processes. Deforestation in the GISS GCM is parameterized with a percentage of each grid point being changed to degraded vegetation. Thus, it is done in a linear way over the deforestation domain and, in that case, it is most likely the change of the ratio between advection and convection that generates these inflection points. This mechanism is also important in RAMS.

5. Summary and conclusions

The Regional Atmospheric Modeling System (RAMS) was used, together with socioeconomic scenarios of land-cover changes, to improve our understanding of the impacts of projected deforestation in the Amazon basin on its hydroclimatology. Simulations show that rainfall gradually decreases as the deforestation increases, but the magnitude of the impacts depends on...
TABLE 3. Moisture convergence and domain-averaged soil moisture content (m$^3$ m$^{-2}$) simulated with the RAMS at a depth of 0.1 m, and total groundwater (kg m$^{-2}$) simulated with the GISS GCM for the period January–February.

<table>
<thead>
<tr>
<th>Moisture convergence (mm day$^{-1}$)</th>
<th>Soil moisture</th>
</tr>
</thead>
<tbody>
<tr>
<td>Current land cover</td>
<td>Total deforestation</td>
</tr>
<tr>
<td>RAMS</td>
<td>2.83</td>
</tr>
<tr>
<td>GISS</td>
<td>3.01</td>
</tr>
</tbody>
</table>

The prevailing climate regime and there is a significant spatial variability in the response. Impacts are stronger under drier El Niño conditions, and are also more intense in the western and southern regions, downstream of the easterly trade winds (which intensify in the basin after deforestation). This leads to large areas of low precipitation, higher temperatures, and higher sensible heat fluxes in the basin. These results agree with those of Voldoire and Royer (2004), who used the Action de Recherche Petite Échelle Grande Échelle (ARPEGE)-Climat GCM and found that during El Niño periods, the effects of deforestation are enhanced.

The climate regime also controls the relationship between deforestation extent and the hydroclimatological response: wetter La Niña conditions lead to a linearly decreasing trend of precipitation as deforestation expands, but under dry conditions, a steep decrease may occur even for minor levels of deforestation. Furthermore, the geographical locations within the basin experience deforestation in a different way. Indeed, some regions (e.g., the high elevation of the eastern part of the basin) even see an increase in rainfall as a result of the basin deforestation. Because it has been speculated that the frequency of El Niño events could increase under the climate change expected to occur in the coming decades, our study indicates that the effects of deforestation on the Amazon hydrology are likely to be exacerbated.

Experiments performed with GCMs suggest that deforestation may establish a permanent savannah in the Amazon, mostly over the east (Oyama and Nobre 2003). Here, we show that progressive deforestation can affect the propagation of squall lines toward the west and with stronger impacts during the El Niño years. Thus, the combination of the large-scale effects on the drying of the eastern part of the basin and the impacts on the local mesoscale systems (i.e., the squall lines) may cause a much stronger effect if they occur during an El Niño year.

Our simulations emphasize that using a high-resolution model is quite valuable for this type of study, because resolving the dynamics associated with mesoscale hydrometeorological processes even partially (such as land–atmosphere–cloud interactions and squall lines, which are important during the wet season in the Amazon basin) significantly affects the regional climate.

Similar to the GISS GCM, RAMS shows that temperature and wind speed increase and latent heat flux decreases as a result of expanding deforestation in the Amazon basin. However, two important differences are noticed when comparing the two models. First, the response to deforestation obtained with RAMS is of lesser magnitude than that obtained with the GISS GCM. Second, while the GISS GCM displays an increase in the domain-averaged sensible heat flux, RAMS shows a decrease. Because RAMS shows spatial patterns of change in the sensible heat flux that correlate with changes in downward solar radiation and fractional cloud cover, we interpret this discrepancy between the models being as caused by the nonlinear land–atmosphere–cloud interactions resolved in RAMS, but neither resolved nor parameterized in the GCM. These interactions, which are discussed in detail in Ramos da Silva and Avisser (2006), induce positive precipitation anomalies and explain why the effects of deforestation are less evident during dry years as compared to dry years, when local circulations and cumulus clouds are better correlated with landscape heterogeneities. Thus, current GCMs may overestimate the effects of deforestation during the wet season.

In the numerical experiments performed here with RAMS, deforested areas were assumed to be replaced with pasture. More realistic scenarios need to include other land use types, such as crop fields, artificial lakes, and urban areas. Also, on the long term, vegetation dynamically adapts to the new environmental conditions. For instance, Laurance et al. (2004) found a significant acceleration of the rate of tree mortality, recruitment, and growth in very remote regions of the Amazon rain forest, and that such anomalies seem to be associated with recent increases in temperature and CO$_2$ concentration. Thus, this type of interaction also needs to be accounted for in climate models because their feedbacks on the hydroclimate system are not obvious. This can be achieved by implementing state-of-the-art dynamical vegetation schemes in the climate models, such as the Ecosystem Demographic Model (ED) developed by Moorcroft et al. (2001).
Finally, the size of the horizontal grid elements used in the RAMS simulations produced for this study was 20 km. At this resolution, a cumulus parameterization is still required because these important clouds (and the precipitation that they produce) are not properly resolved. Ramos da Silva and Avisser (2006) have shown that to explicitly resolve convection in this region, a grid size not larger than 1 km should be used. Given the importance of the nonlinear land–atmosphere–cloud interactions pointed out in this study, one could speculate that resolving convective clouds and precipitation might uncover additional feedbacks not simulated here. We are in the processes of performing such experiments and will report the results in a subsequent publication.

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