

Scale dependence of the simulated impact of Amazonian deforestation on regional climate

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**Abstract**

Using a global climate model, Amazonian deforestation experiments are conducted perturbing 1, 9, 25, 81 and 121 grid points, each with 5 ensemble members. All experiments show warming and drying over Amazonia. The impact of deforestation on temperature, averaged either over the affected area or a wider area, decreases by a factor of two as the scale of the perturbation increases from 1 to 121 grid points. This is associated with changes in the surface energy balance and consequential impacts on the atmosphere above the regions deforested. For precipitation, as the scale of deforestation increases from 9 to 121 grid points, the reduction in rainfall over the perturbed area decreases from ~ 1.5 to ~ 1 mm d^{-1} . However, if the surrounding area is considered and large deforestation perturbations made, compensatory increases in precipitation occur such that there is little net change. This is largely associated with changes in horizontal advection of moisture. Disagreements between climate model experiments on how Amazonian deforestation affects precipitation and temperature are, at least in part, due to the spatial scale of the region deforested, differences in the areas used to calculate averages and whether areas surrounding deforestation are included in the overall averages.

1. Introduction

The future of tropical forests is uncertain as a consequence of climate change, and as a consequence of direct deforestation by humans (Bonan 2008). Tropical deforestation can affect local and regional climate in a variety of ways (D'Almeida *et al* 2007, Pielke *et al* 2011) including via changes in biogeophysical characteristics that affect the surface energy and water balance (Boisier *et al* 2012), and via changes in surface roughness that can affect turbulent transfer of sensible (SH) and latent heat (LH) and horizontal advection of moisture (Polcher and Laval 1994a, Baidya Roy 2009).

Tropical deforestation experiments using climate models almost always simulate warming and drying (Nobre *et al* 1991, Zhang *et al* 1996, Badger and Dirmeyer 2016). The replacement of tropical forests with grasslands or crops typically increases the albedo and reduces the net radiation (R_{net}), which tends to cool the surface. This is offset by reduced evapotranspiration from the more shallow rooted, lower leaf area and

aerodynamically smoother grasslands (Pielke *et al* 2011, Boisier *et al* 2012). The reduced evapotranspiration, which is a very common result in tropical deforestation simulations, warms the surface despite the reduction in R_{net} , but how much the land warms varies greatly across different experiments conducted with climate models and estimates typically range from 1 °C to 3 °C (Badger and Dirmeyer 2016).

While the sign of the change in temperature is reasonably consistent between climate model simulations, the simulated changes in precipitation are much more variable. There are at least two counteracting feedbacks. First, Costa and Foley (2000) noted that lower precipitation recycling following tropical deforestation altered the boundary layer and tended to reduce the likelihood of precipitation. This implies a positive feedback (Runyan *et al* 2012) where less forest leads to less LH and less precipitation. The strength of this feedback depends on the magnitude of local precipitation recycling in the areas of tropical deforestation, which tends to be high (around 25%–35%) in the Amazon (Eltahir and Bras 1994). A second feedback is

linked with moisture flux convergence whereby deforestation reduces the aerodynamic roughness of the surface, which enables stronger horizontal advection. This draws more moisture into a region and has the potential to increase precipitation. The overall balance of these contributors to precipitation change is highly model dependent and consequently some deforestation experiments suggest reduced precipitation (e.g. Costa and Foley 2000, Badger and Dirmeyer 2016, Lejeune *et al* 2015) while others suggest increases (Dirmeyer and Shukla 1994, Polcher and Laval 1994b). Another factor that has less commonly been examined is the role of surface heterogeneity in affecting the response of the atmosphere to deforestation; this likely plays a role in explaining the varying precipitation result (Nobre *et al* 2009, Guillod *et al* 2015).

Whether precipitation increases or decreases as a result of deforestation is very likely linked with the spatial scale of deforestation imposed in the model. For example, beyond some critical threshold of tropical deforestation, substantial and irreversible reductions in precipitation has been predicted (Runyan *et al* 2012, Lawrence and Vandecar 2015). Scheffer *et al* (2005) suggest a critical limit of 30%–40% of forest removal would permanently change a wet forest to a dry savanna. At the opposite extreme, observations over the Amazon point to small-scale deforestation increasing precipitation (Negri *et al* 2004, Chagnon and Bras 2005). Runyan *et al* (2012) note that heterogeneous small-scale deforestation over scales of tens of kilometers might increase vertical instability and increase precipitation, a statement supported by mesoscale modeling (Avissar *et al* 2002, Baidya Roy 2009). Importantly, this increase in precipitation was linked to higher temperatures over the deforested region, which led to increased horizontal pressure gradients and an increased convergence into the areas of deforestation of cool, moist air. This negative feedback, whereby deforestation leads to an increase in precipitation, is in sharp contrast with the positive feedback dominant in most climate models. Overall, the impact of deforestation on precipitation is uncertain and D'Almeida *et al* (2006) concludes that 'intrinsic and interrelated scale and heterogeneity dependencies on the impact of deforestation in Amazonia on the hydrological cycle exist'.

The implementation of a deforestation experiment in a climate model varies widely. Some investigators convert the whole region (say Amazonia) from tropical forest to grass or crops while others perturb much more limited areas. It also likely matters where the deforestation is implemented over Amazonia because the impact of deforestation is linked with the intensity of the simulated precipitation recycling (Runyan *et al* 2012). In addition, an important property of a climate model is how strongly the land is coupled to the atmosphere and a small land cover change in a strongly coupled region can have a larger impact on the atmosphere than a large change in a weakly

coupled region (Lorenz and Pitman 2014). Further, the biogeophysical parameters used in a climate model to represent the forest and the grass or crops can vary and hence 'deforestation' can mean different changes in leaf area index, aerodynamic roughness length, root depths, albedo etc depending on the model used. Even if the changes in parameters use observations, problems with the scale dependency of these observations exist when applied to the wide range of climate model resolutions used in deforestation experiments (see table 1, D'Almeida *et al* 2007). Different experiments also use different vertical resolutions and different parameterizations of non-perturbed land surface components including soil moisture and runoff.

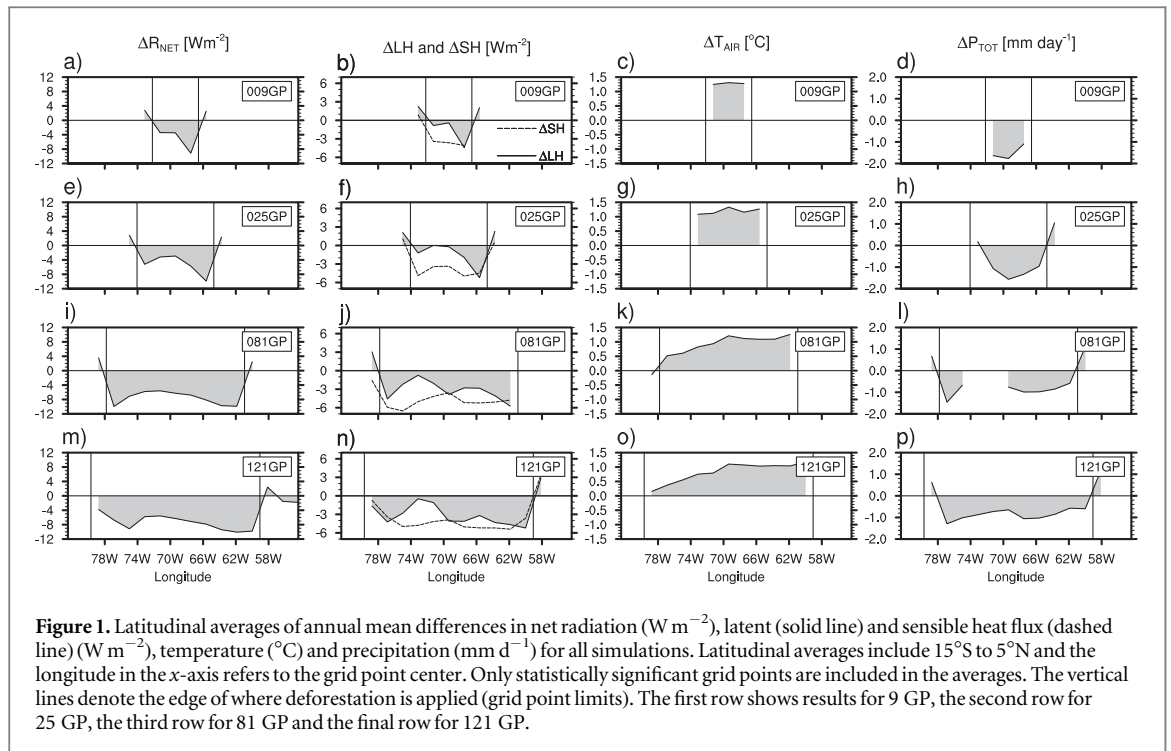
While all of these uncertainties are potentially important, one major area that has not been thoroughly investigated is how the spatial scale of Amazonian deforestation imposed in a model might explain why previous results vary. Here, we examine the relationship between the scale of deforestation implemented in a climate model and the local and regional impact on temperature and precipitation. Our results highlight some interesting scale-dependent features due to deforestation, which help contextualize results from previous studies.

2. Methods and data

We used the Australian Community Climate Earth System Simulator version 1.3b in an Atmospheric Model Inter-comparison Project-style (Gates 1992) configuration with prescribed sea surface temperatures and sea ice based on observations. The resolution is 1.25° latitude by 1.875° longitude with 38 levels in the vertical and we use a 30 min time step. The control simulation was very similar to the CMIP5 contribution, except the land surface model CABLE was updated to version 2.0.1.

In each experiment, and in common with many previous studies, we changed all forest plant functional types (principally evergreen broadleaf forest) to grassland. This affected several biophysical parameters and included an increase in albedo from ~ 0.12 for the forest to 0.16 for the grass and a decrease in the canopy height from ~ 35 to ~ 0.57 m.

In our first experiment, we perturbed a single 1.25° latitude by 1.875° longitude grid point (an area of $28\,051\text{ km}^2$, hereafter 1 GP). We then undertook additional experiments with incrementally expanded deforested area, first perturbing 9 (9 GP, $248\,774\text{ km}^2$), then 25 (25 GP, $693\,735\text{ km}^2$), 81 (81 GP, $2215\,025\text{ km}^2$), and 121 (121 GP, $3106\,963\text{ km}^2$) grid points. The perturbation for the 1 GP experiment was in the center of the region deforested, and larger changes radiated out from this central point. We ran 5 ensemble members for our control experiment (CTL) and for each perturbation experiment starting each ensemble member one year apart. The shortest simulation covered the period



1977–2012 and the longest was for 1973–2012. This enabled us to always have 5 ensemble members for a 32 year period (1981–2012) and to omit at least 4 years of each simulation as a spin-up period.

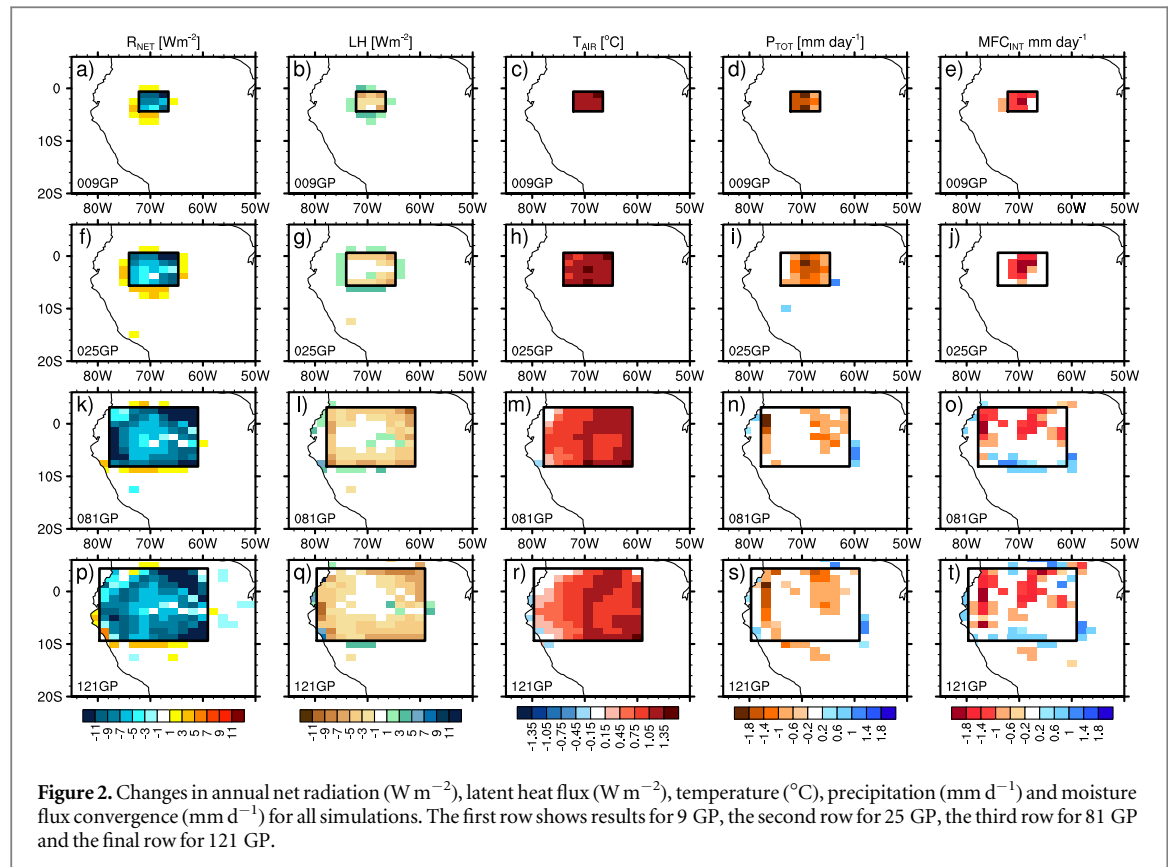
The results are presented as 32 year annual averages over the five ensemble members (so averages are over 160 years) but we also show the ensemble spread in area averaged quantities. We use the modified t-test based on (Zwiers and von Storch 1995) and the False Discovery Rate (Wilks 2006) for statistical significance testing (see Lorenz *et al* 2016 for details of these tests). The modified t-test takes into account autocorrelation in time by adjusting the sample size to the equivalent number of independent samples. If the data is not autocorrelated it reduces to the standard Student t-test. The False Discovery Rate is used as a field significance test to account for the fact that we apply the modified t-test multiple times in space (at every grid point) and correlation in space. Applying a statistical test at every grid point will lead to some statistically significant grid points by chance (the significance level times the number of grid points). The False Discovery Rate identifies the locally significant tests by controlling the expected proportion of rejected local null hypotheses that are actually true. The test is performed on the 32 year time series from each ensemble member and we test if the difference between the experiment and control runs are statistically significant different from zero (one-sample test). We perform the analysis on individual ensemble members first and then calculate ensemble means. Our final results therefore only include data from grid points where differences were statistically significant in the individual ensemble members, and results are only included over land.

3. Results

3.1. Local effects from deforestation and scaling

The change from forest to grassland results in an increase in albedo that in turn decreases R_{net} (figures 1 and 2 first column) in the areas where deforestation is implemented. The decrease in R_{net} , coupled with the reduced capacity of the grassland to evapotranspire, leads to a decrease in LH (figures 1 and 2 second column). SH (figure 1, second column, dashed line) also decreases over the region of deforestation due to the reduction in R_{net} . The air temperature is statistically significantly increased by around 1°C (figures 1(c), (g), (k), (o) and figure 2 third column) and precipitation is decreased in the area deforested by up to 0.8 mm d^{-1} (figures 1(d), (h), (l) and (p)). Hence, in our experiments, Amazonian deforestation leads to warmer and dryer conditions in the area where the deforestation occurs, a result consistent with virtually all earlier studies (e.g. D’Almeida *et al* 2007, Lawrence and Vandecar 2015).

For R_{net} , LH and SH the spatial extent and the magnitude of the decrease due to deforestation scales weakly with the increase in the scale of the perturbation. For small amounts of deforestation (figures 1(a) and 2(a)) decreases of $\sim 4\text{--}8 \text{ W m}^{-2}$ are simulated for R_{net} . Decreases of up to $\sim 5 \text{ W m}^{-2}$ are simulated for LH (figures 1(b) and 2(b)) and SH (figure 1(b)) where deforestation is implemented. Note how the change in LH is larger in the eastern half of the region deforested (e.g. figure 1(b)) following the pattern of change in R_{net} (figure 1(a)). This contrasts with the change in SH which is more spatially consistent. As the area of deforestation expands the area where R_{net} is reduced



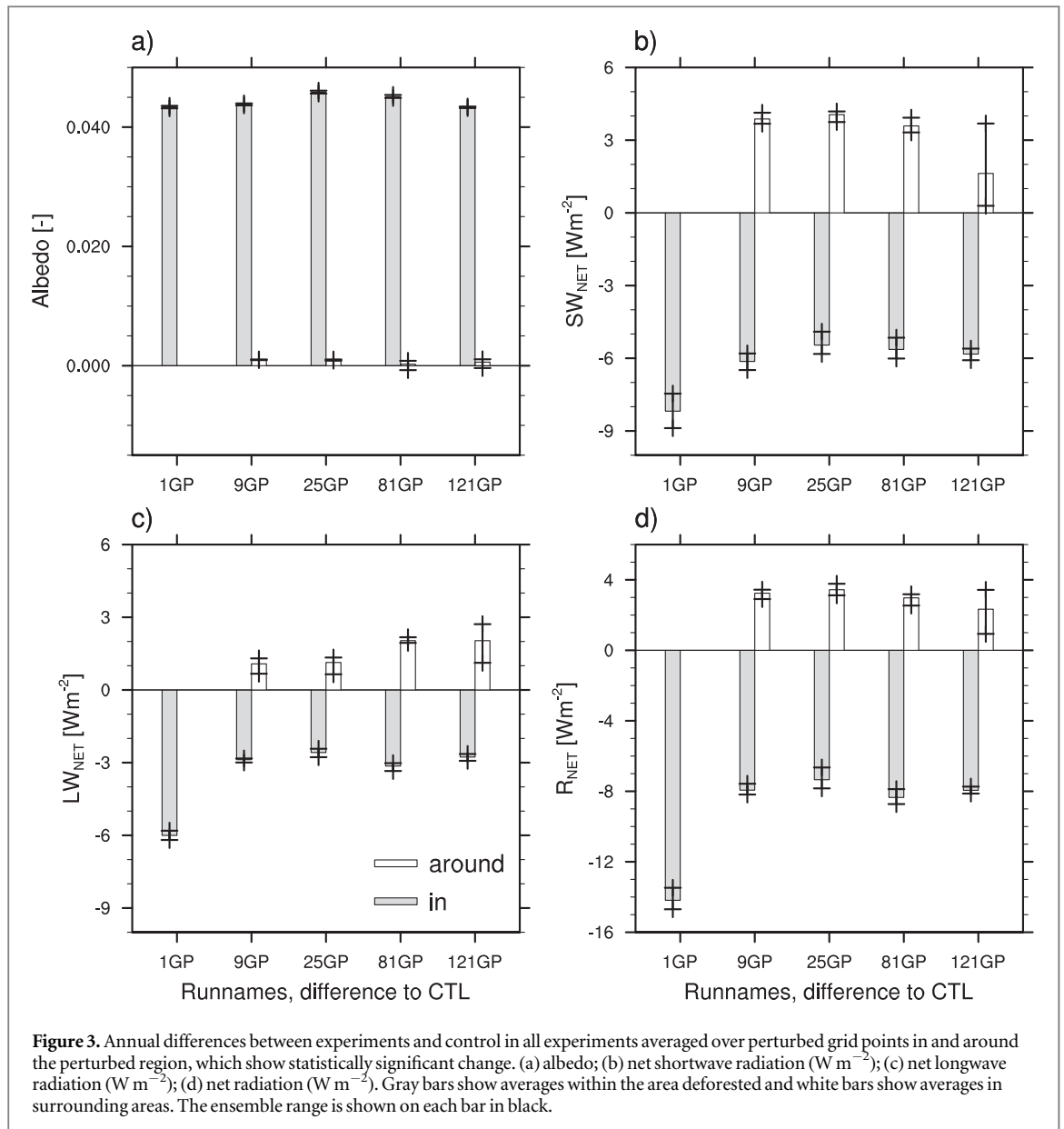
also expands (figures 1(e), (i), (m) and figures 2(f), (k), (p)), but the magnitude of the reduction is similar. For temperature, a small amount of deforestation (e.g. 9 GP, figures 1(c) and 2(c)) causes warming of $\sim 1.0^{\circ}\text{C}$ – 1.5°C locally and as the area deforested increases, the magnitude of the increase reduces (the area that warms does of course increase). At the largest scales examined here heterogeneous patterns of temperature change emerge (e.g. 81 GP, figure 2(m), 121 GP, figure 2(r)). These reflect the detailed nature of the response by the atmosphere to the land perturbation. Finally for precipitation, reductions of ~ 1.0 – 1.5 mm d^{-1} occur for small-scale perturbations (9 GP, figures 1(d), 2(d) and 25 GP, figures 1(h), 2(i)) reflecting a statistically significant decrease in precipitation for many grid points within the deforested region. However, as the scale of deforestation increases (81 GP, figures 1(l), 2(n) and 121 GP, figures 1(p), 2(s)) the precipitation is reduced by smaller amounts and these reductions are not always statistically significant (e.g. 81 GP, figure 1(l)). In addition, the area affected as a proportion of the deforested domain is smaller. These two combine such that the latitude averaged (figure 1) reduction in precipitation is smaller (~ 0.75 – 1.0 mm d^{-1}) at higher amounts of deforestation.

3.2. Effects surrounding the deforested area

When averaging results from deforestation experiments, there is no agreed protocol to determine the

area to use for calculating the average as there is observational evidence of scale dependent impacts on air temperature (e.g. von Randow *et al* 2004) and precipitation (Spracklen *et al* 2012) in areas proximal to deforestation. Our results show this is problematic. Figures 1 and 2 show an interesting phenomenon immediately surrounding the area of deforestation, where the vegetation has not been changed. Whether or not a grid point is deforested in the model is binary; as is common in deforestation experiments a grid square is usually either deforested or it is not. This leads to an edge effect or discontinuity in the model. Figure 2 shows for R_{net} , LH and precipitation that the sign of the change in each quantity reverses in many of the grid squares immediately surrounding the regions of deforestation. This reverse occurs on the eastern edge of the regions deforested for precipitation in the 25 GP experiment (figure 1(h)) and for both the eastern and western edges for the 81 GP and 121 GP experiments (figures 1(l) and (p)).

This edge phenomenon is particularly clear in figure 2. Both R_{net} and LH increase for a distance of several grid points surrounding the area of deforestation. For precipitation, the impact is quite small, but is statistically significant for 2 grid points in figure 2(i) (25 GP) and for more grid points in figure 2(n) (81 GP) and (s) (121 GP). The main areas affected are to the south east of the deforested area and this is common to all experiments independent of the scale of deforestation. Given each experiment is run independently this



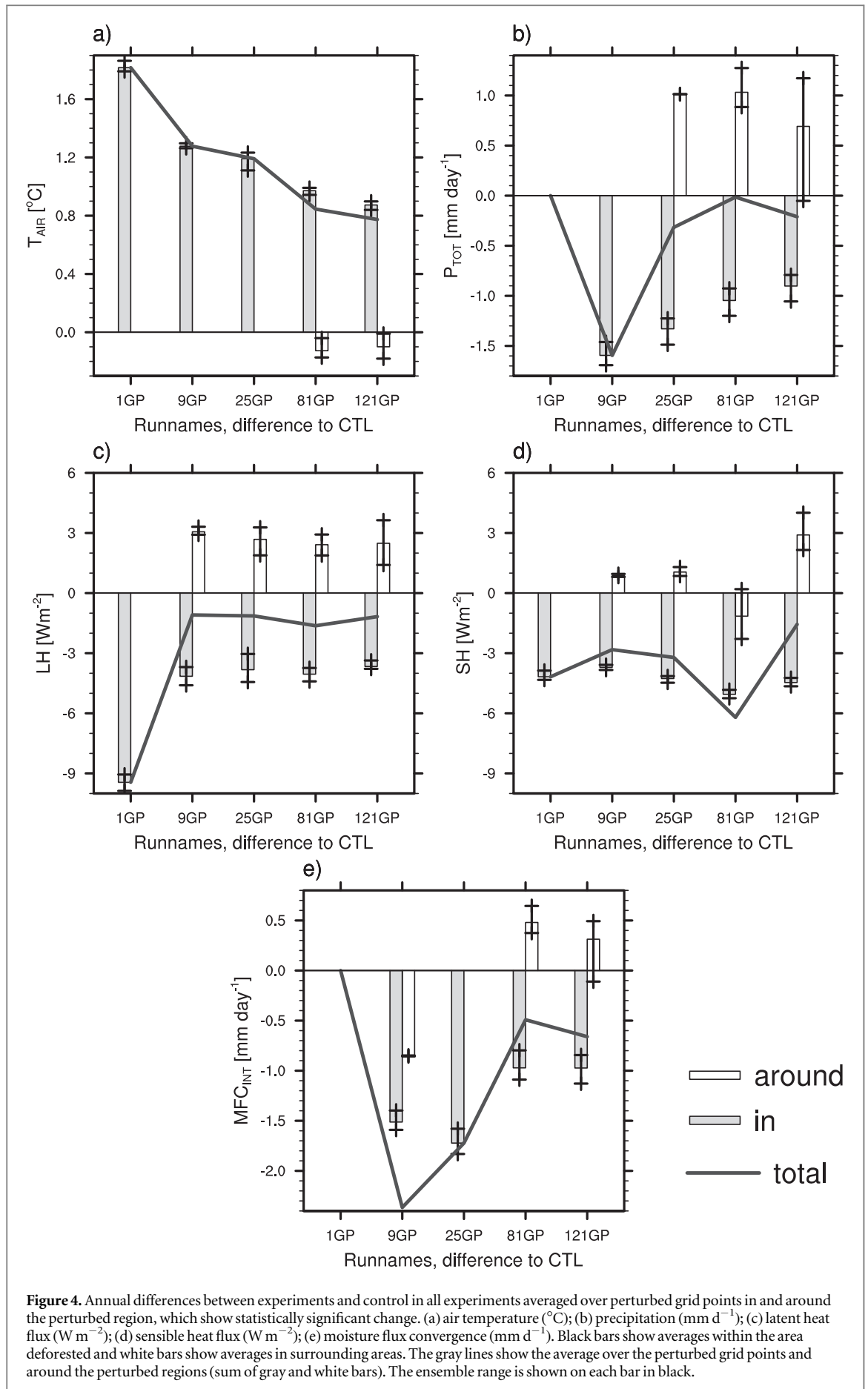
is likely to be a real signal from the model. There is no systematic change in temperature surrounding the area of deforestation because the changes in R_{net} and LH are too small in an evaporation dominated regime such as Amazonia to trigger decreases in local temperature.

Figure 3 shows area-averaged differences between the deforestation experiment and CTL for the area where the deforestation occurs (gray bars) and the grid points around the deforestation (within -18.125°S and 15.625°N , and $82^{\circ}\text{W} - 35^{\circ}\text{W}$, open bars) for albedo, net shortwave radiation (SW_{NET}), net longwave radiation (LW_{NET}) and R_{net} . Results for LH, SH, temperature and precipitation are shown in figure 4. For both figures 3 and 4 only statistically significant changes are taken into account, and the ensemble spread is shown.

The impact of deforestation on albedo is a direct result of the changes in parameter values associated with the change in land cover. The increase in albedo

(figure 3(a)) leads to a decrease in SW_{NET} (figure 3(b)) within the deforested area and an increase in the surrounding area. A similar impact is seen on LW_{NET} (figure 3(c)). In both cases, the decrease within the area of deforestation is largest for 1 GP ($6\text{--}8 \text{ W m}^{-2}$) but is similar for the other experiments ($5\text{--}6 \text{ W m}^{-2}$ for SW_{NET} and $\sim 3 \text{ W m}^{-2}$ for LW_{NET} for all deforestation scales from 9 GP to 121 GP). In all cases above 1 GP, in the areas surrounding deforestation, SW_{NET} and LW_{NET} change in the opposing direction to areas actually deforested. Summing these changes up, the area-averaged difference for R_{net} within the perturbed area is consistent over all experiments above 1 GP (a decrease of $\sim 8\text{--}10 \text{ W m}^{-2}$) while it increases around the deforested area by $\sim 2\text{--}3 \text{ W m}^{-2}$ in all experiments above 1 GP. These changes are consistent among all ensemble members (see range shown in each panel in figure 3).

Air temperature shows the largest area averaged effect in the 1 GP experiment (1.8°C) which then



decreases with increasing deforestation to $0.8\text{ }^{\circ}\text{C}$ in 121 GP (figure 4(a)). Around the deforested area, air temperature shows only a small decrease in the two most extreme experiments ($0.1\text{ }^{\circ}\text{C}$, 81 GP and 121 GP). Total precipitation (figure 4(b)) is decreased within in the deforested area by $0.9\text{--}1.5\text{ mm d}^{-1}$ in all experiments above 1 GP. This effect is largest in the 9 GP case and thereafter decreases with the increasing scale of deforestation. Surrounding the deforested area, precipitation is increased in the 25 GP, 81 GP and 121 GP experiments and this effect is largest in the 25 GP and 81 GP cases ($\sim 1\text{ mm d}^{-1}$, figure 4(b)).

The decrease in precipitation within the deforested area could be caused by local changes (LH, figure 4(c); SH, figure 4(d)) or via changes in moisture flux convergence (figure 4(e)). Our results show changes in LH that match the changes in R_{net} , with changes in SH that are insensitive to the scale of deforestation at least in terms of spatial extent (figure 1, second column). The impact of deforestation on precipitation (figure 4(b)) is relatively strong, declining by 30% between 9 GP and 121 GP (over the area deforested), and does not closely follow the changes in LH. In contrast, as the scale of deforestation increases, moisture flux convergence responds strongly. The decrease in the moisture flux convergence is particularly large in the 9 GP and 25 GP experiments ($1.5\text{--}1.7\text{ mm d}^{-1}$) and declines to $\sim 1.0\text{ mm d}^{-1}$ in the 81 GP and 121 GP experiments (figure 4(e)). The moisture flux convergence also decreases in areas surrounding the deforested area in the 9 GP experiment and changes negligibly in the 25 GP experiments. In contrast, moisture flux convergence increases to $0.3\text{--}0.4\text{ mm d}^{-1}$ in 81 GP and 121 GP experiments. These results are consistent among all ensemble members except for the 121 GP experiment where the ensemble spread for precipitation and moisture flux convergence is inconsistent in terms of the sign of the change. The net impact of the change in the moisture flux convergence is to suppress precipitation most strongly in the 9 GP experiment due to reductions both inside and around the deforested region.

4. Discussion and conclusions

Lawrence and Vandecar (2015) noted that the regional impact of Amazonian deforestation was a warmer and drier climate over the deforested area. They pointed to warming of $0.1\text{ }^{\circ}\text{C}\text{--}3.8\text{ }^{\circ}\text{C}$ and a reduction in precipitation of approximately $0.4\text{--}1.75\text{ mm d}^{-1}$. Our results fall within these ranges, irrespective of the scale of deforestation imposed. Our simulations are therefore broadly consistent with many earlier examples of using climate models to examine the impact of Amazonian deforestation. For large-scale deforestation (81 GP and 121 GP) we find a consistent impact, over the region of deforestation, of warming and reduced precipitation. The large-scale reduction in LH

(figures 2(l) and (q)) associated with lower R_{net} (figures 2(k) and (p)) and changes in biophysical parameters associated with the change in vegetation type can explain the impact of deforestation on temperature (figures 2(m) and (r)). The reduced LH and SH leads to warmer temperatures over the regions deforested because by decreasing both turbulent energy fluxes the land cannot exchange energy as efficiently and consequently warms. The reduced aerodynamic roughness reduces turbulence and vertical mixing such that the lower levels of the atmosphere also warm.

We emphasize that the sign of the change in temperature and precipitation due to deforestation is consistent across all our scales of deforestation. In perturbations ranging from 1 GP to 121 GP, temperatures always increase, but there is a factor of two difference in the amount of increases. Similarly, from 9 GP to 121 GP, precipitation always decreases, but the amount of decrease ranges from $\sim 1.5\text{ mm d}^{-1}$ at 9 GP to $\sim 1.0\text{ mm d}^{-1}$ at 121 GP. While precipitation decreases over the area deforested, which would dry the surface relative to the control experiment and provide a positive feedback on the initial reduction in LH due to deforestation, the main driver of reduced precipitation are changes in the moisture flux convergence. Our results therefore indicate that important fractions of the range in the impact of deforestation summarized by Lawrence and Vandecar (2015) are associated with the scale of deforestation implemented in the various climate model experiments, and that to resolve these differences requires consistency of approach on the scale of deforestation implemented.

Our results also demonstrate the contrasting impact within, relative to outside the deforested area. Figure 4 shows the change in temperature and precipitation averaged over three areas (within the region of deforestation, surrounding the area of deforestation and the combination of these two). For temperature, deforestation causes warming over the area actually deforested but there is little impact on the surrounding grid points, despite the changes in R_{net} and LH shown in figure 2. The impact of deforestation over the combined area is therefore similar to the impact of deforestation over just the area deforested. Figure 4(a) also shows a decreasing impact of deforestation on temperature as the scale of deforestation increases. The changes in R_{net} and LH shown in figure 2 demonstrate that this is associated with an edge effect. Where the deforestation begins, at the edge of the rectangular boxes shown in figure 2, the impact of deforestation is strongest and there is always a region in the center of the overall region of deforestation where there is little change. Therefore, over Amazonia, as the area of deforestation increases, the amount of warming predicted decreases. This result suggests a relatively simple consequence of deforestation on temperature linked with changes in the surface energy balance driven by albedo, through SW_{NET} and LW_{NET} (figure 3).

Changes in R_{net} result in less available energy for the turbulent energy fluxes (figures 4(c) and (d)). This leads to warming, and the lower aerodynamic roughness and consequently lower vertical mixing helps warm the surface air temperature (figure 4(a)).

For precipitation, the existence of horizontal moisture advection leads to a very different impact of scale on deforestation. At small scales of deforestation (1 GP) there is no statistically significant impact on precipitation. This is not to imply that deforesting an area of this scale would have no impact in the real world, rather within a climate model perturbing one grid point does not affect horizontal advection. For larger areas (25 GP and above) enough grid points immediately surrounding the deforested area show a change in precipitation of the opposite sign to the change over the deforested area to affect the overall impact. Figure 4(b) shows that the overall change in precipitation (gray line) reflects decreases in precipitation over the deforested region (gray bars) in combination with increases in precipitation in the grid points surrounding the deforested area (open bars). The impact of deforestation on precipitation decreases relative to the area perturbed as 25 GP, 81 GP and 121 GP are modified and the impact on the surrounding area also decreases. Thus, figure 4(b) shows, as the deforestation scale increases from 25 to 121 grid points, a reducing overall impact on precipitation (a reduction of $\sim 0.0\text{--}0.3\text{ mm d}^{-1}$ shown by the gray line) *only if* both the area deforested and the area immediately surrounding the area of deforestation are combined. If only the area of deforestation is considered then precipitation declines from 1.5 to 1.0 mm d^{-1} while if only the surrounding area is considered then precipitation increases as the scale of deforestation increases.

The reversal of changes from within to outside the region of deforestation needs further discussion. There is evidence that edges, patches or major discontinuities change the sign of the impact of deforestation. As discussed by Lawrence and Vandecar (2015), along borders between forested and deforested regions, mesoscale models predict enhanced convection and, potentially, enhanced rainfall over deforested areas (Baidya Roy and Avissar 2000, Souza *et al* 2000, Wang *et al* 2000). The mechanisms are associated with the warmer temperatures over the cleared surface causing stronger convection leading to an increased convergence of moisture from the adjacent forested region. This would, of course, also affect the simulation of SW_{NET} and LW_{NET} , something we see in our results (figures 3(b) and (c)). We did not save the detailed data from our simulations to enable an analysis of these potential mechanisms, and we caution that this edge reversal explanation is usually applied at local to mesoscales, not at climate model resolutions. However, our simulations were conducted at quite a high resolution and it is possible that these mesoscale

phenomena are resolved at least to some degree and offer an explanation for the edge reversal seen in our results.

A concern in any climate modeling experiment is whether statistical tests appropriately screen internal model variability such that any signal is genuinely associated with the imposed changes in the model. We impose statistical tests that include the modified t-test of Zwiers and von Storch (1995) combined with the Wilks (2006) False Discovery Rate test. We also undertook 5 ensemble members for each experiment. Finally, our experimental design is sequential in the sense that we first perturbed 1 grid point, then in separate simulations perturbed 9 grid points all the way up to 121 grid points. Where we see consistency, and this consistency is statistically significant and exists across the whole range of experiments, we are confident that this is a real signal in the climate model. We therefore are confident that the scaling shown in figure 4 is a genuine reflection of how our climate model behaves. Whether this result is general to all climate models would obviously require further work.

We note two major caveats to our findings. We use a single climate model at a resolution of 1.25° latitude by 1.875° longitude. This is relatively high spatial resolution for Amazonian deforestation experiments but obviously it omits considerable heterogeneity in land processes and mesoscale fluxes which may be extremely important in how deforestation affects the atmosphere (Baidya Roy 2009). Strategies to implement deforestation vary widely among climate modeling groups (Boisier *et al* 2012, de Noblet-Ducoudré *et al* 2012) and these are not reflected in our results. Finally, we use fixed sea surface temperatures; it is uncertain how important this simplification is but some studies have highlighted that larger impacts from deforestation result from using a coupled model (Nobre *et al* 2009, Davin and de Noblet-Ducoudré 2010). Unfortunately, the large number of experiments necessary for this paper would have been prohibitive using a coupled ocean model.

Our results point to several conclusions. First, irrespective of the scale of deforestation imposed in our model, over the regions of deforestation, temperatures increase and precipitation decreases on average. This is apparent for deforestation at the smallest scales we consider (1 GP, a single grid point of area of $28\,051\text{ km}^2$) and at the largest (121 GP, $3\,106\,963\text{ km}^2$). However, the magnitude of this warming is dependent on the scale of deforestation such that both the area averaged warming decreases and the reduction in precipitation decreases as the area of deforestation increases. This is due to different phenomenon. For temperature, the impact is determined internally to the region of deforestation suggesting vertical process dominate the eventual impact via land–atmosphere feedbacks responding to the change in the biophysical parameters and the local surface energy balance. The change in R_{net} , combined with the changes in LH combine through the surface energy

balance and boundary layer response to cause regional scale warming. These changes are local to the areas of deforestation and we see little evidence that temperature is affected over the region of deforestation by processes outside the region of deforestation. The agreement from previous experiments that deforestation causes warming is therefore supported by our results. Our results vary from 0.8 °C to 1.8 °C depending on the scale of deforestation and what area is averaged to provide statistics and this range is a reasonable approximation of the 0.1 °C–3.8 °C noted by Lawrence and Vandecar (2015).

Our results suggest that the uncertainty on the impact of deforestation on precipitation is, at least in part, associated with the scale of deforestation imposed, and what area is averaged to obtain a net result. At scales of 25 GP and above, there is a clear and counter-acting impact from deforestation from areas outside of the region of deforestation. These lead to changes in LH, precipitation and moisture flux convergence (figure 2). As the scale of deforestation increases, the net impact on precipitation decreases (figure 4(b)), and the net impact on rainfall decreases to close to zero for 81 GP and 121 GP if both the within and surrounding areas are taken into account. In other words, reasonable differences in how previous researchers have imposed the scale of deforestation, and whether they include surrounding areas can explain a lot of the 0.4–1.75 mm d⁻¹ range identified by Lawrence and Vandecar (2015). Our range is 0.0–1.6 mm d⁻¹ (figure 4(b)) depending on the scale of deforestation and the area used to derive the average result.

In summary, we find a scale dependency on the impact of Amazonian deforestation associated with an edge effect where the impact of deforestation on R_{net} , LH and temperature, and via moisture flux convergence on precipitation can reverse. This can change the overall impact from deforestation from a significant decrease, to little net change, as the scale of deforestation is increased and the area included in averages is expanded to include areas surrounding deforestation. Our results highlight the need for clear reporting of the imposed area of deforestation, what the impacts are locally, and regionally, and whether a regional scale averages mask changes of opposing signs. We suggest that this would help resolve the wide range of simulated changes in temperature and rainfall reported from deforestation experiments.

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researchdata.anders.org.au/access13b-model-output-experiment-v20/653710. We thank two anonymous reviewers for their extensive and thoughtful comments on our original submission.

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