THE ROLE OF LARGE-SCALE LAND-USE CHANGE ON THE GLOBAL CLIMATE RESPONSE AND SENSITIVITY TO AMAZON DEFORESTATION

by

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The Role of Large-Scale Land-Use Change on the Global Climate - Response and Sensitivity to Amazon Deforestation

A dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy at George Mason University

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Dedication

To Mom and Dad.
For their endless love, support and encouragement.
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Abstract

THE ROLE OF LARGE-SCALE LAND-USE CHANGE ON THE GLOBAL CLIMATE - RESPONSE AND SENSITIVITY TO AMAZON DEFORESTATION

Andrew M. Badger, PhD
George Mason University, 2015
Dissertation Director: Dr. Paul A. Dirmeyer

Large-scale land-use change, such as Amazon deforestation, can have a significant local effect on the climate and has the potential to impact the global climate system. Previous modeling studies have shown non-local responses due to Amazon deforestation. However, a common flaw in these studies is using prescribed ocean conditions, which can dampen the global response. This study uses a set of fully coupled modeling simulations to determine the responses and sensitivities to Amazon deforestation, both locally and globally. In addition, a set of realistic tropical crop vegetation types are developed for the Community Land Model version 4.5. The local increase in surface temperature and decrease in precipitation from this study are consistent with previous modeling studies. After deforestation, it was determined that stronger regions of land-atmospheric coupling are found in the formerly densely forested regions, while areas that receive irrigation become less coupled. This study highlights large-scale changes to the zonal and meridional circulation that are found to have impacts in remote regions throughout the tropics. Lastly, using a set of partial deforestation simulations, areas of non-linear responses to deforestation are found. A metric to quantify the degree of non-linearity over the spatial domain was developed; Amazon deforestation is found to have a tipping point effect for the climate system with less than half of the impacts
within the spatial domain of influence being provided by 50% deforestation.
Chapter 1: Introduction

1.1 Background Information

Land-use change (LUC) has generally been considered a local environmental issue, but it is now becoming a force of global importance (Foley et al., 2005). There has been a rapid growth in human population and the associated development pressures have put a considerable strain on the Earth’s ecosystems (Snyder et al., 2004). LUC occurs on local scales, with real world social and economic benefits, but can potentially cause ecological degradation across local, regional, and global scales (Foley et al., 2005).

A large portion of the Earth’s surface has already been modified for urban and industrial development, agriculture, and pastureland (Snyder, 2010). Worldwide changes to forests, farmlands, waterways, and air are being driven by the need to provide food, fiber, water, and shelter to the more than six billion people inhabiting Earth (Foley et al., 2005). LUC has enabled humans to appropriate an increasing share of Earth’s resources. LUC has also potentially undermined the capacity of ecosystems to sustain food production, maintain freshwater and forest resources, regulate climate and air quality.

The future of tropical forests is at risk in a warmer, more populous 21st-century world (Bonan, 2008b). Forests cover approximately 42 million km$^2$ in tropical, temperate and boreal regions, which is approximately 30% of the Earth’s land surface. In the last two decades, the strong increase of pasturelands over former rainforest areas has raised concerns about the possible climate change impacts that such changes in land cover might have (e.g., Costa et al. 2007). LUC has the potential to have a significant impact on land-atmosphere interactions and modify local climate conditions (Sun and Wang, 2011). Snyder et al. (2004) acknowledges the fact that wide-scale vegetation removal is unrealistic for most of the biomes with the tropical forests being the lone exception.
Loss of natural forests worldwide in the tropics during the 1990s was as high as 152,000 km²/year, and Amazonian forests were cleared at a rate of approximately 25,000 km²/year (Bonan, 2008b). By 1991, 426,000 km² of the Amazon forest had already been removed, approximately 10.5% of the original forest area (Costa and Foley, 2000). More recent estimates suggest that by 2006, 663,177 km² of the Amazon forest had been removed (IBGE, 2006) with approximately an additional 60,000 km² deforested since 2006 (INPE, 2014).

Nepstad et al. (2008) notes that trends in Amazon economies, forests and climate could lead to the replacement or severe degradation of more than half of the closed-canopy forests of the Amazon Basin by the year 2030, even without including the impacts of fire or global warming. The total area of the Amazon that is suitable for industrial agriculture is 33%, suitable indicates areas with no edaphic or climate restrictions. Although cattle pasture remains the dominant use of cleared land, there is becoming a greater need for larger and faster conversion to cropland, mostly for soybean export which has defined the trend in forest loss in Amazonia since the early 2000s (Davidson et al., 2012). Bonan (2008b) points out that such continued land-use pressures can potentially shift the Amazonian region to a permanently drier climate once a critical threshold of clearing is reached.

It is clear that LUC is a danger for the Amazon region that can have drastic consequences due to the role that forests have in mediating the climate. Forests influence the climate through exchanges of energy, water, carbon dioxide, and other chemical species with the atmosphere (Bonan, 2008b). LUC has clearly played a role in changing the global carbon cycle and, possibly, the global climate (Foley et al., 2005). LUC can affect climate conditions through both biogeochemical and biogeophysical processes (Davin and de Noblet-Ducoudré, 2010).

One of the most important roles that forests have in the climate system is their function in the carbon cycle. Forests sequester large amounts of carbon annually, store approximately 45% of terrestrial carbon and contribute approximately 50% of terrestrial net primary production, a rate of carbon sequestration (Bonan, 2008b). Atmospheric analyses suggest that tropical forests are carbon neutral or carbon sinks, which implies potential offsetting of
carbon uptake by undisturbed tropical ecosystems. Bonan (2008b) also notes that carbon uptake by forests contributed to a residual 2.6 PgC/year terrestrial carbon sink in the 1990s, approximately 33% of anthropogenic carbon emission from fossil fuel and LUC; with deforestation releasing 1.6 PgC/year during the 1990s. Nepstad et al. (2008) further details that the trees of the Amazon contain 90-140 billion tons of carbon, equivalent to approximately 9-14 decades of current global, annual, human-induced carbon emissions.

Carbon and water cycles are intimately coupled in terrestrial ecosystems, and water-use efficiency (WUE), carbon gain at the expense of unit water loss, is one of the key parameters of ecohydrology and ecosystem management (Ito and Inatomi, 2011). Ito and Inatomi (2011) states that human activities have affected the water and carbon cycles and therefore the WUE of the terrestrial biosphere, and that LUC from forest to cropland can lower total WUE at broad scales.

Along with impacting the carbon cycle, as inferred by WUE, forests also help to regulate the hydrologic cycle. Forests help to sustain the hydrologic cycle through evapotranspiration, which cools climate through feedbacks with clouds and precipitation (Bonan, 2008b). Bagley et al. (2012) points out that multiple studies have indicated LUC has the potential to drastically reduce evapotranspiration. Climate model simulations show that tropical forests maintain high rates of evapotranspiration, decrease surface air temperature, and increase precipitation compared with pastureland (Bonan, 2008b). LUC can disrupt the surface water balance and the partitioning of precipitation into evapotranspiration, runoff, and groundwater flow (Foley et al., 2005). The Amazon is a critical region for investigation of water vapor transport, with approximately 15% of the Brazilian rainforest already converted to agriculture (Bagley et al., 2012). Through evapotranspiration, forests maintain atmospheric moisture that can return to land as rainfall downwind (Spracklen et al., 2012). Large-scale LUC may alter precipitation hundreds to thousands of kilometers from the region of vegetation change. With current trends of Amazonian deforestation and due to less-efficient moisture recycling caused by LUC, reductions of 12 and 21 percent in wet-season and dry-season precipitation respectively can be expected across the Amazon basin.
by 2050.

Of the remaining land surface, the tropical rainforests of South America, Africa, and Southeast Asia are among the ecosystems most at risk due to the demographic pressures (Snyder et al., 2004). The most studied region is Amazonia, where large-scale conversion of forest to pasture creates a warmer, drier climate (Bonan, 2008b). The large-scale LUC of the Amazon region and its impacts on the physical climate system are the focus of this review.

1.2 Overview of Previous Modeling Experiments

An overview of modeling experiments previously done can be seen in Table 1.1. The most recent experiments tend to have increased resolution and longer integration times. This a feature to be expected as computational resources has increased. The coarser resolution studies cannot resolve the local features of deforestation, but can give a reasonable representation of the regional scale changes. The increased resolution and the associated ability to resolve small-scale features is a desired ability in order to represent the local dynamics involved with deforestation. With increased length of integration, the capability to reach a new equilibrium climate is greatly enhanced; as well as larger sample sizes and more confidence in statistically significant results. Shorter simulations are likely missing some global features associated with Amazon deforestation that have not had a chance to develop in the model integration.

As can be seen in Table 1.1, only one experiment uses a full ocean model, another with an ocean model used from 40°S to 40°N. The remainder of the experiments use a mixed layer ocean model or prescribed SSTs (observed or climatological). There are several pitfalls in not coupling the model in these experiments. Snyder et al. (2004) notes that it is important that the ocean dynamics be realistically represented using a coupled atmosphere-ocean-biosphere model, there are many important feedbacks between the three systems that act to amplify the climate response to vegetation removal. The use of prescribed climatological or observed SSTs can act to dampen the model interannual variability, which could increase
Table 1.1: A summary of previous modelling experiments aimed at determining the impacts of Amazon deforestation.

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<th>Control Simulation</th>
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<td>Costa and Foley (2000)</td>
<td>GENESIS AGCM, 4.5° x 7.5°, 16 vertical levels</td>
<td>IBIS, 4.5° x 7.5°, 6 layers</td>
<td>50-m slab mixed layer ocean model with sea ice</td>
<td>15 year simulation</td>
<td>15 year simulation, rainforest covered grid cells replaced with grasses</td>
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<tr>
<td>Costa et al. (2007)</td>
<td>NCAR AGCM, 2.81° x 2.81°, 18 vertical levels</td>
<td>IBIS, 2.81° x 2.81°</td>
<td>Climatological seasonal cycle</td>
<td>20 year simulation</td>
<td>20 year simulation; 6 ensemble members; 25%, 50% and 75% deforestation with pastureland and soybeans as replacement vegetation</td>
</tr>
<tr>
<td>Davin and de Noblet-Ducoudré (2010)</td>
<td>IPSL LMDZ4, 3.75° x 2.5°, 19 vertical levels</td>
<td>ORCHIDEE</td>
<td>Océan Paralléllisé</td>
<td>110 year simulation, replace all vegetation with forest</td>
<td>110 year simulation, replace all vegetation with grass</td>
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<td>Dickinson and Henderson-Sellers (1988)</td>
<td>NCAR CCM, 4.5° x 7.5°</td>
<td>BATS</td>
<td>Not explicitly stated</td>
<td>13 months</td>
<td>13 months, replace evergreen broadleaf trees with impoverished grassland</td>
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<td>Hasler et al. (2009)</td>
<td>4° x 5°, 26 vertical levels</td>
<td>CLM3</td>
<td>Observed monthly SST</td>
<td>52 year simulation</td>
<td>52 year simulation, all tropical forests replaced with grass and shrubs</td>
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<td>Henderson-Sellers et al. (1993)</td>
<td>CCM1-OZ, 4.5° x 7.5°</td>
<td>BATS</td>
<td>Mixed layer</td>
<td>6 year simulation</td>
<td>6 year simulation, tropical forest replaced with scrub grassland</td>
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<td>Lejeune et al. (2014)</td>
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<td>32 year simulation</td>
<td>32 year simulation, control vegetation replaced by 80% cropland and 11% grassland</td>
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<td>McGuffie et al. (1995)</td>
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<td>BATS</td>
<td>Mixed layer</td>
<td>14 year simulation</td>
<td>6 year simulation, tropical forest replaced with shrub grassland</td>
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<td>Nobre et al. (1991)</td>
<td>NMC, 1.8° x 2.8°, 18 vertical levels</td>
<td>SiB</td>
<td>Prescribed</td>
<td>12.5 months</td>
<td>12.5 months, tropical forest replaced with degraded pasture</td>
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<td>Nobre et al. (2009)</td>
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<td>SSiB</td>
<td>GFDL MOM from 40°S to 40°N, monthly climatological SSTs for remainder</td>
<td>9 20-year ensembles, and one 90 year simulation</td>
<td>9 20-year ensembles, and 90 year simulation; all with forest replaced by savanna</td>
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<tr>
<td>Shukla et al. (1990)</td>
<td>1.8° x 2.8°, 18 vertical levels</td>
<td>SiB</td>
<td>Prescribed</td>
<td>1 year simulation</td>
<td>1 year simulation, tropical forest replaced with degraded pasture</td>
</tr>
<tr>
<td>Snyder et al. (2004)</td>
<td>CCM3, 3.75° x 3.75°, 18 vertical levels</td>
<td>IBIS</td>
<td>Climatologically prescribed SSTs</td>
<td>25 year simulation</td>
<td>25 year simulation, tropical forest replaced with bare soil</td>
</tr>
<tr>
<td>Snyder (2010)</td>
<td>CCM3m 2.8° x 2.8°, 18 vertical levels</td>
<td>IBIS</td>
<td>Climatologically prescribed SSTs</td>
<td>40 year simulation</td>
<td>40 year simulation, tropical forest replace with bare soil</td>
</tr>
<tr>
<td>Werth and Avisar (2002)</td>
<td>NASA GISS, 4° x 5° with 12 layers</td>
<td>8° x 10°</td>
<td>Climatological</td>
<td>Six 12 year simulations</td>
<td>Six 12 year simulations, forests replaced with grass and shrublands</td>
</tr>
</tbody>
</table>
or decrease the magnitude of changes due to deforestation and perhaps dampen global-scale
teleconnections (Davin and de Noblet-Ducoudré, 2010; Hasler et al., 2009; Pitman et al.,
2009). Nobre et al. (2009) adds that the role of ocean-atmosphere interactions on the
climate change issue related to Amazon deforestation should be of particular interest.

Another noticeable inconsistency among the simulations is the replacement vegetations
used. Admittedly, the difference in replacing the tropical forests with grass, pasture, sa-
vanna, shrubs or bare soil may be negligible, some inherent differences may arise. Only one
simulation, Costa et al. (2007), used a crop as replacement vegetation, perhaps the most
realistic replacement vegetation, which can have different impacts of the aforementioned
replacement vegetations.

1.3 Local Response to Deforestation

Eltahir and Bras (1993) suggests that Amazon deforestation will result in a higher surface
temperature, a reduction in evaporation and precipitation, and possibly significant changes
in the tropical circulation. Davin and de Noblet-Ducoudré (2010) add that deforestation in-
volves two different types of forcing mechanisms: a radiative forcing (surface albedo change)
and a nonradiative forcing (change in evapotranspiration efficiency and surface roughness).
Deforestation is expected to take place in the tropics where nonradiative effects are the
dominant forcing type. Amazon deforestation has a series of responses that are associated
with the removal of the tropical rainforest (Figure 1.2). As deforestation occurs, two direct
effects are the reduction of evapotranspiration and increase in surface albedo. The reduc-
tion in evapotranspiration means less latent heat flux to cool the surface; while an increase
in albedo will decrease net radiation and reduce the amount of energy available for latent
heat flux.

As the flux partitioning changes and more sensible heat flux is present, the planetary
boundary layer warms, dries and becomes deeper; which can affect the transfer of energy
from the land surface to the boundary layer, and even influence the transfer of energy
Removal of tropical forest vegetation

ET severely reduced

Increase in surface albedo
Reduction in net radiation

Reduced latent heat flux

Surface temperature increases
Sensible heat flux increases

Precipitation decreases

Less water recycled to the atmosphere, PBL dries and specific humidity decreases

PBL warms, dries and expands

Reduced deep tropical convection

Less moisture available for condensation; overall cloud cover fraction decreases

Increase in incoming solar radiation

Affects intensity and location of high-level tropical outflow

Affects transfers of momentum, heat, and moisture between land surface and PBL

Possible influence on extratropics through anomalous Rossby wave forcing

Figure 1.1: Schematic of processes resulting from deforestation (adapted from Snyder et al. (2004, Fig. 2)).
to the free atmosphere. Increased sensible heat flux also reduces the specific humidity in
the boundary layer and decreases the amount of water recycled to the atmosphere. Less
moisture recycling can be a negative feedback that can continue to reduce precipitation,
and thus reduce the cloud cover. A reduction in cloud cover can increase incoming solar
radiation, which can just further increase the sensible heat flux and cause more warming.
Accompanying less cloud cover is a reduction in convection, which can influence the tropical
atmospheric circulation and the global climate through Rossby wave forcing mechanism.
There is a clear set of processes that show how Amazon deforestation can have local impacts,
as well as possible impacts on the global scale.

1.3.1 Local Changes

A summary of the local changes associated with Amazon deforestation are given in Table
1.2. Not all of the studies reported the local changes, as some studies focused on different
effects of Amazon deforestation.

**Temperature**

The resulting change in annual surface temperature ranges from -1°C to +3°C. Several
studies note that the change in temperature is significant (Dickinson and Henderson-Sellers,
1988; Henderson-Sellers et al., 1993; Lejeune et al., 2014; McGuffie et al., 1995; Nobre et al.,
1991; Shukla et al., 1990; Snyder et al., 2004). Dickinson and Henderson-Sellers (1988) goes
on to add that while surface air temperature increases by 1-3°C, the soil-surface temperature
increase by 2-5°C.

McGuffie et al. (1995) notes that changes in surface temperature over the deforested
region are dipolar: an increase over the central and eastern Amazon and a decrease to the
southwest of the deforestation; the Student’s t-test shows that both changes are statisti-
cally significant at the 95% confidence level. These dipolar changes are exhibited by more
parameters that just temperature. In the southwest region, reversed changes from those
Table 1.2: A summary of results from previous modeling experiments aimed at determining the impacts of Amazon deforestation. All values are annual changes unless otherwise noted. “–” denotes results that are not explicitly stated.

<table>
<thead>
<tr>
<th>Study</th>
<th>Temperature</th>
<th>Precipitation</th>
<th>Evaporation</th>
<th>Runoff</th>
</tr>
</thead>
<tbody>
<tr>
<td>Costa and Foley (2000)</td>
<td>+1.4°</td>
<td>-0.7 mm/day</td>
<td>-0.6 mm/day</td>
<td>-0.1 mm/day</td>
</tr>
<tr>
<td>Davin and de Noblet-Ducoudrè (2010)</td>
<td>-1°</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Dickinson and Henderson-Sellers (1988)</td>
<td>+1.3°C</td>
<td>-20%</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Hasler et al. (2009)</td>
<td>–</td>
<td>-0.4 - 0.7 mm/day</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Henderson-Sellers et al. (1993)</td>
<td>+0.6°C</td>
<td>-1.6 mm/day</td>
<td>-0.6 mm/day</td>
<td>-0.9 mm/day</td>
</tr>
<tr>
<td>Lejeune et al. (2014)</td>
<td>+0.78°C</td>
<td>-0.22 mm/day</td>
<td>-0.4 mm/day</td>
<td>+0.18 mm/day</td>
</tr>
<tr>
<td>McGuffie et al. (1995)</td>
<td>+0.3°C</td>
<td>-437 mm/year</td>
<td>-231 mm/year</td>
<td>–</td>
</tr>
<tr>
<td>Nobre et al. (1991)</td>
<td>+2.5°C</td>
<td>-25%</td>
<td>-30%</td>
<td>-20%</td>
</tr>
<tr>
<td>Nobre et al. (2009)</td>
<td>–</td>
<td>-42.2%</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Ramos da Silva et al. (2008)</td>
<td>+0.3°C</td>
<td>-29.5 mm/month</td>
<td>-25 W/m²</td>
<td>–</td>
</tr>
<tr>
<td>Shukla et al. (1990)</td>
<td>+2.5°C</td>
<td>-26.1%</td>
<td>-30%</td>
<td>+18%</td>
</tr>
<tr>
<td>Snyder et al. (2004)</td>
<td>+1.5°C</td>
<td>-1.4 mm/day</td>
<td>-33.3%</td>
<td>–</td>
</tr>
<tr>
<td>Snyder (2010)</td>
<td>+1.2°C</td>
<td>-1.4 mm/day</td>
<td>-30.9 W/m²</td>
<td>–</td>
</tr>
<tr>
<td>Werth and Avissar (2002)</td>
<td>–</td>
<td>-296 mm/year</td>
<td>-214 mm/year</td>
<td>–</td>
</tr>
</tbody>
</table>
seen in the deforested region: evaporation is increased, total precipitation is increased, sensible heat flux decreased and the total runoff are increased (McGuffie et al., 1995). The climate changes occurring to the southwest of the deforested region indicate that circulation patterns have been modified by the deforestation; this suggestion will be further explored later.

Precipitation

A common feature of all the simulations in Table 1.2 is the decrease in precipitation, although they are of varying intensity. Decreases in annual precipitation are typically found to be significant (Costa and Foley, 2000; Hasler et al., 2009; Henderson-Sellers et al., 1993; Lejeune et al., 2014; McGuffie et al., 1995; Nobre et al., 1991, 2009; Shukla et al., 1990). Nobre et al. (2009) points out a difference in precipitation change in simulations coupled with the ocean; the coupled model produced a rainfall reduction that is nearly 60% larger than was obtained by the use of AGCM uncoupled from the ocean. As previously noted, the effect of different replacement vegetation may play a role. Costa et al. (2007) found that changes in precipitation for 25%, 50% and 75% deforestation, respectively, -6.2%, -11.6%, and -15.7% for the soybean land cover, which is significantly different than the +1.4%, -0.8%, and -3.9% for the pastureland cover.

The results presented in Table 1.2 are for changes in the annual precipitation; changes on the seasonal time scales are also of interest. Depending on the month, precipitation significantly decreases over 30-75% of the deforested area (Hasler et al., 2009), which shows varying impacts on monthly time scales that may be of interest. Hasler et al. (2009) also noted that 6%-13% of the deforested grid cells also experience an annual increased precipitation averaging 0.06-0.18 mm/day, approximately 5% of the annual mean in those cells, with the increase generally occurring during the wet season. Precipitation decreases in the deforested case for all months except December, but the difference is statistically significant at the 95% level in only half of the months (Costa and Foley, 2000). McGuffie et al. (1995) adds that total precipitation decreases all the year-round, particularly at the
times of the rainy season maxima and that most of the Amazon region shows decreased precipitation, but the region of statistical significance is small and located in the northern Amazon Basin. Changes in precipitation for the seasons of JFM and JAS show considerable differences, but mainly in the northern part of the Amazon (Costa and Foley, 2000). Both Costa and Foley (2000) and Lejeune et al. (2014) note the seasonality of the precipitation did not change significantly, with the rainy season and dry season occurring in the same periods.

Using an observed spatial pattern of deforestation, Bagley et al. (2014) adds that deforestation reduces the magnitude of precipitation during dry season and has the potential to enhance regional droughts.

The deforested region experiences a decrease of precipitation; the areas at the edge of that region and at higher elevations receive more rainfall (Ramos da Silva et al., 2008). In a sensitivity study, Ramos da Silva et al. (2008) finds that the basin-averaged rainfall progressively decreases with the increase in deforestation and that the basin experiences much stronger precipitation changes during El Niño events as deforestation increases.

Lejeune et al. (2014) adds that a dipolar of precipitation is present over the deforested region. It was found that there is a significant decrease in precipitation over the central Amazon, while a significant increase in precipitation to the east occurs as larger percentages of the Amazon region are deforested.

**Evaporation**

Evapotranspiration decrease after deforestation is a common result of the Amazon deforestation studies (Costa and Foley, 2000; Dickinson and Henderson-Sellers, 1988; Henderson-Sellers et al., 1993; McGuffie et al., 1995; Nobre et al., 1991; Shukla et al., 1990; Snyder et al., 2004). Costa and Foley (2000) adds that the differences in evapotranspiration are statistically significant in all months.

The decrease in transpiration, 53%, is much larger than the decrease in the total evapotranspiration, 16%; this indicates that increased evaporation from the surface partially
compensates for the drop in transpiration (Costa and Foley, 2000). Henderson-Sellers et al. (1993) notes that as the evaporation decreases, the near-surface specific humidity decreases. This result is of particular interest in the response of planetary boundary layer growth, which is discussed in latter sections.

**Other Impacts**

Along with decrease in annual precipitation, an increased length in the dry season was found in several simulations (Dickinson and Henderson-Sellers, 1988; Nobre et al., 1991; Shukla et al., 1990). Dickinson and Henderson-Sellers (1988) added that the length of the season with dry soil would increase. A significant finding, as soil moisture has an integral role in several aspects of the climate. Annual mean runoff is somewhat reduced by deforestation (Costa and Foley, 2000). Nobre et al. (1991) states runoff is reduced by 20%.

The change in leaf area index (LAI) has a role in the local hydrology. Dickinson and Henderson-Sellers (1988) notes that lower LAI causes a decrease in canopy interception and re-evaporation of precipitation. Snyder et al. (2004) adds that removal of the tropical forest vegetation influences the climate mostly through large reductions in water recycling.

Deforestation of small areas changes the spatial distribution of sensible and latent heat fluxes, resulting heterogeneity in the surface temperature and humidity fields plays a significant role in the initiation of convection and could favor mesoscale circulations, perhaps producing more rainfall (Eltahir and Bras, 1993). Wang et al. (2011) suggests that mesoscale deforestation (which is more realistic) tends to enhance convection, cloudiness, and precipitation over deforested areas due to mesoscale circulations triggered by land surface heterogeneity.

**Explanations of Local Changes**

The leading explanation for local changes is the change in albedo. Land surface albedo changes explain about 96% of the precipitation variance according to Costa et al. (2007). Eltahir and Bras (1993) notes that removal of the rainforest eliminates biomass that absorbs
most of the solar radiation incident on the surface. Thus less biomass results in less absorption and more reflection of solar radiation, hence a higher surface albedo. Costa and Foley (2000) adds that there are several reasons why pasture evapotranspiration is smaller than in the forest, the main reasons are the changes in albedo and roughness length, which have about the same importance. Pasture albedo is higher, causing less energy to be absorbed by the surface, while the reduced roughness length in the pasture implies reduced turbulent transfer between the atmosphere and the land surface.

The decrease in precipitation after a soybean extension is significantly higher when compared to the change after a pastureland extension Costa et al. (2007). The difference in results (soybean vs. pastureland) seems to be directly related to the change in land surface albedo and the water balance. As the reflected radiation increases due to increases in surface albedo and to cloud-radiative feedbacks; surface latent and sensible heat fluxes decrease due to reduced radiation absorbed by the surface, resulting in a cooling of the atmospheric column. Costa et al. (2007) notes that this induces a thermally driven circulation that results in subsidence, with subsequent reduction in convection, cloudiness and precipitation.

Plants mediate exchanges of radiation, water, heat, and momentum between the land and the atmosphere. The forest canopy is particularly important for the surface-energy budget in tropical regions (Dickinson and Henderson-Sellers, 1988). Therefore, changes in the vegetation cover can perturb these fluxes and thus impact the climate (Davin and de Noblet-Ducoudr´e, 2010). Replacement of the forest with a shorter vegetation cover reduces the roughness of the surface layer and causes reduction of the eddy transport of water vapor, heat and momentum near the surface; a shorter roughness length also results in less evaporation and higher surface temperatures (Eltahir and Bras, 1993). Santanello et al. (2011) notes that forested surfaces are considered strongly coupled while smooth grasslands are weakly coupled to the atmosphere. In the tropics the net impact of deforestation is a warming because evapotranspiration efficiency and surface roughness provide the dominant influence in these regions (Davin and de Noblet-Ducoudr´e, 2010). The reduction in evaporation is due largely to the shorter roughness length, while the reduction of precipitation is
due mainly to the smaller amount of energy available for convection following deforestation (Eltahir and Bras, 1993). However, Costa and Foley (2000) notes that differences in leaf area and rooting depth also play an important role.

In tropical forests, transpiration, the evaporation of water in the vascular system of plants with loss through leaf stomata, is overall the largest contributor to total evapotranspiration (Wang and Dickinson, 2012). Therefore, a smaller leaf area implies a reduced area active in transpiration and in canopy storage available for interception (Eltahir and Bras, 1993). Davin and de Noblet-Ducoudré (2010) points out that the sum of changes due to albedo; evapotranspiration and roughness are nearly equal to changes due to deforestation. This result suggests that there may be little nonlinearity involved in the effects of Amazon deforestation.

1.3.2 Effect on the Planetary Boundary Layer

The effects of deforestation on the planetary boundary layer (PBL) are of interest. PBL growth provides a mechanism for the land surface to interact with the free atmosphere. Soil moisture and vegetation feedback may also play a major role in regions like the Amazon where climate is sensitive to land surface conditions (Wang et al., 2011). Studies have shown that LUC has the potential to significantly alter local and regional surface temperature through altered surface fluxes, particularly in tropical regions (e.g., Bagley et al. 2012. These altered surface conditions offer a pathway to perturb the free atmosphere.

**PBL – L-A Interaction Theory**

Moisture and heat fluxes from the land surface to the atmosphere form a critical nexus between surface hydrology and atmospheric processes, particularly the processes relevant to precipitation (e.g., Findell et al. 2011. Land-atmosphere interactions play a critical role in determining the diurnal evolution of the PBL, land surface temperature and soil moisture states (Santanello et al., 2009). Findell and Eltahir (2003) states that the lowest 300 mb of a sounding are critical in processes related to feedbacks from the land surface. Within
this critical convective triggering potential region, the temperature lapse rate is important for determining the ease with which entrainment, and therefore boundary layer growth, can occur. The convective triggering potential provides critical information about the boundary layer response to surface fluxes in a given atmospheric setting. In the energy cycle, both latent and sensible heat fluxes at the surface may be gateways between soil moisture forcing and the atmospheric response (Dirmeyer et al., 2012). Dirmeyer et al. (2012) adds that while surface latent heat flux moistens the PBL, it is the sensible heat flux that contributes to its growth, and that only after the PBL top has penetrated the lifting condensation level (LCL), can convective clouds and precipitation form. Kim and Entekhabi (1998b) further details that during the day the PBL grows mainly in response to the surface sensible heat flux and when the sensible heat flux vanishes, the turbulence dissipates and the boundary layer collapses.

Seneviratne et al. (2010) notes that through soil moisture’s impact on the partitioning of the incoming energy in the latent and sensible heat fluxes, soil moisture has several additional impacts on climate processes, in particular on air temperature, boundary-layer stability and in some instances on precipitation. The effect of soil moisture anomalies on the lower atmosphere results in a direct correlation with near-surface humidity, but an inverse relationship with temperature, and thus the growth of the boundary layer (Dirmeyer, 2011). Decrease of latent heat flux over a large area leads to a drying and warming of the PBL (Kim and Entekhabi, 1998b). The impact of soil moisture on clouds and precipitation is dependent on several sensitivities. Santanello et al. (2011) provides a pathway for these effects: 1) the surface fluxes respond to soil moisture, 2) PBL evolution is dependent on surface fluxes, 3) entrainment fluxes at the PBL-top are dependent on PBL evolution, and thus shows the collective feedback of the atmosphere (via the PBL) due to surface fluxes. In brief, the processes that contribute to land-precipitation coupling are the relationship between soil moisture anomalies and the resulting evapotranspiration anomalies, then how these evapotranspiration anomalies affect precipitation anomalies, which provide the relationship between precipitation and the land surface state.
PBL Response to Deforestation

With increasing spatial scales, the signature of the land surface on the conditions in the PBL becomes clearly significant (Kim and Entekhabi, 1998a). Thus large scale LUC can play a major role in the PBL. Deforestation of large areas reduces evaporation and results in a drier boundary layer, which affects precipitation significantly (Eltahir and Bras, 1993). This can be explained due to the fact there is less water available to form clouds and there is a reduction in energy available for convection. The decrease in precipitation is coherent with the larger albedo increase, as dry soils are more reflective (Snyder et al., 2004). There is a positive feedback mechanism-taking place in which a reduction in net radiation at the surface due to the higher albedo dampens the potential development of convective precipitation due to less available energy. Bonan (2008b) adds that flux tower measurements in the Brazilian Amazon confirm that forests have lower albedo compared with pasture, greater net radiation, and greater evapotranspiration; producing a shallower, cooler, and moister boundary layer. Costa et al. (2007) also notes that there are two mechanisms that reduce precipitation: 1) the suppression of precipitation due to the cooling of the atmospheric column, and 2) the reduction of moisture from local evapotranspiration resulting in a drying of the column.

Mechanism for Deforestation’s Effect on the PBL

Less removal of energy and water from the land surface by latent heat results in increased temperatures and increased runoff (Dickinson and Henderson-Sellers, 1988). Due to the fact that evaporation cools the surface, the reduction in evaporation results in an increase in surface temperature (Eltahir and Bras, 1993). Even though removal of the tropical forest vegetation causes a reduction in the net radiation absorbed at the surface, the surface temperature increases as the loss of net radiation is more than compensated for by the increased energy at the surface due to reduced latent cooling (Snyder et al., 2004).

In forested areas, there is sensitivity of the evaporative fraction (EF) to soil moisture variations when soils are dry but little sensitivity in moderate to wet soils (Dirmeyer et al.,
Dirmeyer et al. (2000) states that the forest canopy can sustain a nearly constant EF across a wide range of soil wetness; while shallow-rooted grasses and short vegetation are not as capable as forests of accessing water in dry conditions. This implies that replacement vegetation in Amazonia would produce a more variable evaporative fraction. Due to root uptake and transpiration of water, the forest allows for much higher values of evaporative fraction in dry soil conditions than could be expected by soil evaporation alone (Dirmeyer et al., 2000). Deforestation has a direct effect on the top soil layer due to greater exposure to different environmental factors, such as increased radiation reaching the top soil layer and increased wind speed enhancing mixing in the boundary layer. The top soil layer controls surface runoff and stores the moisture necessary for transpiration; therefore changes to the top soil layer can affect multiple processes (Eltahir and Bras, 1993). Shukla et al. (1990) added that the reduced storage capacity for soil moisture in the deforestation case reduces the transpiration rate.

Increasing the specific humidity enhances the greenhouse effect and thus the equilibrium temperature. As temperatures increase the latent heat flux becomes increasingly more efficient in removing energy. Kim and Entekhabi (1998a) explains this due to the fact that Clausius-Clapeyron relation predicts the specific humidity deficit sharply increases with temperature and the latent heat flux increases with increasing specific humidity.

Increasing PBL depths will tend to weaken the surface energy contribution to the PBL by distributing it over a greater mass of air (Dirmeyer et al., 2012). Eltahir and Bras (1993) states that the change in runoff from the Amazon region can be estimated from the difference between the anomaly in atmospheric moisture convergence in the PBL and the corresponding anomaly in atmospheric moisture divergence in the atmosphere above the PBL. Henderson-Sellers et al. (1993) finds the moisture divergence in the deforested regime to be only 55% of the control case. Month-by-month analysis indicates that the greatest decrease in convergence occurs in the period November to February, the rainy season (Henderson-Sellers et al., 1993). Nobre et al. (1991) and Shukla et al. (1990) also indicate that the dynamical convergence of moisture flux decreased as a result of deforestation.
McGuffie et al. (1995) further details that decreased dynamic convergence of moisture over the Amazon after deforestation indicates a modification in the atmospheric circulation over the Amazon following deforestation.

Precipitation recycling drops by nearly 5% from the forested to the deforested simulations (Costa and Foley, 2000). Precipitation recycling is precipitation over a defined region that originated as evaporation from the same region (Dirmeyer et al., 2009). A high recycling ratio suggests strong local climate feedbacks. Costa and Foley (2000) also states that an Amazonia covered by pasture is more dependent on external sources of water vapor.

1.3.3 Local and Regional Circulation

As previously noted, the temperature and precipitation anomalies induce a circulation change in the Amazon region. Eltahir and Bras (1993) states that the increased temperature causes convergence towards the Amazon region, while the decrease in precipitation induces a divergence of similar strength. These two anomalies have opposite effects resulting in the circulation anomaly being smaller in magnitude than the larger of the two components. It is possible to have significant anomalies in surface temperature and precipitation with only a negligible effect on the circulation. One anticipated outcome of tropical deforestation is a reduction in the vertical ascent over the deforested region caused by the increase in surface albedo as Henderson-Sellers et al. (1993) explains is due to the net loss of energy to the column and also by the decrease in net surface radiation caused by the smaller turbulent exchanges. A shallow response of the tropical atmosphere to LUC effects is limited to diabatic heating in the tropics (Jonko et al., 2010).

LUC in the Amazon region causes decreased vertical motion in the atmosphere above the deforested area. Costa and Foley (2000) points out that there are some small regions with increased vertical motion. The ascent over the Amazon is considerably diminished after deforestation (Henderson-Sellers et al., 1993; McGuffie et al., 1995). McGuffie et al. (1995) adds that at the same time, south of the deforested region, decreased descending motion occurs. Davidson et al. (2012) makes note that heterogeneous deforestation at large
scales leads to more complex circulation changes, with suppressed rainfall over core clearings and unchanged or increased rainfall over large remnant forest patches.

It can be expected that the changes from deforestation would also be important for the regional-scale atmospheric circulation due to the major changes in surface fluxes of sensible and latent heat. Snyder et al. (2004) states that increased precipitation over the northern Amazon basin and Venezuela, may be due to the advection of moisture from other regions or changes in the regional circulation or boundary layer thermodynamics, which alter the conditions favorable for development of convective precipitation. Deforested regions not only see a reduction in precipitation, but also a redistribution of the regions of convection as well as a change in the intensity of deep versus shallow convection. These changes in convective activity can have an impact on the climate outside of the tropics.

1.4 Remote Response to Deforestation

The climatic influence of the tropical rainforests may extend to the extratropics through atmospheric teleconnections (Bonan, 2008b). The tropics are an important source of energy for the extratropical atmosphere; the changes in the thermodynamics and dynamics of the tropical atmosphere can be felt around the globe (Snyder, 2010). Jonko et al. (2010) suggests that nonlinearities could arise from the interactions of tropical and extratropical anomalies or from the modification of intraseasonal tropical variability imposed by LUC.

A mechanism for propagation to middle and high latitudes of disturbances arising from tropical deforestation is proposed by Zhang et al. (1996), based on Rossby wave propagation, providing a dynamical mechanism for possibly explaining how tropical deforestation can have global climate effects.

1.4.1 Case Study of Tropical Heating

Gill (1980) showed solutions for diabatically induced tropical circulations (Figure 1.2). Although Gill centered heating north of the equator, inferences about heating in the Amazon region can be made. The tropical heating caused poleward motion in the lower layer and
equatorward motion aloft. The changes in meridional flow produce a low-level easterly flow to the east through the propagation of Kelvin waves into the heating region and produces a low-level westerly flow to the west due to planetary waves. The influenced region to the west is about 1/3 the size of the region influenced to the east. The upper-level flow that is towards the equator is west of the heating region. Also present is cyclonic flow obtained around lows formed on the western margins of the heating region. Figure 1.2 can be flipped vertically to give an idea of how heating the Amazon region can affect the circulation, and is very similar to the resulting circulation change presented by Eltahir and Bras (1993).

Evidence of a wave train forced by the tropical changes has been seen in numerous studies. A Rossby wave response was found over areas of large-scale LUC (Jonko et al., 2010). Snyder et al. (2004) adds that removal of tropical forest leads to a change in the intensity and location of the high-level tropical outflow from deep tropical convection; which can influence the extratropics through anomalous Rossby wave forcing. In a later study, Snyder (2010) confirmed that removal of the tropical forests excites a Rossby wave train emanating northeastward away from the South American continent. Also of note is an eastward propagation of Kelvin waves in the tropics due to Amazon deforestation (McGuffie et al., 1995). Werth and Avissar (2002) hypothesizes that teleconnections between the Amazon and the remote areas are due to the fact that the large scale LUC in the Amazon serves as a wave source. The geopotential anomalies tend to reinforce each other, producing
a strong wave pattern over Asia and North America, and are suggestive of a deforestation effect (Hasler et al., 2009).

1.4.2 Large-Scale Circulation Changes

By mass continuity, the decrease in ascent associated with the deforested regions will cause changes beyond the areas of disturbance. There must be either a decrease in descent or an increase in ascent elsewhere. Therefore, it seems more than reasonable that deforestation can lead to a disturbance in some aspects of the general circulation, most notable the Hadley and Walker circulations.

Jonko et al. (2010) notes that LUC based on the IPCC SRES A2 scenario have a discernible impact on the large-scale tropical circulation. Results indicate that a small fractional change in the evapotranspiration causes a large change in the net energy supply and results in a large modification to the meridional structure of the Hadley circulation (Zhang et al., 1996). As previously noted, deforestation has a dramatic effect on evapotranspiration, and thus can have a considerable effect on the Hadley circulation. A reduction on Amazon deep convection acts to reduce the trade winds, through a reduction of the mean meridional atmospheric circulation and its effect on the subtropical high pressure systems (Nobre et al., 2009).

**Hadley and Walker Circulations**

Significant modifications to both the Hadley and Walker circulations are found after deforestation (Zhang et al., 1996). Zhang et al. (1996) suggests that these modifications can cause changes in regions distant from the deforestation. After deforestation, a decrease in the circulation strength of the Hadley cell over South and Central America was found (Henderson-Sellers et al., 1993). Over the Amazon region, this effect can be interpreted as a weakening of the ascending branch of both the local Hadley and Walker and Hadley circulations (Jonko et al., 2010).
Henderson-Sellers et al. (1993) notes that after deforestation, both the ascent in the ITCZ and the descent in both subtropical regions is weakened, but these effects do not extend outside 40°N and 40°S. The major cause of disturbances outside the deforested region seems to be caused by the reduced ascent in the ITCZ. Snyder (2010) further suggests that after deforestation, changes in European storm-track activity cause an intensification and northward shift in the Ferrell cell that leads to anomalous adiabatic warming over a broad region of Eurasia.

The impact on the general circulation of the atmosphere can be detected in the cells of the Walker circulation in both January and July and results in diminished ascent in the ITCZ over both deforested areas (Henderson-Sellers et al., 1993). The impact of deforestation on the Walker circulation is essentially to remove the Atlantic and east Pacific cells in July so that descent predominates from approximately 140°W to 100°E.

**Transports**

Along with the circulation change associated with Amazon deforestation, modifications to atmospheric transports can be expected. Zhang et al. (1996) states that deforestation can affect meridional energy transport in the middle and high latitudes. Poleward transports of sensible heat energy were reduced in both hemispheres, with large changes in the high latitudes of the Southern hemisphere. Hasler et al. (2009) also notes that the zonal and vertical mean potential energy fluxes further emphasize the changes in energy transports due to deforestation from the tropics into the northern midlatitudes.

Deforestation not only has impacts on the regional precipitation, but can also influence the regional hydrology through modifying the energy associated with the ITCZ (McGuffie et al., 1995). This is of particular interest due to the large role that evaporation from the Amazon rainforest has in supplying water to the neighboring regions.

Approximately eight trillion tons of water evaporates from the Amazon rainforest each year, which can have important influences on global atmospheric circulation (Nepstad et al., 2008). Dirmeyer et al. (2009) states that Brazil, which has a large area, humid climate,
Table 1.3: Percentage of water vapor supplied by Brazil for South American countries (from Dirmeyer et al. 2009).

<table>
<thead>
<tr>
<th>Country</th>
<th>Percentage of Water Vapor Supplied by Brazil</th>
</tr>
</thead>
<tbody>
<tr>
<td>Argentina</td>
<td>10.8%</td>
</tr>
<tr>
<td>Bolivia</td>
<td>34.5%</td>
</tr>
<tr>
<td>Colombia</td>
<td>19.8%</td>
</tr>
<tr>
<td>Ecuador</td>
<td>29.3%</td>
</tr>
<tr>
<td>French Guiana</td>
<td>2.8%</td>
</tr>
<tr>
<td>Guyana</td>
<td>4.6%</td>
</tr>
<tr>
<td>Paraguay</td>
<td>40.0%</td>
</tr>
<tr>
<td>Peru</td>
<td>34.4%</td>
</tr>
<tr>
<td>Suriname</td>
<td>3.9%</td>
</tr>
<tr>
<td>Uruguay</td>
<td>25.0%</td>
</tr>
<tr>
<td>Venezuela</td>
<td>7.0%</td>
</tr>
</tbody>
</table>

and lies under the western edge of the South Atlantic subtropical anticyclone between the Atlantic Ocean and much of the rest of South America, is a major source of water vapor to nearly every other nation on the continent. A summary is shown in Table 1.3.

It can clearly be seen that Amazon deforestation can have an impact on the water supplies of other South American countries; expanding this climate issue into a social issue. (Dirmeyer et al., 2009) points out that several South American nations show the largest dependence on another nation as their water vapor source; Bolivia (5th largest dependence in the world), Ecuador (10th), Paraguay (2nd), and Peru (6th).

1.4.3 Comparison to El Niño

It is interesting to compare El Niño, the classical ocean-atmosphere interaction phenomenon with the Amazon deforestation problem. Eltahir and Bras (1993) suggests that these two problems are similar in the sense that in both cases, the atmosphere responds to anomalies in the lower boundary conditions in the tropics, specifically in evaporation and surface temperature. Deforestation is similar to El Niño events in the sense that it provides a pathway for the dispersion of the tropical disturbances to high latitudes (Zhang et al.,
Nobre et al. (2009) found significant remote atmospheric responses to Amazon deforestation scenarios in coupled simulations, which revealed global ocean and atmosphere circulation changes due to enhanced ocean-atmosphere variability over the Pacific Ocean. The responses are interpreted as enhanced El Niño activity over the Pacific and a positive feedback contributing to the extra rainfall reduction over the Amazon in the coupled simulations.

1.4.4 Teleconnections

By modifying the atmospheric circulation, impacts also extend beyond the deforested region (McGuffie et al., 1995). The anomalous forcing of Rossby waves can have a direct impact on the Northern Hemisphere general circulation and climate through atmospheric teleconnections (Snyder, 2010). Werth and Avissar (2002) found several remote areas where this local change to the surface had a noticeable response.

Disturbances in South America extend beyond the region of land-surface change causing temperature reductions and precipitation increases to the south of the deforested Amazon (Henderson-Sellers et al., 1993). Snyder (2010) states that the extratropical response was found to be strongest across Eurasia in Northern Hemisphere winter, but the temperature anomalies from deforestation exist throughout the year. Outside the deforested regions, statistically significant changes in precipitation can be found at a few locations in North America, Africa, and in the western tropical Atlantic and Pacific (Hasler et al., 2009). Hasler et al. (2009) also suggests that the significant changes confirmed the existence of a teleconnection mechanism.

Model results indicate that Amazon deforestation can influence the extratropical general circulation in other ways. Results from Snyder (2010) show that with deforestation there is a deepening in Northern Hemisphere polar sea level pressure with lower geopotential heights, and a thickening of the troposphere with an increase in geopotential heights, south of 60°N is present.
1.4.5 Ocean Changes

One aspect of deforestation that has not been researched heavily is changes to the oceans. A modification of the general circulation can have impacts on the ocean surface, and over a long period of time, the ocean sub-surface. Davin and de Noblet-Ducoudré (2010) suggests that the ocean surface responds to deforestation by a cooling and even the temperature change over land is strongly affected by the ocean coupling. Davin and de Noblet-Ducoudré (2010) also acknowledges that by not taking into account the coupling with the ocean, the results would have concluded that the net effect of deforestation, averaged over all land areas, is a warming; by accounting for the ocean coupling, this net effect is of opposite sign. The main parameter involved in the coupling with the ocean is surface albedo, this is because the change in albedo modifies temperature and humidity in the whole troposphere, thus enabling the initial land perturbation to be transferred to the ocean.

Nobre et al. (2009) state that total deforestation shows weaker trade winds in almost all latitudes from 20°S to 20°N, and as a result, the Pacific ocean surface flow is weaker at the equator in the northern branch of the South Equatorial Current (nSEC). The weakening of the nSEC causes an upward displacement of the equatorial undercurrent (EUC) core, and is seen by an area of intensification above the thermocline and weakening below it. Due to a weakening of the Hadley circulation, there is less upwelling along the equator and less downwelling in the North Hemisphere. The eastern Pacific is nearly 1°C warmer in the deforestation simulation than the control simulation and the western Pacific is cooler by the same amount, Nobre et al. (2009) suggests this is shifting the system closer to warm El Niño conditions. The warmer eastern Pacific displaces convection typically over the western Pacific to the central and eastern Pacific, thus producing increased rainfall close to the South American continent, a similar result of warm El Niño events. With the displacement of the warm-pool atmospheric convection to the eastern Pacific, enhanced large-scale subsidence over the Amazon rainforest can be seen, and thus further reducing rainfall over the Amazon region.
1.5 Inspiration for Further Research

From the conclusions presented, the problem of LUC, and more specifically Amazon deforestation, has several areas that need significant research. Costa et al. (2007) notes that further study of the role of soybeans, and croplands in general, on the physical climate system of Amazonia need further research. Future research needs to include global as well as regional climate experiments that determine Amazon deforestation influences on climate. Together with a diagnosis of the planetary boundary layers ability to convert surface flux anomalies into weather and climate anomalies, a thorough diagnosis of potential land surface impacts on climate variability may be possible (Dirmeyer, 2011).

1.5.1 The Need for Long Fully-Coupled Simulations

To understand the effects of deforestation in the fully-coupled Earth system, it is necessary to perform long simulations with an interacting ocean. Long coupled simulations will allow for the effects of El Niño with Amazon deforestation (Werth and Avissar, 2002). Nobre et al. (2009) suggests that a detailed investigation of the intraseasonal, annual, and interannual variability is necessary to understand the effect of Amazon deforestation on the Atlantic and Indian Oceans and need to be the subject of further study. In a green planet versus desert planet experiment, Kleidon et al. (2000) states that the differences over the ocean could lead to subsequent changes in the oceans, sea surface temperatures and the meridional heat transport. Kleidon et al. (2000) suggests it would be interesting to find out in which regions oceanic feedbacks would intensify or reduce the response of LUC.

Hasler et al. (2009) notes that reproducing the results associated with Amazon deforestation with transient observed SSTs, or with coupled ocean-atmosphere models is an essential next step to provide additional insights on the spatial and temporal variability of the teleconnections. Further work must examine how LUC affects the local and regional circulation as well as how certain biomes influence the climate through global teleconnections (Snyder et al., 2004). Large-scale LUC may alter precipitation hundreds to thousands
of kilometers from the region of vegetation change (Spracklen et al., 2012). Nobre et al. (2009) notes that another aspect emerging from this coupled modeling research is the lack of a linear relationship between rainfall departures over the eastern Pacific and the extent of deforested area over the Amazon.

1.5.2 Linearity of Deforestation

Costa et al. (2007) investigated the local linearity of deforestation by replacing 25%, 50% and 75% of the Amazon with soybeans and pastureland. The responses for varying the degree of deforestation were different for each case producing respectively a 6.2%, 11.6% and 15.7% decrease in precipitation. Meanwhile, pastureland as a replacement vegetation produced a +1.4%, -0.8% and -3.9% change in precipitation. A common assumption is that the degree of response to land surface changes like this will vary linearly with the degree of change. These results suggest non-linear responses are possible. While precipitation response to increasing levels of soybean coverage (Costa et al., 2007) was relatively linear, the response to varying levels of pastureland was not.

Lejeune et al. (2014) simulated 33%, 66%, and 100% deforestation using predicted deforestation distributions. Lejeune et al. (2014) finds through the analysis of areal averages of temperature and precipitation that partial deforestation behaves relatively linear. However, it is suggested that this behavior may be model-dependent and that feedback may be suppressed due to model set-up.

The results of Costa et al. (2007) and Lejeune et al. (2014) raise several key questions about partial deforestation. Most notably, does 50% deforestation produce significantly more or less than 50% of a total deforestation signal. If a non-linear response would be indicated, this would suggest feedback processes in the climate system were amplifying or dampening the response.
1.5.3 Societal Effects

The effects of Amazon deforestation extend beyond the physical climate and can have socio-economic impacts in the region. Spracklen et al. (2012) states that reduced regional precipitation will have consequences for rainfall-reliant industries both within and outside the Amazon basin, including agriculture and hydroelectric power generation, which contribute substantially to South American economies.

Changes in climate induced LUC can have profound impacts on food production (Bagley et al., 2012). Much of the moisture that contributes to soybean crops comes from the north and northwest, with a substantial fraction of the evaporative source being located over the Brazilian Amazon. In a sensitivity study of LUC and food production, Bagley et al. (2012) suggests the largest impact on moisture availability was found to be South American soybeans. Bagley et al. (2012) adds that a large contributor to this was that the moisture source for this region tends to be largely concentrated over heavily forested biomes, where LUC has been found to most strongly influence growing season evapotranspiration. For South American soybeans where approximately 60% of the evaporative source for regions soybeans was recycled within the region, and a significant portion of that was from tropical rainforests.

1.6 Overview of this Study

Chapter 2 describes the model of choice, description of model simulations and the modifications to the model, including the development of tropical crops. The local response of temperature, precipitation and surface fluxes to Amazon deforestation are discussed in Chapter 3. Chapter 3 further discusses the modifications to land-atmosphere coupling and the role that irrigation has. Chapter 4 describes large-scale circulation changes and highlight notable remote responses. The spatial patterns of non-linearity to partial deforestation is discussed in Chapter 5. Chapter 5 provides a method to quantify the degree of non-linearity over a spatial domain and shows that Amazon deforestation has a non-linear impact on the
climate system. Conclusions and further research is discussed in Chapter 6.
Chapter 2: Methods

This section contains a description of the climate model used in this study to obtain the results in the following chapters. The model of choice is the Community Earth System Model (CESM) version 1.2.0 developed at the National Center for Atmospheric Research (NCAR). CESM is a coupled model system for simulating the Earth's climate and is composed of separate models simulating the Earth's atmosphere, ocean, land, land-ice and sea-ice (Vertenstein et al., 2013). In addition to a description of CESM, this section also details additional model development that has been necessary to execute the proposed experiments, and a description of the model simulations performed.

2.1 Model overview of CESM 1.2.0

The CESM project is a cooperative effort among many U.S. climate researchers. CESM development is primarily supported by the National Science Foundation (NSF) and based at the NCAR in Boulder, Colorado. CESM consists of the following geophysical models: the Community Atmosphere Model (CAM), the Community Land Model (CLM), the Parallel Ocean Program (POP), the Community Ice CodE (CICE), the River Transport Model (RTM) and the Community Ice Sheet Model (CISM) (Vertenstein et al., 2013). CESM can be configured in multiple ways, supports numerous resolutions and component configurations. Each model component can either be active, data, stub, or dead, which allows for a variety of combinations. "Active" means that model component is fully prognostic. "Data" means that model component is supplying prescribed input data (e.g., climatological or observed forcing). "Stub" components are used to satisfy technical requirements when the component is not necessary for the simulation (e.g., using climatological atmospheric forcing to drive the land surface would not require an ocean model, it would be run as a stub), and
"dead" components are used only for system testing. Additionally, each model component has options to configure specific model physics and parameterizations. All active model components are described in the following sub-sections.

2.1.1 Community Atmosphere Model version 4 (CAM4)

CAM4 is a global atmospheric general circulation model and was released as part of the Community Climate System Model version 4 (CCSM4) (Neale et al., 2010). It has a dynamical core component using the finite volume (FV) numerical scheme, which is used as the default dynamical core due to its superior transport properties. The FV dynamics and physics are time split, meaning that prognostic variables are updated sequentially by the dynamics and then by the physics. The time integration within the FV dynamics is fully explicit and uses sub-cycling within the two-dimensional Lagrangian dynamics to stabilize the fastest wave. The horizontal discretization in CAM4 is based on a conservative flux-form semi-Lagrangian approach with vertical discretization described as Lagrangian with a conservative remapping.

CAM4 physics consists of a sequence of components for moist precipitation processes, clouds and radiation, a surface model and turbulent mixing (see Neale et al. (2010) for complete information). Each of these components is divided into subcomponents. Moist precipitation processes include a dry adiabatic adjustment applied in the stratosphere, moist penetrative convection, shallow convection and large-scale stable condensation. The cloud and radiation component calculates the cloud parameterization based on clouds and convection in the atmospheric column and then the radiation parameterization based on cloud distribution. The surface model transmits the surface fluxes from the land, ocean and sea ice models, or calculates them from specified surface conditions. The turbulent mixing component uses the provided fluxes for lower boundary conditions, which includes the representation of the planetary boundary layer (PBL), vertical diffusion and gravity wave drag.

The exchanges of heat, moisture and momentum between the atmosphere and surface
(land, ocean and ice surfaces) are treated with a bulk exchange formulation. The turbulence parameterizations in CAM4 include diffusivities for the free atmosphere and an explicit, non-local atmospheric boundary layer (ABL) parameterization. The ABL parameterization includes a determination of the boundary layer depth. In CAM4, the free atmosphere diffusivities are determined at all levels and then the ABL scheme determines the ABL height and diffusivities, replacing the free atmosphere values for all levels within the ABL.

The calculation of convective available potential energy (CAPE) uses an entraining plume to provide the in-cloud temperature and humidity profiles used to determine buoyancy and various cloud properties (Neale et al., 2010). This replaces the method used in previous versions of CAM and increases convective sensitivity to tropospheric moisture while reducing the amplitude of the diurnal cycle of precipitation over land surfaces. The convective scheme is based on a plume ensemble approach; it is assumed that an ensemble of convective scale updrafts exists when the atmosphere is conditionally unstable in the lower troposphere. The updraft ensemble plumes are sufficiently buoyant to penetrate the unstable layer. The moist convection occurs when there is enough CAPE for parcel ascent from the sub-cloud layer.

2.1.2 The Community Land Model version 4.5 (CLM4.5)

The Community Land Model (CLM) is the product of a collaborative project between scientists in the Terrestrial Sciences Section of the Climate and Global Dynamics Division of NCAR, the CESM Land Model Working Group and the CESM Biogeochemistry Model Working Group, as well as scientists at universities and government laboratories (Vertenstein et al., 2013). CLM4.5 incorporates recent scientific advances in the understanding and representation of land surface process, provides expanded model capabilities, and improves surface and atmospheric forcing datasets (Oleson et al., 2013). CLM4.5 is a model developers release of CLM and provides improvements prior to the public release of CLM version 5.

Land surface heterogeneity in CLM4.5 is accomplished with a nested sub-grid hierarchy
(see Figure 2.1) in which grid cells are comprised of multiple land units, soil columns, snow columns, and plant functional types (PFTs) (Oleson et al., 2013). The PFT level, which also includes bare ground, is intended to capture the biogeophysical and biogeochemical differences between broad categories of plants in terms of their functional characteristics. Fluxes to and from the surface are defined at the PFT level, as well as the vegetation state variables, such as vegetation temperature and canopy water storage. A list of default CLM4.5 PFTs (additional PFTs are added if the crop model and/or irrigation model is active) include:

- Needleleaf evergreen tree - temperate
- Needleleaf evergreen tree - boreal
- Needleleaf deciduous tree - boreal
- Broadleaf evergreen tree - tropical
- Broadleaf evergreen tree - temperate
- Broadleaf deciduous tree - tropical
- Broadleaf deciduous tree - temperate
- Broadleaf deciduous tree - boreal
- Broadleaf evergreen shrub - temperate
- Broadleaf deciduous shrub - temperate
- Broadleaf deciduous shrub - boreal
- C3 Arctic Grass
- C3 Grass
- C4 Grass
- C3 Unmanaged Rainfed Crop
The different PFTs are characterized by parameters that differ in leaf and stem optical properties to determine the reflection, transmittance and absorption of solar radiation (Oleson et al., 2013). Each PFT has a specific root distribution to allow for root uptake of water from the soil. Different PFTs have aerodynamic parameters that determine heat, moisture and momentum transfers. Photosynthetic parameters that determine stomatal resistance, photosynthesis and transpiration are used for each PFT. These parameterizations are used to make unique PFTs to best represent their respective behaviors.

Biogeophysical processes included in CLM are solar and longwave radiation interactions with vegetation, canopy and soil; momentum and turbulent fluxes from the canopy and soil; heat transfer in soil and snow; hydrology of the canopy, soil and snow; and stomatal physiology and photosynthesis (Lawrence et al., 2011). Biogeophysical and biogeochemical processes are simulated for each subgrid land unit, column, and PFT (see Figure 2.1) independently (Oleson et al., 2013). The same atmospheric forcing is used to force all subgrid units within a grid cell with each subgrid unit maintaining its own prognostic
variables. A weighted average by fractional area of each subgrid is calculated to obtain the surface variables and fluxes required by the atmosphere.

The zonal and meridional momentum fluxes, sensible heat flux, and water vapor flux are derived from Monin-Obukhov similarity theory developed for the surface layer (Oleson et al., 2013). The sensible heat and water vapor fluxes are partitioned into vegetation and ground fluxes that depend on vegetation and ground temperatures. Surface fluxes are also dependent on surface layer specific humidity. The air within the canopy is assumed to have negligible capacity to store heat; therefore the sensible heat flux between the surface and the atmosphere at a height within the canopy must be balanced by the sum of the sensible heat from the vegetation and the ground. Similarly, the air within the canopy is assumed to have negligible capacity to store water vapor; analogously, the water vapor flux between the surface and the atmosphere at a height within the canopy must be balanced by the sum of the water vapor flux from the vegetation and the ground.

The model parameterizes many aspects of canopy hydrology, the snow layer, surface and sub-surface hydrology; such as interception of precipitation, throughfall, canopy drip, snow accumulation, snow melt, water transfer between snow layers, infiltration, evaporation, surface runoff, sub-surface drainage, redistribution within the soil column, and groundwater discharge and recharge (Oleson et al., 2013). Precipitation is either intercepted by the canopy, falls directly to the surface (called throughfall), or drips off the vegetation (called canopy drip). The rate of interception by vegetation is dependent upon leaf and stem areas. The moisture input at the grid cell surface is the sum of liquid precipitation reaching the ground surface and melt water from snow. The moisture flux is then partitioned between surface runoff, surface water storage, and infiltration into the soil. The surface water storage and outflow are functions of fine spatial scale elevation variations called micro-topography, the spatial scale of the micro-topography is small relative to that of the grid cell and it is assumed that the inundated areas are distributed randomly within the grid cell. Soil water is predicted from a multi-layer model, in which the vertical soil moisture transport is governed by infiltration, surface and sub-surface runoff, gradient diffusion, gravity, canopy
transpiration through root extraction, and interactions with groundwater.

Leaf stomatal resistance, which is needed for transpiration, is coupled to leaf photosynthesis (Oleson et al., 2013). Leaf stomatal resistance is calculated from the Ball-Berry conductance model (Ball and Berry, 1982), which relates stomatal conductance to net leaf photosynthesis scaled by the relative humidity at the leaf surface and the CO2 concentration at the leaf surface. Soil water influences stomatal conductance directly with a soil water stress function that ranges from one when the soil is wet to near zero when the soil is dry and depends on the soil water potential of each soil layer, the root distribution of the PFT, and a plant-dependent response to soil water stress.

**CN mode**

CLM4.5 includes a fully-prognostic treatment of the terrestrial carbon and nitrogen cycles (Oleson et al., 2013). The model is fully prognostic for all carbon and nitrogen state variables in the vegetation, litter, and soil organic matter. The seasonal timing of new vegetation growth and litterfall for each PFT is also prognostic, responding to soil and air temperature, soil water availability, and day length. When the biogeochemistry model is active, leaf area index (LAI), stem area index (SAI), canopy top and bottom heights are calculated prognostically, which are utilized by the biophysical model. The carbon and nitrogen allocation routines in CLM4.5 determine the fate of newly assimilated carbon coming from the calculation of photosynthesis, and available mineral nitrogen coming from plant uptake of mineral nitrogen in the soil or being drawn out of plant reserves. The model treats maintenance and growth respiration fluxes separately; maintenance respiration is defined as the carbon cost to support the metabolic activity of existing live plants, while growth respiration is defined as the additional carbon cost for the development of new growth.

The CLM4.5 phenology parameterization consists of several algorithms controlling the transfer of stored carbon and nitrogen out of storage pools for new growth and into litter pools for losses of growth (Oleson et al., 2013). PFTs are classified into three distinct
phenological types that are represented by independent algorithms: an evergreen type that has some fraction of annual leaf growth displayed for longer than one year; a seasonal-deciduous type with a single growing season per year controlled mainly by temperature and day length; and a stress-deciduous type with the potential for multiple growing seasons per year, controlled by temperature and soil moisture conditions. Within the evergreen phenology algorithm, litterfall is specified to occur only through the background litterfall, dependent on specified leaf longevity; there are no distinct periods of litterfall for evergreen types, just a slow and continuous shedding of foliage. The seasonal-deciduous phenology algorithm is based on the parameterizations for leaf onset and senescence for temperate deciduous broadleaf forest (Lawrence et al., 2011). Initiation of leaf onset is triggered when a common degree-day summation exceeds a critical value, and leaf litterfall is initiated when day length is shorter than a critical value. In relatively warm climates, onset triggering depends solely on soil water availability; in cold climates, onset triggering depends on both accumulated soil temperature summation and adequate soil moisture. Any one of the following three conditions is used to trigger the initiation of leaf senescence: a sustained period of dry soil, a sustained period of cold temperature, or a day length shorter than six hours.

**Crop model**

CLMs default list of PFTs includes an unmanaged crop, essentially treated as a second C3 grass PFT (Levis et al., 2012; Oleson et al., 2013). In CLM4.5, a crop model based on the AgroIBIS (Kucharik et al., 2000) crop phenology algorithm has been added, consisting of three distinct phases. Phase 1 starts at planting and ends with leaf emergence, phase 2 continues from leaf emergence to the beginning of grain fill, and phase 3 starts from the beginning of grain fill and ends with physiological maturity and harvest.

Within the specified planting dates (calendar dates specified as the earliest and latest planting dates), the 10-day running mean of the 2-meter air temperature and 10-day running mean of the daily minimum 2-meter air temperature must be greater than the crop-specific
Coldest planting temperatures; as well as the 20-year running mean of growing degree days tracked from April through September must be higher than the minimum growing degree day requirement (Oleson et al., 2013). If these requirements are not met within the planting period, the crops are planted after the last day in the planting period, as long as the 20-year running mean of growing degree-days (GDD) is greater than zero. At planting, the model updates the average GDD necessary for the crop to reach vegetative and physiological maturity. According to AgroIBIS, leaf emergence (phase 2) occurs when the GDD of soil temperature to 0.05 m depth tracked since planting reaches 3 to 5% of the GDD to physiological maturity. The LAI generally increases and reaches a maximum value during phase 2. For phase 3 to begin, GDD for 2-m air temperature must reach 40 to 70% of GDD to physiological maturity. In phase 3, the leaf area index begins to decline in response to a background litterfall rate calculated in the CN part of the model. Harvest is assumed to occur as soon as the crop reaches maturity or the number of days past planting reaches a crop-specific maximum; harvest occurs in one time step using CNs leaf offset algorithm.

CLM4.5 introduces three new PFTs: corn (CLM’s only C4 crop), soybean, and temperate cereals, i.e. spring wheat and winter wheat (Levis et al., 2012). Temperate cereals represent wheat, barley, and rye, assuming that these three crops have very similar characteristics and could be treated as one PFT. The changing of several PFT parameter values following AgroIBIS, further distinguishes corn (a C4 crop), soybean, and temperate cereals from the unmanaged crop. The most notable difference between C3 and C4 photosynthesis is that the C4 photosynthetic pathway allows for stomata to close more often, thus transpiring less, allowing for higher water-use efficiency in C4 plants. With the crop model active in CLM4.5, the vegetated land unit is split into two parts, unmanaged and managed. PFTs in the unmanaged land unit all share the same below ground properties per grid cell, including water and nutrients, while PFTs in the managed land unit occupy separate soil columns and do not interact with each other below the ground, and thus do not compete for water and nutrients. Having PFTs in the managed land unit separated allows for different management practices, such as irrigation and fertilization, for each crop PFT.
CLM4.5 includes the option to irrigate cropland areas. The model application of irrigation responds dynamically to the soil moisture conditions simulated within the model (Oleson et al., 2013). When irrigation is enabled, the crop areas of each grid cell are divided into irrigated and rainfed fractions according to an input dataset of areas equipped for irrigation. Irrigated and rainfed crops are placed on separate soil columns, so that irrigation is only applied to the soil beneath irrigated crops. In irrigated croplands, a check is made once per day to determine whether irrigation is required on that day, this check is made in the first time step after 6 AM local time. Irrigation is required if crop leaf area is greater than zero, and water is the limiting factor for photosynthesis.

CLM4.5 adds nitrogen directly to the soil mineral nitrogen pool to meet crop fertilization demands (Oleson et al., 2013). CLM4.5’s managed land unit ensures that natural vegetation will not have access the fertilizer applied to crops. Fertilizer application begins during the leaf emergence phase (phase 2) of crop development and continues for 20 days, the fertilization helps reduce large losses of nitrogen from leaching and denitrification during the early stage of crop development.

2.1.3 Other CESM Components

The River Transport Model (RTM) was previously part of CLM and was developed to route total runoff from the land surface model to either the active ocean or marginal seas, which enables the hydrologic cycle to be closed (Vertenstein et al., 2013). This is needed to model ocean convection and circulation, which is affected by freshwater input. The RTM is generally run at a time step greater than that of the CLM because of computational constraints. The total runoff from the land model at each time step is accumulated by the driver until the RTM is invoked. The runoff at the land model resolution is interpolated by the coupler to the resolution of RTM and converted to units volume per second.

The ocean model is an extension of the Parallel Ocean Program version 2 (POP2) from Los Alamos National Laboratory (Danabasoglu et al., 2011; Vertenstein et al., 2013). POP2 solves the primitive equations in general orthogonal coordinates in the horizontal with the
hydrostatic and Boussinesq approximations. A linearized, implicit free-surface formulation is used for the barotropic equation, which allows variations of the surface layer thickness (Danabasoglu et al., 2011). Runoff from the RTM is discharged to the ocean model via the flux coupler. These runoff fluxes are treated as surface freshwater fluxes, and they are distributed over coastal ocean points near the river mouths, with higher concentrations at the mouths.

The sea-ice component, the Community Ice CodE model version 4 (CICE4), is an extension of the Los Alamos National Laboratory sea-ice model and was developed through collaboration within the CESM Polar Climate Working Group (Vertenstein et al., 2013). CICE4 is supported on high and low resolution Greenland Pole and tripole grids, which are identical to those used by POP (Bailey et al., 2013). CICE4 is a dynamic-thermodynamic model that includes a subgrid-scale ice thickness distribution. CICE4 uses the energy conserving thermodynamics of and has multiple layers in each thickness category, also accounting for the influences of brine pockets within the ice cover.

2.2 Model Simulations

The CESM with active components of CAM4, CLM4.5, POP2, CICE4 and RTM are used for the model simulations in this study. The simulations are run at an atmospheric model resolution of 0.9° x 1.25° and a nominal 1-degree ocean resolution grid with a displaced pole over Greenland for present day, year 2000, initial conditions. Before starting the coupled runs, a spin-up simulation for the land surface is implemented to achieve a steady state for the carbon and nitrogen processes of the interactive phenology. The CLM4.5 spin-up procedure is a 650-year offline simulation with present day atmospheric forcing, done by repeatedly cycling through the Qian et al. (2006) input dataset; the last land state from the offline simulations is then used at the land initial condition in the coupled simulations. A separate spin-up simulation is done for each coupled experiment as explained below. In simulations utilizing tropical crops (see Section 2.3), the crop model and irrigation models are active. In all simulations, the fire model (see Section 2.4) is turned off.
Table 2.1: Summary of CESM simulations

<table>
<thead>
<tr>
<th>Simulation Name</th>
<th>Active Model Components</th>
<th>Length</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control (CON)</td>
<td>CAM4, CLM4.5, POP2, CICE4, RTM</td>
<td>250</td>
</tr>
<tr>
<td>Complete Deforestation (DEF)</td>
<td>CAM4, CLM4.5, POP2, CICE4, RTM</td>
<td>250</td>
</tr>
<tr>
<td>East Deforestation (E.DEF)</td>
<td>CAM4, CLM4.5, POP2, CICE4, RTM</td>
<td>250</td>
</tr>
<tr>
<td>West Deforestation (W.DEF)</td>
<td>CAM4, CLM4.5, POP2, CICE4, RTM</td>
<td>250</td>
</tr>
<tr>
<td>50% Deforestation (DEF.50)</td>
<td>CAM4, CLM4.5, POP2, CICE4, RTM</td>
<td>250</td>
</tr>
</tbody>
</table>

A summary of the simulations performed is given in Table 2.1. Five fully coupled CESM simulations have been done: a control run (CON), complete deforestation of the Amazon region (DEF), eastern deforestation (E.DEF), western deforestation (W.DEF), and 50% deforestation (DEF.50). Each of these fully coupled simulations has a length of 250 years, in which the first 125 years are used as spin-up for the atmosphere and ocean, and the last 125 years is used for analysis. In addition to the model simulations listed in Table 2.1, offline simulations were performed to test tropical crop PFTs that were added to CLM4.5 and to determine the fire impacts in CLM4.5.

In the DEF, E.DEF, W.DEF and DEF.50 simulations, a distribution (discussed in Section 2.3) of tropical crops is used to replace tropical forest (see Figure 2.4). Figure 2.2 displays the fractional distribution of different PFTs in the Amazon region for CON. In the E.DEF and W.DEF simulations, the Amazon is split in half to provide equal deforestation areas of the Amazon. It was determined that 62°W is the halfway mark of the annual LAI distribution in the Amazon region. The gridboxes immediately east and west of this longitude were used as a transition zone; depending on the portion being deforested, these gridboxes were deforested 66% and 33%. In the DEF.50 simulation, 50% of the forest area is removed from each gridbox and replaced with the tropical crop distribution.
2.3 Development of Tropical Crops

2.3.1 Shortcomings in the CLM Crop Model

In performing offline CLM4 simulations, the need to develop more realistic PFTs for the tropics became apparent. In the simulations, the tropical broadleaf evergreen tree PFT was replaced with the unmanaged crop PFT and C3 grass PFT. It was thought that there would be a reduction in LAI when replacing the broadleaf evergreen trees; however, it was found that there was a basin wide increase in LAI as seen in Figure 2.3.

It was determined from discussions with colleagues at the 2013 Land Model Working Group (LMWG) meeting at NCAR that the crop and C3 grass PFTs were parameterized solely for the mid-latitude conditions. The winter season temperature in the Amazon does not get cold enough to trigger senescence; the survival temperature for C3 grass is -17°C and the establishment temperature for C3 grass is 15.5°C, while the planting temperature for
managed crops is 7°-13°C. The Amazon has an annual average temperature of approximate 27°C, meaning minimum temperature thresholds for each PFT are always met. Another aspect is the greater moisture availability in the Amazon; plants are not usually stressed by a lack of available moisture.

This misrepresentation in LAI is significant when it comes to modeling land-use change. LAI impacts numerous processes, including, but not limited to, the radiation budget (albedo, longwave radiation), gross primary production (photosynthesis), evaporative fraction (transpiration, sensible heat from ground and vegetation), and the carbon cycle. Misrepresentation of LAI for replacement vegetation in the tropics will lead to errors in fully coupled global simulations. It was determined that the development of tropical crop PFTs is necessary as replacement vegetation in large-scale tropical land-use change simulations. Without adequate replacement vegetation, the simulation of large-scale land-use change will not give an accurate representation of the local, regional or global responses.

2.3.2 Adding Tropical Crops in CLM4.5

The default crop PFTs in CLM4.5 include soybean, corn, spring wheat and winter wheat; with each crop type having rainfed and irrigated PFT versions. Using the Sacks et al.
(2010) and Portmann et al. (2010) datasets of global crop distribution, it was determined
by summing crop acreage over Brazil that the most prevalent crops were soybean, corn (two
versions), cotton, rice and sugarcane. These crops were then selected as the tropical crops
that should be added to CLM4.5. Two separate corn PFTs were added to simulate the
two separate corn harvests that occur in the region. Given the long growing season in the
tropics, after the first corn harvest of the year a second crop of corn is planted and harvested
later in the year. For each crop added, a rainfed and irrigated PFT were constructed.

The new tropical crops are based on existing crops in CLM4.5, with adjustments to
some of the physiology parameters to get realistic behavior. Tropical soybean was based
on the existing soybean PFT and tropical corn based on the existing corn PFT. Tropical
sugarcane is also based on the existing corn PFT, tropical rice is based on the existing
spring wheat, and tropical cotton is based on the existing soybean.

Sugarcane was based on the existing corn because sugarcane is a C4 plant that behaves
similarly to corn, which is the only C4 crop in CLM4.5. Rice was based on the existing
spring wheat because they are both cereal grain crops. Cotton uses soybean as a basis
because they are both bushy C3 crops, with neither being a cereal grain crop, as the other
C3 crops in CLM4.5 are. It is of note that sugarcane is a multi-year perennial crop, while
all the other crops are annual; CLM4.5 does not currently have the capability to simulate
perennial crops. Sugarcane was modeled to have a planting date just after the previous
harvest; this was done with the intention of simulating large crop coverage throughout the
year with a decrease once a year when a portion of the sugarcane is typically harvested.

The Sacks et al. (2010) data was used to determine planting dates, GDD, maximum LAI
and maximum number of days to plant maturity for the tropical crops being added; Table
2.2 shows original crop PFT and tropical crop PFT parameters that were modified. In
addition to changing those physiology parameters; the albedo and radiative transmissivity
of crop leaves were changed to match those of Bonan (2008a). The amount of fertilizer
applied to each crop was modified to allow for a more realistic seasonal cycle.
Table 2.2: Key parameters used in developing CLM4.5 tropical crops. Planting dates are in the format of last two digits being the day of the month, the preceding digit(s) being the month number (example: 415 is 15 April). “–” denotes a parameter that is not specified.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>C3 Crop</th>
<th>Corn</th>
<th>Spring Wheat</th>
<th>Winter Wheat</th>
<th>Soybean</th>
<th>Tropical Soybean</th>
<th>Tropical Corn</th>
<th>Tropical Corn (2)</th>
<th>Tropical Sugar-cane</th>
<th>Tropical Rice</th>
<th>Tropical Cotton</th>
</tr>
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<td>Photosynthesis</td>
<td>C3</td>
<td>C4</td>
<td>C3</td>
<td>C3</td>
<td>C3</td>
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<td>C4</td>
<td>C4</td>
<td>C3</td>
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<td>7</td>
<td>6</td>
<td>6</td>
<td>5</td>
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<td>6</td>
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<tr>
<td>Leaf Transmittance - near IR</td>
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<td>0.34</td>
<td>0.34</td>
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</tr>
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<td>0.05</td>
<td>0.05</td>
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</tr>
</tbody>
</table>

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2.3.3 Tropical Crop Distribution after Deforestation

The 5-minute spatial resolution Portmann et al. (2010) data was regridded for use in CLM4.5. At each CLM gridbox in the specified domain (85°W-35°W, 30°S-13°N), those with a sum of tree PFTs (tropical broadleaf evergreen and tropical broadleaf deciduous) percentage greater than zero were deforested; all existing PFTs in that gridbox were removed. Each respective deforested gridbox is checked for crops being present in the regridded Portmann data. If any crops are present in a deforested gridbox, the ratio of acreage for present crops is used to determine the percentage of each crop to be placed in the gridbox. There is a maximum of five crops allowed in a gridbox. If all six crops are present, the lowest acreage crop was not used in that gridbox. For deforested gridboxes with no crops present, a Cressman Analysis is used to interpolate nearby crops to the respective gridbox. The calculated distribution of the tropical crops can be seen in Figure 2.4.
2.4 Fire model

When coupling CLM4.5 with CAM, specific humidity has been found to be too low over the Amazon region (W. Sacks and D. Lawrence, personal communications). Fires in CLM4.5 are invoked as a function of relative humidity, soil wetness, temperature and precipitation (Oleson et al., 2013). With specific humidity being too low, relative humidity was low enough to engage the fire model in vast areas of the Amazon region, predominantly regions neighboring the closed canopy forests (gridbox with greater than 60% tree PFT). Along with a reduction in humidity, there is a decrease in precipitation total that is enough to invoke fire in the closed canopy as well. Figure 2.5 displays the change in LAI due to fire impacts in the Amazon region. It can be seen that fire starts in year 1 by occurring along the edge of the closed canopy and LAI is reduced. As time continues, LAI become significantly reduced in the northeast by year 4 and large reductions in LAI propagate westward into the closed canopy in the following years.

CLM4.5 was tested in short coupled simulations with the fire model active and inactive. The results showed that canopy height was no longer decreasing with the fire model inactive, although the LAI was reduced by approximately 30% from offline simulations. These LAI impacts with the fire model active are much more severe to both the canopy height and LAI. Although with the fire model inactive the LAI is reduced, presumably from further impacts with low humidity and precipitation impacting the phenology algorithms that were previously discussed.

Thus, it has been determined that the simulations used in this study should have the fire model turned off. The LAI impacts due to deforestation will still be a large change capable of producing a significant signal, even with a reduction of LAI in the coupled model. In addition, the large changes exhibited in surface roughness will also be present to provide a boundary condition to the atmosphere capable of demonstrating the impacts of large-scale land-use change.
Figure 2.5: Change in LAI for a 12-year coupled run with the fire model active, year number in lower-left corner of each map.
Chapter 3: Local Response to Amazon Deforestation

This section discusses the local response to Amazon deforestation; highlighting changes in surface temperature, precipitation, surface fluxes, and land-atmosphere coupling. All analysis in this section is done with three custom-defined seasons: NDJFM, AMJ and JASO. NDJFM largely coincides with crop growth in the region south of the equator and planting north of the equator. AMJ is the main growing season north of the equator. JASO is predominantly a period after harvest has occurred and planting south of the equator is taking place in the last month. Additionally, these seasons correspond to the seasons of peak precipitation as NDJFM has precipitation predominantly south of the equator, AMJ precipitation is centered on the equator and extends into northern South America, and JASO is the driest period for a majority of the region with precipitation centered over the northwest portion of South America.

Due to deforestation, the initial seasonal mean changes to the land surface can be seen in Figure 3.1. There is a basin wide increase in surface albedo across the deforested region. In the closed canopy region where the highest percentages of broadleaf evergreen trees are located, there is a large reduction in both LAI and canopy height across all seasons. To the southeast of that region, an area where C4 grass was predominant, there is an increase in both LAI and canopy height in NDJFM, the main growing season of the dominant crops, soybean and rice, in that region. The other months show a general decrease in LAI and canopy height in that region.

This chapter is the basis of Badger and Dirmeyer (2015a), submitted to Hydrological and Earth System Science.
Figure 3.1: Changes to surface properties after deforestation in NDJFM, AMJ and JASO; albedo (top-row), leaf area index (middle-row) and canopy height [m] (bottom-row). Shading indicates significance at the 95% confidence level.
Figure 3.2: Change in surface temperature [K] for NDJFM, AMJ and JASO. Shading indicates significance at the 95% confidence level.

3.1 Temperature

As can be seen in Figure 3.2, in the initially dense forest region there is an increase in surface temperature in all seasons; a majority of the region warms 1-3 K with the central region warming more than 7 K. To the southeast, there is a region of temperature decrease, typically less than 3 K. This temperature decrease is largely over the region that was predominantly C4 grass. McGuffie et al. (1995) noted that changes in surface temperature over the deforested region are dipolar: an increase over the central and eastern Amazon and a decrease to the southwest of the deforestation. The region of decrease is shifted eastward in these findings, but such a dipolar change has precedent. Despite the region of cooling, the areal average for each season shows an increase; +0.8 K in NDJFM, +1.6 K in AMJ, and +2.1 K in JASO.

The contrast in temperature change between the densely forested and C4 grass areas becomes more apparent in the change of maximum monthly surface temperature. The forested region experiences an increase in all months, typically between 2-6 K. In the C4 grass area, the maximum monthly surface temperature decreases from August to January by 4-6 K, with the remaining months having a mixed change between -2 to 2 K. The same pattern tends to hold up for minimum monthly temperature, with the changes about half
the magnitude. The overall range in extremes for the densely forested area increases by 2-4 K, while in the C4 grass area, the range of extremes is reduced by 2-4 K from August to January and increase less than 2 K in the remaining months. It is worth noting that C4 grass in CLM can behave unrealistically at times by dying off and then regrowing a couple months later (Dirmeyer et al., 2013), which can affect surface temperature drastically.

The annual areal average increase in surface temperature of 1.4 K is consistent with previous modeling studies; Costa and Foley (2000) found a 1.4 K increase, Snyder et al. (2004) found a 1.5 K increase and Snyder (2010) found a 1.2 K increase. However, some studies found smaller or larger temperature increases: 0.6 K (Henderson-Sellers et al., 1993), 0.3 K (McGuffie et al., 1995), 0.3 K (Ramos da Silva et al., 2008), 2.5 K (Nobre et al., 1991) and 2.5 K (Shukla et al., 1990). The results in this study lie within the range of previous findings.

### 3.2 Precipitation

There is a significant decrease of at least 1 mm/day in precipitation over the originally densely forested region throughout the year, with some areas experiencing decreases larger than 4 mm/day, see Figure 3.3. A majority of this region sees decreases of more than 50%. During NDJFM, when a majority of the Amazon region experiences at least 8 mm/day in precipitation in the control simulation, there is a largely statistically significant decrease in precipitation for the deforested region. An area of increase is present in a region that is mainly irrigated rice. During AMJ when precipitation is largely occurring within a few degrees of the equator, there is a significant decrease across this region of the equator, while a significant increase is present to the south. The driest season in the control simulation, JASO, has a significant decrease in precipitation over much of the deforested region. All seasons experience a decrease in the areal average: -0.27 mm/day in NDJFM, -0.37 mm/day in AMJ, and -0.44 mm/day in JASO.
Most of the precipitation changes can be explained by changes to convective precipitation, which decreases in all seasons (not shown), with the only exception being the region with irrigated rice. The reduction in convective precipitation suggests changes in flux partitioning at the surface may modify the properties and growth of the planetary boundary layer, as well as the land-atmosphere coupling in the region.

The decreases exhibited in this study are consistent with previous modeling studies; however, the magnitude of the decrease is smaller. This study found an annual areal average decrease of 0.35 mm/day, while previous studies found decreases of 0.7 mm/day (Costa and Foley, 2000), 0.4-0.7 mm/day (Hasler et al., 2009), 1.6 mm/day (Henderson-Sellers et al., 1993), 1.2 mm/day (McGuffie et al., 1995), 1.4 mm/day (Snyder et al., 2004; Snyder, 2010), and 0.8 mm/day (Werth and Avissar, 2002). The smaller decrease in precipitation may be due to previously mentioned model shortcomings with low humidity and less climatological precipitation in the region.

### 3.3 Radiation and fluxes

Net radiation is shown (Figure 3.4) to be significantly reduced over the densely forest region in all seasons, typically by 30-50 W/m². To the southeast over the C4 grass area,
increase is shown during NDJFM, changes between -10 and 10 W/m$^2$ are present in AMJ and decreases of 10 W/m$^2$ exist in JASO. These changes are driven by changes to albedo (seen in Figure 3.1) and impacts the partitioning of latent and sensible heat flux.

Latent heat flux is primarily reduced across the region in all seasons; the major exception is an increase during NDJFM in the former C4 grass area. Sensible heat flux increases in the formerly densely forested area in all seasons and is surrounded by a region of decrease in sensible heat flux. There is an increase in sensible heat flux in the southeast during both NDJFM and AMJ, while JASO has a mix of both increases and decreases, with most of the area not experiencing a significant change. The annual areal average of latent heat flux and sensible heat flux over land area both decrease, -8.1 W/m$^2$ and -1.7 W/m$^2$ respectively. This change in the fluxes has reduced the evaporative fraction (Figure 3.5) in the region and indicates that the Amazon would shift to a drier climate.

Evaporative fraction (Figure 3.5) is the ratio of latent heat flux to the sum of latent and sensible heat fluxes. After deforestation, the nearly the entirety of the deforested region in AMJ and JASO have significant decreases in evaporative fraction, indicating a drier climate in the region. NDJFM experiences an increase of evaporative fraction over a large portion of the area, this is due to being the season of main crop growth over that area. The formerly densely forested region in NDJFM experiences a decreases in evaporative fraction, this is probably due to the deeper root profile of tree PFTs that would have access to a larger soil moisture reservoir.

### 3.4 Land-Atmosphere Coupling

A two-legged coupling metric (Dirmeyer, 2011; Guo et al., 2006) uses correlations between a land surface state variable (soil moisture) and surface flux (latent heat) as a means to assess terrestrial climate feedbacks, or a surface flux (sensible heat) and an atmospheric property (PBL height) for the atmospheric climate feedback. It is used here to describe the feedbacks present in the system and how they have changed after deforestation. Positive values in
Figure 3.4: Changes of surface energy fluxes in NDJFM, AMJ and JASO; net radiation \([W/m^2]\) (top-row), latent heat flux \([W/m^2]\) (middle-row) and sensible heat flux \([W/m^2]\) (bottom-row). Shading indicates significance at the 95\% confidence level.
these two instances would imply that the land surface is controlling the feedback. We multiply these correlations by the standard deviation (SD) of the response variable (latent heat and PBL height respectively) to determine the magnitude of the feedback (Guo et al., 2006).

In the terrestrial leg of the coupling (Figure 3.6) for the control simulation, a large band of negative values during NDJFM corresponds to the heavy rains during that season when soil moisture is not a limiting factor for surface fluxes. As the rains shift throughout the year, this region shifts accordingly. During the drier seasons in the south, the sign switches to positive, indication that soil moisture is controlling the latent heat flux (cf. Dirmeyer et al., 2013).

After deforestation, the previously densely forested areas become more strongly coupled throughout the year (Figure 3.6). This is probably due to the shallower roots of crops, which have access to a smaller soil moisture reservoir. There are also large areas of decreased coupling, particularly over the southeast in JASO and south of the densely forested area in NDJFM. During AMJ, nearly the whole region sees an increase in coupling.

The changes in coupling can occur due to changes in the correlation, variability, or both. In NDJFM, the correlation increases in 54.8% of the region and flux variability decreases in 56.4% of the region. Neither component appears to be the leading agent of the changes; the
Figure 3.6: Terrestrial leg of coupling strength [W/m²] between soil moisture and latent heat flux for the control simulation (top-row) and change due to deforestation (bottom-row) for NDJFM, AMJ and JASO.
changes in NDJFM (the rainy season) are largely atmospherically driven due to changes in precipitation. Areas with the largest reduction in precipitation have correlation increases; they also have increases in variability and are becoming more strongly coupled.

In AMJ and JASO, the changes in correlation are much larger: 69.5% and 76.6% of the region has an increase in correlation, respectively. Increases in correlation do not necessarily imply increased coupling, while a majority of the region in AMJ has stronger coupling, JASO has a majority of the region showing a decrease in coupling. JASO has a decrease in variability for 62.7% of the region, with 46.2% of the region having an increase in correlation and decrease in variability, largely taking place in the southeast where there was lower initial tree cover. In contrast, the more densely forested regions largely experience an increase in correlation and an increase in variability.

For the atmospheric leg of the coupling, in the control run, the entire region is positively coupled (Figure 3.7). The areas of strongest coupling occur in locations that were initially less tree-covered, as the dense canopy acts to dampen the coupling between surface sensible heat flux and PBL height.

In all seasons, the densely forested areas have an increase in coupling after deforestation (Figure 3.7). The southeast region largely experiences a decrease in coupling during all seasons. The largest contrast between the densely forested area and the southeast occurs in JASO, which is after most of the crops have been harvested and LAI is low.

For the atmospheric leg, a majority of the region either experiences an increase in both correlation and variability or decrease in both. There are co-located correlation and variability increases over 31.0%, 41.6%, and 33.3% of the region for NDJFM, AMJ and JASO respectively. These regions are predominantly along the southeast coast where increased temperature and decreased precipitation occur, and in the previously forested areas. Regions experiencing decreases in both were 36.9% 29.7% and 40.2% for those same seasons. These changes largely occurred in the southeast area where lower initial tree cover is located.
Figure 3.7: Atmospheric leg of coupling strength [m] between sensible heat flux and planetary boundary layer height for the control simulation (top-row) and change due to deforestation (bottom-row) for NDJFM, AMJ and JASO.
3.5 Discussion of Local Response

Replacement of natural vegetation with crops typical of tropical agriculture over the Amazon results in an albedo increase, lowering net radiation, which in turn modifies the surface fluxes. As previously made notion to, latent heat flux is largely reduced across the domain, with the exception being the former C4 grass region in NDJFM; sensible heat flux has a more detailed spatial change with decreases in all seasons over the former densely forested area and a seasonality to the changes in the surrounding regions. The areal averages for latent heat flux and sensible heat flux are reduced, but the evaporative fraction decreases, modifying the region toward a drier climate. Combining the surface temperature increase with the surface flux changes, a warmer, drier and deeper PBL is expected. There is a decrease in precipitation, largely due to decreased convection. By modifying PBL properties and PBL growth, modified interaction between the PBL and the free atmosphere will decrease vertical moisture transport and increase vertical heat transport. These changes in vertical transport provide a mechanism that can impact the circulation and may affect remote regions, with large scale circulation changes enhancing the precipitation changes being discussed in the coming chapter.

An added level of complexity that previous studies did not consider is irrigation. The irrigation impact is difficult to isolate, due to the gridboxes with irrigated rice also having other crops present. Irrigation adds water to the surface when water is a limiting factor for photosynthesis and can have an impact on land-atmosphere interactions. Irrigation does appear to have an impact on the coupling between land and atmosphere (Figure 3.8). Irrigation is active in eight months (ONDJFM in the southern hemisphere and JFMAM in the northern hemisphere) when rice is widely grown. In the months that irrigation is added, there is a negative correlation between irrigation water added and the change in the terrestrial leg of the coupling. The more irrigation water that is added, the less coupled the soil moisture becomes from the latent heat flux.

By affecting the surface coupling, irrigation can also impact the atmospheric leg of the coupling (Figure 3.8). A negative relationship between irrigation water added and the
Figure 3.8: Top row: Change in terrestrial leg of coupling strength [W/m²] versus irrigation water added [mm/day] for irrigated gridboxes in NDJFM, AMJ and JASO. Bottom row: Change in atmospheric leg of coupling strength [m] versus initial tree cover percentage for NDJFM, AMJ and JASO. Shaded dots represent irrigated gridboxes with the shading being equivalent to the shading for irrigation water added [mm/day] in top row.
change in SH-PBLH (atmospheric leg) coupling further shows that irrigation is modifying land-atmosphere interactions.

Although irrigation is shown to have an impact on the atmospheric leg of the coupling, the larger contributor appears to be the percentage of tree cover lost (Figure 3.8). The coupling changes are largely the same for non-irrigated gridboxes with original tree percentage less than 80%, typically between -50 and 50 m. JASO, the driest season, does have a larger spread, but comparable magnitudes of increases and decreases. When the initial tree cover is greater than 80%, the coupling strength is predominantly increasing and has a greater magnitude of the change. This signal is also common in climate change scenarios driven by greenhouse gas increases (Dirmeyer et al., 2013), suggesting land use change could further amplify sensitivity to land surface anomalies in the tropics.

Irrigation largely decreases the coupling strength when the initial tree cover is less than 80% and increases the magnitude of the change. When the initial tree cover is greater than 80%, the gridboxes that experience a decrease in coupling are typically irrigated; with the more strongly irrigated gridboxes showing the largest decreases and less irrigated gridboxes showing an increase in coupling that is comparable to non-irrigated gridboxes. Just as with the terrestrial leg, more irrigation water added decreases the coupling strength of the atmospheric leg of the coupling.

By using a realistic crop distribution in the Amazon region, as opposed to homogenous vegetation coverage used in previous studies, there is still general agreement with previous modeling studies. The higher resolution and heterogeneity of the land cover shows smaller scale features and regions of opposite change, particularly in the southeast Amazon where the region has higher coverage of C4 grass. With crops being planted in different regions at different times of the year, a level of complexity not present in previous Amazon deforestation studies, seasonality to land-surface changes that were not previously modeled, are now seen.

A warming and drying of the region has impacted how the land-surface and atmosphere interact. By modifying the flux partitioning between latent and sensible heat fluxes, the
region shifts to drier climate with a warmer, drier and deeper PBL. By altering how the
PBL grows, interaction with the free atmosphere is altered; this can lead to a warmer and
drier atmospheric column above the region and may cause impacts to remote regions by
modifying the general circulation and transports of moisture and heat.
Chapter 4: Remote Response to Amazon Deforestation

Replacing natural vegetation with tropical crops over the Amazon region impacts vertical transport of heat and moisture, modify the interaction between the atmospheric boundary layer and the free atmosphere. Vertical velocity is decreased over a majority of the Amazon region, shifting the ascending branch and modifying the seasonality of the Hadley circulation over the Atlantic and eastern Pacific oceans. Using a simple model that relates circulation changes to heating anomalies and generalizing the upper-atmosphere temperature response to deforestation, agreement is found between the response in the fully-coupled model and the simple solution. These changes to the large-scale dynamics significantly impact precipitation in several remote regions, namely sub-Saharan Africa, Mexico, the southwestern United States and extratropical South America.

This chapter is the basis of Badger and Dirmeyer (2015c), submitted to Climate Dynamics.

4.1 Circulation Changes

Significant changes in the tropics are found in the zonal mean meridional mass stream function (Figure 4.1), particularly in the annual mean and seasonal means of DJF, MAM and SON. JJA shows little significant change; this could largely be because the rising branch of the Hadley circulation has shifted north of the main deforested area in this season. Significant changes occur above 800 hPa and are centered over the rising branch with the largest impact near 300 hPa. These changes show a decrease in the strength of the Hadley cell in the Northern Hemisphere and a strengthening in the Southern Hemisphere Hadley cell. The overall meridional mass transport is modified, less poleward transport in the Northern Hemisphere and increased poleward transport in the Southern Hemisphere.
Figure 4.1: Meridional mass stream function \([10^{10} \text{ kg/s}]\) using the globally averaged zonal v-wind. Black contours indicate the control simulations meridional mass stream function, positive values indicate a clockwise rotation and negative values indicate a counter-clockwise rotation. Colored contours indicate changes and shading indicates significant change at the 95% confidence level. Panels on the right show meridional mass transport globally (black line), in the northern hemisphere (long-dashed blue line) and the southern hemisphere (short-dashed red line).
In the cross-equatorial cell during DJF, significant weakening can be seen in the rising branch as well as in the poleward motion near the top of the troposphere. This seasonal Hadley circulation shows decreased transport into the Northern Hemisphere; the cross-equatorial cell is integral in supplying heat and moisture to the Northern Hemisphere in this season.

The transition seasons of MAM and SON, when the rising branch is switching hemispheres, show significant changes. MAM shows an increase in stream function along the rising branch, a strengthening in the Northern Hemisphere and a weakening in the Southern Hemisphere. In addition, centered in the Northern Hemisphere Hadley cell is a significant decrease, indicating a weakening in the core of this circulation. These changes indicate a shifting and weakening of the MAM Hadley cells. The zonal mean around the entire globe can hide the complexity of regional changes. Looking at changes in the zonal mean of precipitation (Figure 4.2), MAM has different changes over the eastern Pacific (120°W - 80°W) and Atlantic (40°W - 0°). In the eastern Pacific, there is a southward shift in precipitation, while the Atlantic shows a decrease in precipitation during MAM. The changes in precipitation indicate different adjustments in the Hadley circulation over different ocean basins to the east and west of the deforested region.

In SON, where the rising branch is shifting from the Northern to Southern Hemisphere, a large significant decrease in stream function (Figure 4.1) is found. The decrease in stream function indicates a weakening of the rising in the Northern Hemisphere cell and a strengthening in the rising motion of the Southern Hemisphere cell. Just as in MAM, there are separate changes occurring over the eastern Pacific and the Atlantic. Looking at the zonal mean of precipitation (Figure 4.2), a southward shift in the maximum precipitation occurs over the eastern Pacific, while there is an increase in the maximum precipitation over the Atlantic. Associated with these changes in the Hadley circulation is a decrease in poleward transport in the Northern Hemisphere and increased poleward transport in the Southern Hemisphere above 400 hPa.

In JJA, there is a southward shift in precipitation over the eastern Pacific and northward
shift in the Atlantic. There is little significant change in the meridional stream function (Figure 4.1) in JJA. These opposite changes over the eastern Pacific and Atlantic likely cancel in the zonal mean.

Looking at the velocity potential field in the upper troposphere (~200 hPa; Figure 4.3), the dominant feature annually and in each season is the significant reduction in upper-level divergence over the deforested region. This feature was hypothesized by Snyder et al. (2004), where it was stated that altering the natural vegetation of tropical forests could impact the intensity of high-level tropical outflow. The significant change in velocity potential over the deforested region has an elongated extension towards the southeast. This feature corresponds to a weakening in the South Atlantic Convergence Zone (SACZ). Decreased divergence aloft is paired with decreased convergence in the lower troposphere in the SACZ. This is discussed further in the context of precipitation in the next section.

Another feature present annually and seasonally is a decrease in large-scale upper-tropospheric convergence in remote regions. Annually, and in DJF, MAM and SON, the decreased convergence occurs to the northeast of the deforested region over North America.
Figure 4.3: Significant change at the 95% confidence level for annual and seasonal velocity potential $[10^{10} \text{ m}^2/\text{s}]$. Vectors experiencing a significant change to either the $u$ or $v$ component for divergent wind are overlaid.
and the neighboring Pacific Ocean, with the largest change occurring in DJF at approximately 30°N, which is the northern edge of the cross-equatorial Hadley cell. Significant decreases in the upper-level flow using the meridional mass stream function (Figure 4.1) were found during that season. SON features a large area of significant change (Figure 4.3) with significant reductions in convergence extending across the equatorial and northern hemisphere Pacific. SON was shown to have the most significant changes to the meridional circulation, as well as the largest reductions in upper-level poleward mass transport in the Northern Hemisphere.

In JJA, a significant decrease in convergence (Figure 4.3) appears over the central Pacific near the equator. This dipole of change could be indicative of a modification of the equatorial zonal circulation, i.e. the Walker circulation. The Walker circulation can impact central Pacific sea surface temperatures, making these changes of particular interest.

4.2 A Mechanism for the Changes

Gill (1980) proposed a simple solution to the tropical atmospheric circulation in the presence of anomalous diabatic heating. The solutions show modifications to the horizontal wind and the zonal circulations. To compare to the situation of Amazon deforestation, modifications to the cases presented by (Gill, 1980, Fig. 3) can be made. Gill (1980) placed a heating anomaly north of the equator, whereas the region of deforestation occurs predominately south of the equator. Gill (1980) induced a source of diabatic warming throughout the atmospheric column. Although there is anomalous warming at the surface in response to deforestation (see Section 3.1), the response of the atmospheric column over tropical deforestation is a cooling. With decreased moisture present over the region of change, the rising air is dryer, which leads to less condensation (latent heat release) in the atmosphere above the deforested region. The cooler and drier atmospheric column leads to a significantly colder upper troposphere (Figure 4.4), with cooling to the north and south of the equator.
Figure 4.4: Significant changes at the 95% confidence level for the annual geopotential height [m] (top panel), temperature [K] (middle panel) and omega [Pa/s] (bottom panel). Note that positive changes to omega indicates less rising motion.
Following the framework laid out by Gill (1980), the expected response for this deforestation study should be decreased rising motion and a trough in the upper atmosphere due to less heating in the troposphere. This would induce increased westerlies along the equator, with strong cyclonic flow south of the equator and weaker cyclonic flow north of the equator. Accompanied by a weakening of the zonal circulation along the equator.

Figure 4.4 shows that the simulated changes are similar to those put forth by the Gill (1980) solution. The largest increases in westerly flow in the upper atmosphere (~200 hPa) occur over the region with significant change in omega [Pa/s] in the mid-troposphere (~550 hPa), showing the decreased rising motion as expected. Two regions of decreased geopotential height in the upper atmosphere are found north and south of the equator, with the stronger of the depressions in the Southern Hemisphere. The upper-level wind flow shows increased westerlies along the equator with stronger cyclonic wind flow south of the equator and slightly more cyclonic wind flow north of the equator. Analyzing the zonal mass stream function averaged from 5°S to 5°N, a weakening in this zonal circulation can also be seen, furthering the similarities to the expected response based on the Gill (1980) solution.

Further analysis was done using a simple solution from Zebiak (1982) that analyzes wind changes based on temperature anomalies. The Zebiak (1982) solution is similar to the Gill (1980) solution in the sense that wind flow is altered by SST anomalies on a horizontal plane, instead of diabatic heating. The Gill (1980) solution uses two layers with mass transfer between layers; Zebiak (1982) uses one layer with an internal mode to simulate vertical structure. Using the solution of Zebiak (1982) with an idealized version of the upper-atmosphere temperature anomaly (Figure 4.4), the wind response in the simple model (Figure 4.5) is similar to that of the observed wind response (Figure 4.4) in the fully-coupled model. In the simple solution, stronger cyclonic flow to the south and weaker cyclonic flow to the north can be seen; both present features in Figure 4.4. The largest difference between the simple solution and full-model solution is the flow across the largest temperature change; north-south flow in the simple solution and east-west flow in the full
model. Regardless of this difference, the simple solutions of Gill (1980) and Zebiak (1982) show remarkable consistency with the fully-coupled model results and provide a framework for a forcing mechanism that reaches beyond the region of deforestation.

4.3 Remote Responses

The previously discussed circulation changes have caused remote regions of change, particularly in regards to precipitation. Significant changes can be seen south of the deforested region in South America, as well as over Africa and North America.

With the weakening of the SACZ (Figure 4.3), decreased moisture convergence in that
region has been found that leads to significant increases in precipitable water over areas just south of the SACZ during austral spring and summer (Figure 4.6) and increased precipitation over that southern margin. Thus, there are potential seasonal implications for climate in the mid-latitude regions of South America. In the remaining months of the year (not shown), there is no appreciable areas of significant change to precipitable water or precipitation in the region south of the deforestation.

Significant changes in precipitation over sub-Saharan West Africa and the Congo (Figure 4.7) are also found. As previously discussed, there is a northward shift and intensification in the rising branch of the Hadley circulation over the Atlantic in JJA (Figure 4.2). This circulation change leads to increased moisture convergence over sub-Saharan West Africa, increasing precipitable water and precipitation during JAS (Figure 4.7). There appears to be a delayed response to the southward transition of the rising branch over the Atlantic. This delay, accompanied by increased precipitation in sub-Saharan West Africa, appears to lead to significantly less precipitable water and precipitation during October and November over the Congo (Figure 4.8), normally two of the rainiest months for this region. Changes in precipitation over these regions could have drastic impacts to the local climate and vegetation.

Figure 4.9 shows significant increases in precipitable water over the southwest US and Mexico during MAMJJ and increased precipitation during MAMJJ. By analyzing changes along a cross-section (105°W, 30°N to 60°W, -10°S) between northern Mexico and South America, notable changes can be seen (Figure 4.10). The lower-level winds have intensified along the transect towards Mexico. Although the specific humidity has decreased, moisture transport from the Amazon towards Mexico has increased. At the northern edge of the significant increase in moisture transport, an increase in precipitation can be seen, corresponding to the significant precipitation increases shown in Figure 4.9. It appears that additional moisture is being advected northwestward from South America, contributing to increased specific humidity, cloud cover and precipitation north of the Yucatan Peninsula along the cross-section.
Figure 4.6: Significant changes at the 95% confidence level for SONDJ precipitable water [mm] (top panel) and precipitation [mm/day] (bottom panel) over the South American region.
Figure 4.7: Significant changes at the 95% confidence level for JAS precipitable water [mm] (top panel) and precipitation [mm/day] (bottom panel) over the African region.
Figure 4.8: Significant changes at the 95% confidence level for ON precipitable water [mm] (top panel) and precipitation [mm/day] (bottom panel) over the African region.
Figure 4.9: Significant changes at the 95% confidence level for MAMJJ precipitable water [mm] (top panel) and precipitation [mm/day] (bottom panel) over the North American region.
Figure 4.10: Changes to MAMJJ wind velocity [m/s] (first panel), specific humidity [g/kg] (second panel), moisture transport [g/kg\*m/s] (third panel), and percent change in precipitation (fourth panel) along a cross section from 105°W, 30°N (A-Mexico) to 60°W, -10°S (B-South America). Black contours are control mean, colored contours are change with shading at the 95% confidence level. The red line in the percent change of precipitation denotes significant change at the 95% confidence level. Map of cross section location can be seen at the bottom.
4.4 Discussion of large-scale response

By converting the surface vegetation over the Amazon region to cropland in a coupled Earth system model, large-scale circulation changes are induced throughout the year. Shifts and magnitude changes in the rising branch of the Hadley circulation have been found, with differing changes occurring over the eastern Pacific and Atlantic. Accompanying these meridional circulation changes are significant changes to the upper atmosphere and zonal circulations.

Using the framework of anomalous low-latitude heating by Gill (1980), the predicted circulation response has been found to agree well with the Earth system model output. Using an idealized temperature anomaly and the simple solution of Zebiak (1982), the upper atmosphere temperature changes are suspected to be the driving force in modifying the upper atmosphere circulation.

These changes in large-scale circulation lead to significant changes in precipitation outside of the deforested region; particularly over southern South America, sub-Saharan West Africa, the Congo and southwest North America. Although not shown here, remote responses in a variety of atmospheric variables can be seen across the tropics, sub-tropics and extra-tropics (i.e. surface temperature, surface fluxes, PBL height and geopotential height).

The study uses monthly output from the climate model, which prevents smaller time-scale features, namely transient eddies, from being analyzed. Both stationary and transient eddies could be contributors to the changes presented in this study, largely by modifying meridional transports of heat and moisture.

With modifications to cross-equatorial heat transport due to Amazon deforestation, the thermal contrast between hemispheres could impact global monsoon patterns (Wang et al., 2013). This study helps to exhibit that Amazon deforestation can be considered a geographically isolated event. Although the results of this study are model dependent, and other climate models may or may not show similar impacts to Amazon deforestation, this study does exhibit changes globally caused by Amazon deforestation and provides a realistic physical mechanism for the changes presented.
Chapter 5: Response to Partial Deforestation

From the discussion in 1.5.2, the results of Costa et al. (2007) and Lejeune et al. (2014) raise several questions that will be addressed in this chapter: 1) How linear is the local deforestation response in fields in addition to precipitation? 2) Is the non-local response linear or non-linear? 3) Can we quantify the degree of non-linear response in a useful manner? Using a realistic heterogeneous crop distribution of corn, soybean, sugarcane, irrigated rice and cotton as replacement vegetation, we approach the question of the linearity of the response to Amazon deforestation by comparing experiments with 50% deforestation to a case with total deforestation. If 50% deforestation were to produce significantly more or less than 50% of the total deforestation signal, a non-linear response would be indicated, suggesting feedback processes in the climate system were amplifying or dampening the response.

This section aims to better quantify the linearity of the climate system response. Of particular interest is how does the response change when replacing half of the Amazon region and is deforestation a linear process (i.e. does half deforestation provide half of the total deforestation response). This section will investigate the varying deforestation responses due to partial deforestation and analyzes the climate response to differing spatial patterns of deforestation. Regions providing a non-linear response are discussed and to what extent differing spatial distributions of deforestation impact remote regions. This section will provide a metric that quantifies the linearity of the deforestation responses over a spatial domain to determine the climate sensitivity to spatial patterns of deforestation.

This chapter is the basis of Badger and Dirmeyer (2015b), submitted to Journal of Climate.
5.1 Methods of partial deforestation

Three partial deforestation simulations were completed: eastern deforestation (E.DEF), western deforestation (W.DEF) and half deforestation (DEF.50). As stated earlier, in the E.DEF and W.DEF simulations, the Amazon is divided in half to provide nearly equal deforestation areas (Figure 5.1). 62°W is the halfway line based on annual leaf area index (LAI, i.e. the area integral of annual mean LAI in the Amazon region to the east and west of 62°W is nearly the same). The columns of gridboxes immediately east and west of 62°W were used as a transition zone. Depending on the portion being deforested, these gridboxes were deforested 66% and 33% to smooth the transition. In the DEF.50 simulation, 50% of the present PFTs are removed from each gridbox and replaced with the tropical crop distribution (Figure 2.4).

5.2 A context for linearity

The response in partial deforestation simulations (DEF.50, E.DEF and W.DEF) will differ; the question is what information can be gathered from their comparisons to CON and DEF. For any climate variable under consideration, locally or remotely, we define the following responses:
\( DEF - CON = \Delta DEF \)  

(5.1)

\( E.DEF - CON = \Delta E.DEF \)  

(5.2)

\( W.DEF - CON = \Delta W.DEF \)  

(5.3)

\( DEF.50 - CON = \Delta DEF.50 \)  

(5.4)

If the response to deforestation is linear, the following will be true:

\[ \Delta W.DEF + \Delta E.DEF = \Delta DEF \]  

(5.5)

\[ 2 \times \Delta DEF.50 = \Delta DEF \]  

(5.6)

If eqns. 5.5 and 5.6 are not true, the response to Amazon deforestation is not linear. Realistically, eqns. 5.5 and 5.6 will not be exactly satisfied due to sampling error and natural variability within the Earth system model; a statistical test can be applied to determine whether they are significantly different, which would indicate non-linearity. When subtracting \( \Delta DEF \) from the partial anomalies in eqns. 5.5 and 5.6, when \( \Delta DEF \) is positive (negative) a positive (negative) significant difference would indicate a saturating response where more than half of the response to complete deforestation is already realized with half deforestation. A significant negative (positive) difference would indicate an accelerating response where the degree of change speeds up between half and total deforestation. So-called tipping point responses of climate to land use change would manifest as a saturating response if the tipping point is relatively close to the current land cover state, and as the accelerating response if the tipping point is closer to complete deforestation. Statistical
measures used in this chapter will be introduced as they are discussed in context with the results.

### 5.3 Results

All analyses in this section are within the latitudinal limits of 45°S and 45°N, as these were the limits to which most remote responses were seen. Furthermore, at higher latitudes internal variability in the climate system associated with baroclinic waves produces statistically significant but highly transient responses that are likely meaningless in the context of this study. Four variables are analyzed: temperature, precipitation, mid-troposphere geopotential height and zonal wind at the lowest model level.

#### 5.3.1 Spatial distribution of non-linearities

To determine the spatial pattern of significant non-linearities, a Student’s t-test has been used. The first step is to remove the climatological annual cycle of CON from all the deforestation simulations, leaving a time-series of anomalies relative to CON. The null hypotheses quantified in eqns. 5.5 and 5.6 are tested; with rejection indicating the progressive deforestation response does not behave linearly. The denominator of the test statistic uses a pooled variance from DEF and the respective partial deforestation case(s) being tested with a coefficient of $\sqrt{3}$. The coefficient change is due to three samples being used in the numerator of the test statistic, whereas traditional t-test’s use two samples in the numerator and have a coefficient of $\sqrt{2}$ for the denominator.

When comparing the spatial patterns of temperature change (Figure 5.2), three areas of change appear prevalent; the local response over the Amazon region, and remote responses over the equatorial Atlantic and Africa. The local response in DEF.50 shows more than half of the temperature response from DEF, indicating a saturating response as defined above. In contrast, the summed response $\Delta W$.DEF and $\Delta E$.DEF shows a more linear local response with non-linearities arising near the transition region (Figure 5.1). The changes
for both scenarios shows a saturating response, but as a cooling. The Atlantic region shows a prominent region in both scenarios that is an accelerating response.

Precipitation (Figure 5.3) largely has a linear response annually and seasonally for both scenarios. Regions of non-linearity appear near the rising branch of the Hadley circulation. It appears the DEF.50 response has more non-linear features, but not to the extent that temperature (Figure 5.2) exhibits.

Geopotential height and winds give an indication of the large-scale dynamical response of the atmosphere. In analyzing geopotential height (Figure 5.4), it is seen that 50% deforestation provides a large area of non-linear response, both locally in the Amazon region and remotely over the Atlantic, Indian and western Pacific oceans. These appear in all seasons. In JJA and SON, a non-linear response is found over a majority of central Africa. When combining the anomalies for ΔW.DEF and ΔE.DEF, a significant non-linear response extends around the equatorial region annually and in all seasons. Both scenarios primarily exhibit changes indicative of an accelerating response.

Zonal wind (Figure 5.5) has two primary areas where non-linearities occur; locally and over Africa. Over the Amazon, DEF.50 does not produce half of the wind change seen in DEF, indicating a local accelerating response. The same can be said about the summed response from ΔW.DEF and ΔE.DEF. While DEF produces a significant change in zonal winds over Africa in JJA and SON, a non-linear response appears in nearly all season for the partial deforestation scenarios. It is of note that the partial deforestation scenarios show a dipole response over Africa, while DEF provides a homogeneous response, indicating partial deforestation may provide either saturating or accelerating responses over the region.

5.3.2 Degree of non-linearity

We propose a quantitative method to estimate the degree of non-linearity (DNL) in the response of climate variables to deforestation. By the definition of collinearity, if the area of a triangle defined by vertices at three points has an area of zero, then the three points lie along the same line and are completely linear (Figure 5.6). If the triangle has an area greater
Figure 5.2: $\Delta$DEF (left column), $2^*\Delta$DEF.50 - $\Delta$DEF (middle column) and ($\Delta$W.DEF + $\Delta$E.DEF) - $\Delta$DEF (right column) for temperature [K]. Significant differences at the 95% level are shaded and implies non-linear response.
Figure 5.3: ΔDEF (left column), 2*ΔDEF.50 - ΔDEF (middle column) and (ΔW.DEF + ΔE.DEF) - ΔDEF (right column) for precipitation [mm/day]. Significant differences at the 95% level are shaded and implies non-linear response.
Figure 5.4: $\Delta$DEF (left column), $2^*\Delta$DEF.50 - $\Delta$DEF (middle column) and ($\Delta$W.DEF + $\Delta$E.DEF) - $\Delta$DEF (right column) for geopotential height [m]. Significant differences at the 95% level are shaded and implies non-linear response.
Figure 5.5: $\Delta$DEF (left column), $2\cdot\Delta$DEF.50 - DEF (middle column) and $(\Delta$W.DEF + $\Delta$E.DEF) - $\Delta$DEF (right column) for zonal wind [m/s]. Significant differences at the 95% level are shaded and implies non-linear response.
than zero, then the three points are not on the same line, and are therefore non-linear to a certain degree.

Using this concept, a metric has been developed to quantify a measure of non-linearity at each point in the domain (45°S - 45°N). The degree of deforestation is known for all three simulations and can be used as the abscissa to define triangles; CON=0; DEF.50, E.DEF, W.DEF=0.5 and DEF = 1. The corresponding ordinate values are the means (annual or seasonal) of the climate variable for the respective simulation. At each gridbox a triangle can be defined by the three vertices: (0,CON), (0.5,DEFX) and (1,DEF) and are seen as points A, D and C, respectively in Figure 5.6; where DEFX corresponds to one of the partial deforestation simulations. A second triangle is defined for each gridbox, the reference triangle, with the points (0,CON), (0.5,CON + DEF + σ) and (1,DEF); corresponding to points A, B+σ and C in 5.6. The partial deforestation is replaced in the reference triangle by the average of CON and DEF (which alone would give an area of zero) with the addition of one standard deviation based on the pooled variance of CON and DEF for the respective annual or seasonal means (Figure 5.6). The reference triangle defines a confidence level for determining linearity.

Summing the area of all the real triangles would quantify the total non-linear area and
summing the area of all the reference triangles would quantify an expected degree of non-linearity in an imperfectly sampled linear system. By normalizing the real triangle sum by the reference triangle sum, a single number can be given to state the DNL for the entire domain. The smaller the ratio, the closer the system is to being perfectly linear, providing a relative measure.

This method can be taken one step further to determine whether the dominant source of non-linearity is from an accelerating response or a saturating response (Figure 5.6). By binning the real triangles into either an accelerating or saturating response, and signing the areas of the former as positive and the latter as negative, the sum of the signed triangle areas quantify the dominant response, called the binned DNL. The binned sum is normalized by the sum of the reference triangle; a positive value would denote more of the non-linearities arise from an accelerating climate response to deforestation and a negative value would indicate non-linearities arise more from a saturating response.

To determine significance of the DNL and binned DNL, a bootstrap method has been utilized. Two sets of 125 years have been selected with replacement from each CON and DEF. One sample (Sample 1) of CON and DEF corresponds to the same years used for the end points above (i.e. (0,CON) and (1,DEF); A and C in 5.6). The remaining CON and DEF random samples (Sample 2) are averaged together to provide a half response. Real and binned triangles are calculated as previously described using the average of averaged CON and DEF from Sample 2 (i.e. D in 5.6) with the CON and DEF from Sample 1. The reference triangles are calculated as previously described using the CON and DEF from the Sample 1 (i.e. B+σ in 5.6). A distribution of 1000 values for DNL and binned DNL are collected to determine non-parametrically the percentiles of the half-deforestation cases. A DNL lying above the 95th percentile or a binned DNL either below the 2.5th or above the 97.5th percentile would indicate a significantly non-linear response at the 95% confidence level.

Table 5.1 displays the DNL and binned DNL for each variable, season and partial scenario as well as their associated percentiles. The DNL for temperature is significant annually
Table 5.1: Degree of non-linearity (DNL) and binned DNL annually and seasonally for temperature, precipitation, geopotential height and zonal wind in all partial deforestation scenarios. Italic refers to significant at the 95% level using the bootstrap method described in the text.

<table>
<thead>
<tr>
<th></th>
<th>Half DNL</th>
<th>Binned DNL</th>
<th>West DNL</th>
<th>Binned DNL</th>
<th>East DNL</th>
<th>Binned DNL</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Temperature</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Annual</td>
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<td>0.0369</td>
</tr>
<tr>
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<td>0.0331</td>
<td>0.1304</td>
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<td>0.1210</td>
<td>0.0326</td>
</tr>
<tr>
<td>MAM</td>
<td>0.1523</td>
<td>0.0395</td>
<td>0.1633</td>
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<td>0.1478</td>
<td>0.0221</td>
</tr>
<tr>
<td>JJA</td>
<td>0.1392</td>
<td>0.0428</td>
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<td>0.1612</td>
<td>0.0377</td>
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<tr>
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<td>0.0603</td>
<td>0.1375</td>
<td>0.0273</td>
<td>0.1502</td>
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<th>Binned DNL</th>
<th>East DNL</th>
<th>Binned DNL</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Precipitation</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Annual</td>
<td>0.0557</td>
<td>0.0057</td>
<td>0.0620</td>
<td>0.0064</td>
<td>0.0689</td>
<td>-0.0049</td>
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<tr>
<td>DJF</td>
<td>0.0802</td>
<td>0.0073</td>
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<td>0.0012</td>
<td>0.1041</td>
<td>-0.0124</td>
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<tr>
<td>MAM</td>
<td>0.0868</td>
<td>0.0037</td>
<td>0.1060</td>
<td>0.0014</td>
<td>0.1086</td>
<td>0.0013</td>
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<tr>
<td>JJA</td>
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<td>0.0952</td>
<td>-0.0065</td>
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<tr>
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<td>0.0094</td>
<td>0.1066</td>
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<td>0.1063</td>
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<th>Binned DNL</th>
<th>East DNL</th>
<th>Binned DNL</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Geopotential Height</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Annual</td>
<td>0.4450</td>
<td>0.3438</td>
<td>0.3498</td>
<td>0.2036</td>
<td>0.5269</td>
<td>0.4039</td>
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<tr>
<td>DJF</td>
<td>0.1236</td>
<td>0.0747</td>
<td>0.1137</td>
<td>0.0466</td>
<td>0.1506</td>
<td>0.0967</td>
</tr>
<tr>
<td>MAM</td>
<td>0.1522</td>
<td>0.0937</td>
<td>0.1289</td>
<td>0.0623</td>
<td>0.1622</td>
<td>0.0996</td>
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<tr>
<td>JJA</td>
<td>0.1434</td>
<td>0.1029</td>
<td>0.1039</td>
<td>0.0443</td>
<td>0.1914</td>
<td>0.1258</td>
</tr>
<tr>
<td>SON</td>
<td>0.1491</td>
<td>0.1118</td>
<td>0.1253</td>
<td>0.0734</td>
<td>0.1627</td>
<td>0.1034</td>
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<table>
<thead>
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<th>Binned DNL</th>
<th>West DNL</th>
<th>Binned DNL</th>
<th>East DNL</th>
<th>Binned DNL</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Zonal Wind</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Annual</td>
<td>0.0614</td>
<td>0.0077</td>
<td>0.0669</td>
<td>0.0088</td>
<td>0.0711</td>
<td>-0.0011</td>
</tr>
<tr>
<td>DJF</td>
<td>0.0856</td>
<td>0.0160</td>
<td>0.1038</td>
<td>0.0133</td>
<td>0.1090</td>
<td>-0.0087</td>
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<tr>
<td>MAM</td>
<td>0.1039</td>
<td>-0.0120</td>
<td>0.1119</td>
<td>-0.0067</td>
<td>0.1076</td>
<td>0.0053</td>
</tr>
<tr>
<td>JJA</td>
<td>0.1119</td>
<td>-0.0033</td>
<td>0.1164</td>
<td>-0.0101</td>
<td>0.1145</td>
<td>0.0067</td>
</tr>
<tr>
<td>SON</td>
<td>0.1101</td>
<td>0.0502</td>
<td>0.1085</td>
<td>0.0397</td>
<td>0.1014</td>
<td>0.0113</td>
</tr>
</tbody>
</table>
and for all seasons across all simulations. The DNL for precipitation is significant annually and during all seasons for E.DEF and W.DEF. For DEF.50, the precipitation response is only significant in JJA and SON. Geopotential height is significant for DEF.50 and E.DEF annually and for all seasons. DEF.W has a significant DNL for annual, MAM and SON means. Zonal wind shows a significant DNL annually and seasonally across all simulations.

For binned DNL (Table 5.1), temperature has a significant response annually and for all seasons in the DEF.50 and E.DEF simulations corresponding to an accelerating response from partial to total deforestation. W.DEF has a significant accelerating response only for SON. Precipitation does not have any significant binned DNL values, suggesting that although there are many areas with non-linear responses, the accelerating and saturating responses largely balance out Geopotential height has a significant binned DNL annually and for all seasons across all simulations indicating an accelerated response. Zonal wind exhibits an accelerating response in SON for DEF.50 and W.DEF due to a significant binned DNL.

### 5.3.3 Spatial correlations

In order to determine how the pattern of change for each variable and scenario compares to the complete deforestation response, a spatial correlation was calculated over the domain of interest. In order to determine the significance threshold, the spatial degrees of freedom (DOF) between 45°S-45°N are estimated using the method proposed by Bretherton et al. (1999). This method requires summing the inverse squares of the explained variance for each EOF of the time-series (Bretherton et al., 1999, Eqn. 4). The method is applied to each individual simulation to obtain the DOF for each season and variable. For each variable across all simulations and in each season, the DOF were similar, so a mean DOF across all experiments was used to calculate significance. In calculating the correlations between ΔDEF and ΔDEF.50, and between ΔDEF and the sum of ΔE.DEF and ΔW.DEF, a significant result indicates a linear response cannot be rejected as a possibility.

Table 5.2 shows the spatial correlations for each scenario, variable and season, along
Table 5.2: Spatial correlations of $\Delta E.\text{DEF} + \Delta W.\text{DEF}$ and $\Delta \text{DEF}.50$ to $\Delta \text{DEF}$ and associated p-values for temperature, precipitation, geopotential height and zonal wind for each season. Degrees of freedom for each correlation in parentheses next to respective season.

<table>
<thead>
<tr>
<th>Variable</th>
<th>DJF (Cor. p-value)</th>
<th>MAM (Cor. p-value)</th>
<th>JJA (Cor. p-value)</th>
<th>SON (Cor. p-value)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Temperature</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>East+West</td>
<td>0.83 0.0001</td>
<td>0.86 0.0000</td>
<td>0.94 0.0000</td>
<td>0.90 0.0000</td>
</tr>
<tr>
<td>2*Half</td>
<td>0.69 0.0024</td>
<td>0.72 0.0002</td>
<td>0.87 0.0000</td>
<td>0.84 0.0001</td>
</tr>
<tr>
<td><strong>Precipitation</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>East+West</td>
<td>0.81 0.0006</td>
<td>0.85 0.0000</td>
<td>0.86 0.0000</td>
<td>0.86 0.0000</td>
</tr>
<tr>
<td>2*Half</td>
<td>0.73 0.0031</td>
<td>0.83 0.0000</td>
<td>0.72 0.0005</td>
<td>0.74 0.0006</td>
</tr>
<tr>
<td><strong>Geopotential Height</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>East+West</td>
<td>0.70 0.0192</td>
<td>0.75 0.0020</td>
<td>0.71 0.0037</td>
<td>0.81 0.0001</td>
</tr>
<tr>
<td>2*Half</td>
<td>0.61 0.0440</td>
<td>0.51 0.0430</td>
<td>0.50 0.0450</td>
<td>0.55 0.0147</td>
</tr>
<tr>
<td><strong>Zonal Wind</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>East+West</td>
<td>0.84 0.0022</td>
<td>0.84 0.0000</td>
<td>0.92 0.0000</td>
<td>0.91 0.0000</td>
</tr>
<tr>
<td>2*Half</td>
<td>0.74 0.0109</td>
<td>0.72 0.0002</td>
<td>0.75 0.0003</td>
<td>0.77 0.0003</td>
</tr>
</tbody>
</table>

with the associated p-values. A significant result at the 95% confidence level using single-tailed test (i.e. linear response), was found for all temperature, precipitation, geopotential height and zonal wind correlations. This indicates that the pattern of changes exhibited by the partial deforestation scenarios is not significantly different from a linear interpolation between control and total deforestation simulations.

### 5.4 Discussion

Using three methods to analyze the linearity of the climate system response to progressive deforestation of the Amazon in a state-of-the-art Earth system model has provided some notable results. By testing eqns. 5.5 and 5.6, it is seen that regions in the tropics respond differently for each variable and deforestation pattern. Locally over the Amazon region, deforesting 50% of each grid box provides large areas of with a non-linear response for
temperature, geopotential height and zonal wind. In contrast, by adding the anomalies of E.DEF and W.DEF, non-linearity is locally present for temperature and geopotential height.

When looking remotely, prominent regions of non-linear responses are seen over the Atlantic Ocean and Africa, implying that these are the most sensitive non-local regions to partial deforestation. One notable impact is when comparing partial deforestation changes to total deforestation for temperature over Africa. The Congo Basin is ringed by areas of non-linear response, while the basin itself is not. This implies that the Congo region is less sensitive to partial deforestation than the surrounding areas.

Another inference that can be made by looking at the impacts of partial deforestation is that different elements of the climate system respond differently. Temperature, geopotential height and zonal wind all have large areas that behave non-linearly. Temperature response appears non-linear locally, over the Atlantic and Africa, zonal wind appears non-linear locally and over Africa, while geopotential height appears non-linear across a majority of the tropics. Precipitation exhibits a predominantly linear response both locally and remotely.

Analyzing the spatial correlation (Table 5.2) between partial and total deforestation, it can be seen that all correlations are significant; meaning the pattern of the climate response to deforestation is robust regardless of the degree of deforestation. This result implies that remote regions that experience changes due to total deforestation are also the regions that experience changes to partial deforestation. Likewise, remote regions that are not sensitive to partial deforestation of the Amazon are also not sensitive to total deforestation.

The deviation from a linear response quantified across 45°S-45°N is significant for nearly all cases and variables, meaning the climate system does not necessarily respond linearly as deforestation progresses. Determining whether the non-linearity is due to an accelerating response as opposed to a saturating response can provide some real world insight about the sensitivity of the climate system.

The only variable to show a significant binned DNL with respect to all partial deforestation scenarios is geopotential height, showing an accelerating response. This should be of
particular interest as geopotential height can provide some insight into the large-scale
circulation. Having a significant binned DNL means that partial deforestation will not alter
the circulation by as much as half of the total deforestation response. Indicating that there
is a threshold of deforestation that the circulation would be sensitive to and cause large
changes across the domain.

The fact that temperature has an accelerating response annually and seasonally for
DEF.50 and E.DEF highlights an interesting sensitivity to partial deforestation. This indi-
cates that in regards to temperature, the climate system is sensitive to the spatial pattern
of deforestation, as W.DEF does not provide the same binned DNL results. This is un-
like the other variables tested that show a similar binned DNL for their respective partial
deforestation scenarios.

When comparing W.DEF and E.DEF to DEF, an interesting result arises for tempera-
ture (Figure 5.7). If W.DEF or E.DEF are significantly different from DEF in a region where
significant changes are found for eqn. 5.1 (shown in Figure 5.2), then it can be inferred
that the spatial pattern of deforestation does not cause that change. On the other hand,
if there is not a significant change between either W.DEF or E.DEF and DEF, then that
spatial pattern can explain that change. Along the coast of South America in the eastern
Pacific, significant changes in temperature are seen when comparing DEF to CON (Figure
5.2). In both W.DEF and E.DEF (Figure 5.7), the temperature change shown by DEF is
not significantly different from those partial deforestation scenarios, implying that region
is sensitive to Amazon deforestation. In contrast, the significant temperature changes near
the South American coast over the western Atlantic (Figure 5.2) cannot be explained by
either W.DEF or E.DEF (Figure 5.7). This result indicates that region is less sensitive to
those partial deforestation scenarios, but significant changes occur for total deforestation.
Regions that behave like the western Atlantic can be found for other variables (not shown)
and further justifies the findings of an accelerated response to deforestation.

Furthermore, when analyzing the percent change of sensible and latent heat flux over the
deforested region; DEF.50 provides 82.5% of the sensible heat flux reduction found in DEF
Figure 5.7: W.DEF DEF (left column) and E.DEF (right column) for temperature [K], shading indicates significance.
but only 45.5% of the latent heat flux reduction found in DEF. These differing changes as the surface further indicate that the response locally is not linear and can initiate a non-linear response to the climate system.

5.5 Conclusions

This section has provided insight into the global response to progressive deforestation of the Amazon by examining a case of total deforestation and three scenarios each representing deforestation of half of the Amazon: 50% deforestation in each gridbox, deforestation of the eastern Amazon and deforestation of the western Amazon.

Using a Students t-test to analyze eqns. 5.5 and 5.6, it was found that there are regions that respond non-linearly. A notable finding is that different variables provide different regions and degrees of response of non-linearity. Using a spatial correlation to determine how well the patterns of deforestation are in comparison to total deforestation, it was determined that all correlations were significant, meaning that spatially, the system behaves linearly.

This chapter provides a method to determine the magnitude that a system is non-linear. Using this method it has been determined that nearly all variables behaved non-linearly. Taking this method a step further to determine whether the non-linearity was due to a saturation response or accelerating response, it was found that a majority of the non-linearity is occurring due to the system being an accelerating response.

This section has highlighted what regions are sensitive to deforestation as well as whether they respond linearly. With a majority of the response in this study showing an accelerating response, this implies that a threshold for sensitivity to deforestation exists. In comparison to the real world, this means that the global impacts of Amazon deforestation are not being felt yet, but once the threshold for sensitivity is crossed, then global implications could become apparent.

An investigation into the degree of non-linearity in the response of climate to land use change should be expanded to other models as well. Knowing this sensitivity will provide insight into how resilient the climate system is to large-scale land-use change and the regions
most impacted by such.
6.1 Summary

Previous modeling studies have detailed the local impacts of Amazon deforestation, with few showing notable remote responses. Part of the reason for this could be due to model setup; either through the use of homogeneous replacement vegetation or having a non-interactive ocean model.

This study has furthered the research regarding modeling of Amazon deforestation in Earth system models by using a heterogeneous crop distribution to provide a more realistic land-surface for Amazon deforestation. In addition, this study used a fully-coupled climate model with and interactive ocean, helping to further characterize the impacts of Amazon deforestation.

To help provide a more realistic response to Amazon deforestation, a suite of tropical crops were developed for use in the Community Land Model version 4.5. Soybean, corn, sugarcane, rice and cotton were developed and parameterized for the tropics, incorporated into CLM4.5 and shared with the modeling community. These tropic crops are being added to CLM for the next public release, as well as having already been used by others in the community.

In regards to temperature, precipitation and surface fluxes; the local response to deforestation in this study falls in line with the results of previous studies. Detailing land-atmosphere interactions, it was determined that the land-surface becomes more strongly coupled when densely forested areas are removed. In contrast, it was found that through irrigating the soil, the land-surface becomes more weakly coupled to the atmosphere.

By altering the land-atmosphere interactions over such a large region, modifications to the atmosphere above the region were found. With cooling in the upper atmosphere...
above the deforested region, a thermally induced circulation response occurs. This cir-
culation response in the upper atmosphere is rather remarkable in the sense that it is in
agreement with circulation response to tropical heating proposed by (Gill, 1980). Addi-
tionally, modifications to the meridional circulation, particularly the rising branch of the
Hadley circulation. These modifications to the large-scale circulation were found to impact
remote regions, such as the South Atlantic Convergence Zone, west and central Africa, and
southwest North America.

Through the use modeling three different partial deforestation simulations, the linear
response to Amazon deforestation was analyzed. Few studies have analyzed the extent
to which the response to partial Amazon deforestation is non-linear, with respect to total
deforestation. It was shown that spatially, areas of non-linearity arise to due to partial
deforestation scenarios. A metric to quantify the degree of non-linearity over a spatial
domain was proposed and used in this study. Through the use of this new metric, it was
determined that the magnitude of the response of variables analyzed behaved non-linearly
with regards to partial deforestation, although the spatial pattern of the response varied
little. It was also determined that half deforestation provides less than half of the response
to total deforestation. This result shows that there climate system has some resiliency to
large-scale land-use change, but that there is a tipping point to which the climate system
is sensitive to Amazon deforestation.

6.2 Societal Implications

The modification to the physical climate system due to Amazon deforestation can have
socio-economic impacts both locally and globally. Spracklen et al. (2012) notes that re-
duced precipitation due to Amazon deforestation will have consequences on rainfall-reliant
industries in the region. These industries include, but are not limited to, agriculture and
hydroelectric power generation, both of which play a substantial role in South American
economies. Brazil is one of the largest producers hydroelectric power in the world, reduc-
tions to precipitation across the region would directly impact the Brazilian economy.
Agriculture would be impacted both locally and remotely. Regions locally (Figure 4.6) and remotely (Figure 4.8) that are important for crop production show significant precipitation decreases. The crop type most largely impacted would be cereal grains, which are integral for feeding developing nations, particularly in Africa.

By analyzing the linearity of the response to partial deforestation, it is likely that remote regions will not entirely feel the impacts of Amazon deforestation for some time, perhaps decades. Locally, the impacts have been felt and have not gone unnoticed as policy initiatives have slowed the present rate of Amazon deforestation (Nepstad et al., 2014).

6.3 Future Work

The atmospheric model used in this study does not represent the large changes in atmospheric carbon dioxide that would accompany Amazon deforestation. Foley et al. (2005) notes that LUC can have an impact on the global carbon cycle. Nepstad et al. (2008) adds that the trees of the Amazon region contains 90-140 billion tons of carbon, equivalent to 9-14 decades of current anthropogenic carbon emissions. Amazon deforestation’s impact could have large consequences on the global climate system that were not realized in this study. Further work should be done that takes into account that impact to the global carbon cycle and could further enhance regions of remote change.

A notable impact found, but not discussed, was modifications to ENSO. By removing the seasonal cycle and normalizing by the control simulations standard deviation for areal averaged SST in the NINO3.4 region, an ENSO index for each simulation was created and analyzed. Although significant changes to the power spectrum of ENSO were not found, changes in the extremes of ENSO were present (Table 6.1).

Using the 5th and 95th percentiles for the control to determine thresholds for warm and cold ENSO events, large decreases in the number of cold events for eastern and western deforestation simulations occurred. Modifications to ENSO would have substantial impacts on the climate system and is an area that should be research further.

All the results in this study are model dependent and could vary across different climate
Table 6.1: Counts of warm and cold events for ENSO. Warm events defined as being greater than or equal to the 95th percentile in the control simulation, and cold events are defined as being less than or equal to the 5th percentile for the control simulation.

<table>
<thead>
<tr>
<th></th>
<th>Annual</th>
<th>DJF</th>
<th>MAM</th>
<th>JJA</th>
<th>SON</th>
</tr>
</thead>
<tbody>
<tr>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Control</td>
<td>75</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>7</td>
</tr>
<tr>
<td>Deforested</td>
<td>46</td>
<td>3</td>
<td>4</td>
<td>8</td>
<td>4</td>
</tr>
<tr>
<td>Half</td>
<td>72</td>
<td>5</td>
<td>6</td>
<td>8</td>
<td>7</td>
</tr>
<tr>
<td>Eastern</td>
<td>87</td>
<td>9</td>
<td>5</td>
<td>6</td>
<td>9</td>
</tr>
<tr>
<td>Western</td>
<td>55</td>
<td>2</td>
<td>5</td>
<td>8</td>
<td>4</td>
</tr>
<tr>
<td><strong>COLD EVENTS</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Control</td>
<td>75</td>
<td>7</td>
<td>7</td>
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models. Analyzing the global impacts to Amazon deforestation with other models would help in the understanding of how Amazon deforestation impacts the local climate and how local changes feed into the climate system. Through the use of other model, areas of agreement for remote impacts and areas that are sensitive to Amazon deforestation may become more apparent and help to provide further insight as to how the land-surface and climate interact.

In closing, Amazon deforestation is not just a local problem, but also an issue that can have far-reaching implications around the world. Amazon deforestation is not just a progressive problem where the impacts will be felt as changes to the region occur, there is a feedback that shows climate sensitivity to the modification to the land-surface.
Bibliography


Henderson-Sellers, A., R. E. Dickinson, T. B. Durbidge, P. J. Kennedy, K. McGuffie,


Andrew M. Badger grew up in Asheville, NC. He attended North Carolina State University, where he received his Bachelor of Science in Meteorology in 2010. During his time as an undergraduate, he twice interned with the National Aeronautics and Space Administration (NASA) at Marshall Space Flight Center in Huntsville, AL researching the atmosphere of Mars for entry, descent and landing purposes. In 2010, Andrew joined the Climate Dynamics program at George Mason University.

As part of his Ph.D. studies at George Mason University, Andrew worked on understanding the role that large-scale land-use change has on the local and global climate. In addition, Andrew worked with researchers at the National Center for Atmospheric Research (NCAR) to make improvements to the Community Land Model version 4.5, which are being incorporated into next public release of the Community Land Model. This work has been presented at the Land Model Working Group Meeting hosted by NCAR, the American Geophysical Union Fall Meeting and has been submitted to three peer-reviewed journals: Hydrology and Earth System Sciences, Climate Dynamics and Journal of Climate.

Andrew is interested in advancing the understanding of the relationship between the land-surface and climate system, investigating how the land-surface changes in a changing climate, and exploring the societal impact these changes can have.