

Modeling the Impact of Afforestation on Global Climate: A 2-Box EBM

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Abstract

Afforestation programs have become increasingly prevalent around the world as trees are considered crucial in mitigating climate change due to their carbon sequestration potential. In recent years, international agreements such as the Clean Development Mechanism established under the United Nations Framework Convention for Climate Change have notably fueled afforestation activities. However, several complicating factors are often neglected when evaluating the effects of afforestation on global climate. For instance, while carbon uptake by forests reduces the greenhouse effect, the increase in evapotranspiration due to afforestation tends to increase it. An increase in forest cover also lowers the albedo of afforested regions due to the fact that afforestation efforts tend to be carried out on barren lands having relatively high albedo. Further, atmospheric transport exacerbates the cumulative effect of afforestation on global temperatures due to the interaction of poleward transport of sensible and latent heat with ice-albedo feedback.

In this study, we assess the impact of afforestation on global and regional temperatures utilizing a mathematical climate model incorporating carbon dioxide forcing, land/ice albedo feedback, evapotranspiration, and atmospheric heat transport. We investigate the extent to which changes in surface reflectivity and moisture content of the atmosphere caused by afforestation offset the cooling potential of carbon sequestration. In addition, we examine the degree to which these climatic responses depend on the latitude of the afforested region. Considerations such as these have the potential to increase the positive impact of afforestation efforts by identifying land types and latitude regions that, when planted, result in greater mitigation of global warming.

Keywords: afforestation, conceptual climate models.

1. Introduction

The United Nations Conference on Environment and Development, held in 1992, has been successful in generating widespread awareness regarding the need for sustainable development. Following the conference, international market-based development mechanisms, such as the Clean Development Mechanism, were established under the United Nations Framework Convention on Climate Change and have notably fueled afforestation activities. In addition, initiatives such as the Billion Tree Campaign, led by the United Nations Environment Programme, have raised awareness about the carbon-sequestration potential of large-scale plantations, leading to increased interest in afforestation as an effective means of climate change mitigation.

However, a growing body of literature has consistently challenged advocates of afforestation-driven carbon sequestration with the admonition that afforestation can potentially result in a net positive radiative forcing resulting in an overall warming of the global climate. Gibbard et al. (2005) observe that “when changing from grass and croplands to forest, there are two competing effects of land cover change on climate: an albedo effect which leads to warming and an evapotranspiration effect which tends to produce cooling” (p. 1). In this same context, Bonan et al. (2008) argues that while boreal forests create a positive forcing due to a low albedo, tropical forests create a negative forcing through evaporative cooling.

It should be noted, however, that the cooling due to evaporation is local and may only be confined to the region of plantation. In general, the moisture added to the atmosphere from increased surface evaporation will be transported to higher latitudes where it condenses, thereby releasing heat. A number of studies have shown water vapor and atmospheric transport of latent heat are very important in shaping the polar amplified response of the climate to forcing (Flannery 1984, Schneider et al. 1997; Alexeev 2003; Rodgers et al. 2003; Alexeev et al. 2005; Langen and Alexeev 2005; Cai 2005; Langen and Alexeev 2007; Graversen and Wang 2009). Therefore, an increase in poleward latent heat transport due to large scale afforestation in the tropics has the potential to lead to extra-tropical warming despite the cooling effect caused by carbon uptake in the forest. In fact, interactions between the atmospheric transport and the ice/snow albedo feedback could lead to an overall global warming response to afforestation.

Because of its strong greenhouse effect and positive dependence on atmospheric temperature the presence of water vapor in the climate system results in a strong positive feedback independent of transport (Hall and Manabe 1999; Held and Soden 2000). As such, the impact on atmospheric water vapor should be taken into account when assessing the potential climate response to large scale afforestation, not simply the negative forcing due to carbon uptake. For instance, Soden et al. (2002) showed that atmospheric drying resulting from a simulated volcanic eruption in a GCM amplifies the negative radiative forcing due to injected volcanic aerosols. Since large afforestation projects on dry/barren lands lead to increased evapotranspiration, there is the possibility that the water vapor feedback in this case will dampen, or even negate, the radiative forcing (Pielke et al. 2002).

Changes in the albedo of afforested regions should also be taken into account given that many large afforestation projects are carried out on lands that have a relatively high albedo. This decrease in albedo due to afforestation will have a local warming effect, though, again, the potential exists for this local effect to be felt in different latitudes due to the atmospheric circulation. Cess (1978) has shown that extremely long term changes in the surface albedo can double the sensitivity of the global climate to factors which produce climate change.

Betts (2000) simulated radiative forcings due to changes in land surface albedo and argues that for boreal forests the positive forcing induced by the decreases in albedo can fully offset the negative forcing induced by carbon sequestration so that afforestation in high latitudes can lead to warming. In such cases, it may even be argued that deforestation is a preferred strategy for mitigating climate change. Foley et al. (2005), relying on Bonan et al. (1992), says that deforestation in high latitudes can cool the climate due to an increase in surface albedo.

Both high-altitude and high-latitude regions – where one can expect consistent snow cover – have been accepted as regions where deliberate land-use change in the form of afforestation can lead to a net positive forcing due to a lower surface albedo. In addition, GCM simulations by Gibbard et al. (2005) showed that total replacement of current vegetation by trees would lead to warming similar to $2 \times \text{CO}_2$ scenarios while replacement of vegetation by grassland would lead to moderate cooling. Their simulations also indicate that mid-latitude forestation shows the possibility of a potential positive forcing and net warming.

It is evident, then, that there is more than one effect of land-use change on the global climate. In essence, afforestation can do more than simply sequester carbon. Further, the relative effects of the different feedback processes involved are not well understood and are difficult to estimate using GCMs. Even very large plantations are negligible when compared to the land area that is currently forested. Hence, extracting a simulated climate response to a given plantation in a GCM will be problematic given the inherent variability on multiple timescales that exists in most large models.

Forests have complex non-linear interactions with the atmosphere and affect planetary energetics, the hydrological cycle, and atmospheric composition which can dampen or amplify anthropogenic climate change. An additional complication that is important for modeling, as well as model validation, relates to carbon stock assessment in the field. Unfortunately, different assessment technologies sometimes give different estimates of carbon content. Hence, a consensus view on the best methods to use to gauge carbon stocks has not materialized. Of course, it is even more difficult, if not impossible, to develop accurate general formulae for biomass carbon densities across the board (Christie and Scholes, 1995). This seems to call for a more regional, project-specific approach while evaluating afforestation activities.

Most of our current understanding about forests and their interactions with the climate system comes from models, which are abstractions of many complex systems in our atmosphere. It is these models that contribute to policy making under treaties such as the Kyoto Protocol. Because of this, accurate quantification/parametrization of model processes is essential if the policies we enact are to have the effect we intend them to have. But we would add that a knowledge of important climate processes, including the mechanics of their interaction, is of equal importance in shaping policy going forward.

As models become more complex they allow for greater climate prediction, but they also become less useful for understanding and conceptualizing climate systems. For this reason, we chose to consider the impact of afforestation on the global climate using a simplified 2-box energy balance model. The model is sufficiently detailed in that it incorporates the main climate processes governing the interaction between forests and the climate system as discussed above. However, the model's simplicity (it has only two prognostic variables) allows for a focused study of competing climate feedbacks via a qualitative analysis. Also, individual climate processes can be easily switched off in the model to isolate their effect by means of,

for instance, a formal feedback analysis.

Afforestation is widely recognized for its carbon sequestration potential in the policy-world. We think this recognition should be expanded to include albedo changes, water vapor feedbacks, and atmospheric transport of heat. We echo the contention of Pielke et al. (2002) that a system which takes regional effects into account in a new metric will be useful in developing a more comprehensive protocol than what we have currently. A more complete assessment and understanding of the ways in which afforestation can impact the climate system can only aid in our ability to craft sound policies for guiding the implementation of large scale afforestation efforts so they have their intended effect.

2. Model Schematics

The model used in this paper is adapted from a 2-box energy balance model used by Alexeev and Jackson (2013) to assess the relative roles of atmospheric heat transport (AHT) and surface albedo feedback (SAF) in shaping the polar amplified response of the global climate to uniform forcing. It consists of two boxes or regions, shown schematically in Figure 1, one topical and one extra-tropical, dividing the hemisphere equally area-wise at 30°N. Each box contains equal parts land and ocean. The model incorporates surface albedo feedback, atmospheric heat transport, CO₂-dependent emissivity, evapotranspiration, and water vapor feedback in the simplest possible formulation. The change in temperature of the regions is modeled as a function of incoming shortwave solar fluxes, atmospheric heat fluxes (sensible and latent), outgoing longwave radiation, and CO₂ forcing. The model takes into account albedo of the regions as well. Moisture availability for latent heat transport depends on the temperature of the tropical atmosphere as well as the area available for evapotranspiration. We assume free evapotranspiration over both ocean and forested regions, while barren (non-forested) lands are assumed to be dry.

3. Model Equations

Model state variables are T_1 and T_2 , the average temperature of the tropical and extra-tropical boxes, respectively. The temperature of the tropical box is assumed to be independent of latitude, while the temperature of the extra-tropical box is assumed to decrease linearly from T_1 at 30°N. This assumption is justified by the annual zonally averaged meridional temperature profile described, say, in Piexoto and Oort (1992). The extent of the ice cap is determined as the area north of the latitude where the temperature crosses a prescribed freezing temperature, here taken to be -2°C . Model equations are given by an energy balance:

$$\begin{aligned} H dT_1/dt &= S_1 - F - (A + B T_1) + \varepsilon \\ H dT_2/dt &= S_2 (1 - 2\alpha a) + F - (A + B T_2) + \varepsilon \end{aligned} \tag{1}$$

Here H is the heat content of each region, determined primarily by the upper ocean layer heat content; S_1 and S_2 are the net incoming solar fluxes in the tropical and extra-tropical boxes, respectively; A and B are the Budyko-Sellers constants for parametrization of the outgoing long-wave radiation as a function of surface temperature; a is the fractional area of the hemisphere covered by snow/ice; α is the effective ice albedo; and ε represents

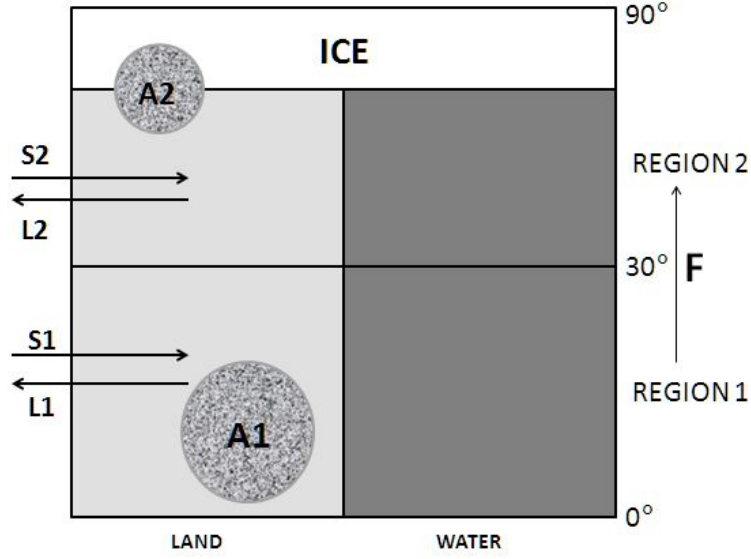


Figure 1: Schematics of the model. S_1 and S_2 are incoming shortwave fluxes. L_1 and L_2 are outgoing longwave radiation. F shows the poleward transfer of heat (latent and sensible) from the tropics to extra-tropics. Forested regions are depicted by A_1 and A_2 .

the carbon forcing. Units for these parameters are in petawatts where 1 PW in either box is equivalent to $10^{15}/\pi r^2 = 7.8 \text{ W/m}^2$ at the top of the atmosphere. The atmosphere is assumed to have minimal heat capacity as compared to the land and ocean. In any case, for purposes of evaluation of the model, H determines only the relative time scale of the model response, hence precision in the actual value is completely unnecessary.

The snow/ice area as a fraction of the hemisphere is determined geometrically as mentioned above:

$$a = 1 - \sin \left(30^\circ + 30^\circ \frac{T_1 - T_{\text{ice}}}{T_1 - T_2} \right). \quad (2)$$

Albedo effects due to afforestation are incorporated into both the solar flux terms, S_1 and S_2 , as well as effective ice albedo, α . Forest area is parametrized as a forest fraction f_i , which represents the fraction of the land surface of region i that is forested. Ice-free surface albedo is then calculated by an area-weighted average of ocean, forest, and barren land albedo. This surface albedo is then used to scale the incoming solar flux. Hence, both S_1 and S_2 depend, respectively, on independent forest fractions f_1 and f_2 .

To determine the effect of ice on the radiation budget we first assume that the extra-tropical forest is well-distributed throughout the region. Secondly, where forest and snow/ice-covered regions overlap we take the albedo to be that of the darker forest. These are both somewhat unrealistic assumptions, but they have been made in order not to underestimate the effect of extra-tropical afforestation on ice albedo. In fact, this assumption will very likely overestimate the albedo effect of extra-tropical afforestation since such efforts tend to occur in the mid-latitudes and the albedo of a snow-covered forested region will be necessarily be a value strictly between that of snow and forest (Betts and Ball 1997).

The ice albedo, $\alpha_I(f)$, a function of extra-tropical forest fraction, is then converted to an effective ice albedo which describes the net reflective effect of the ice-covered surface over that of the ice-free surface, $\alpha_L(f)$:

$$\alpha(f) = \frac{\alpha_I(f) - \alpha_L(f)}{1 - \alpha_L(f)}. \quad (3)$$

The atmospheric heat transport F is parameterized as follows:

$$F = F_0 + \gamma_1(T_1 - T_2) + \gamma_2 C(T_1)(T_1 - T_2) \\ C(T_1) = 6.11 \exp\left(17.23 \frac{T_1 - 273.15}{T_1 - 35.86}\right) \quad (4)$$

The first term in this formula for F describes the mean background value; the second and third terms are included to mimic the sensible and latent heat transports, respectively. Exponential dependence of latent heat transport on T_1 describes the moisture availability in the atmosphere. The particular form given here is that of the Magnus-Tetens approximation to the Clausius-Clapeyron equation which takes into account the temperature dependence of the latent heat of vaporization at the phase-change boundary. We assume that the majority of the moisture in the extra-tropical atmosphere comes from the tropics and therefore T_2 is not included in the expression for $C(T_1)$. Additionally, as discussed above, we relate moisture availability in the atmosphere to the tropical forest fraction, f_1 , by scaling γ_2 by $(1 + f_1)/2$, which represents the fraction of the tropical box that is ‘wet.’

Lastly, we assume the radiative forcing due to carbon sequestration in forests is uniform across both regions due to the fact that CO_2 is a well mixed greenhouse gas. This TOA forcing is computed via the relation given in REF:

$$\varepsilon = c \log\left(\frac{C_0 - C_1 - C_2}{C_0}\right) \quad (5)$$

where C_0 is a base atmospheric CO_2 value and $C_i = (\mu_i * 0.5 * 3.67) f_i L$. That is, C_i expresses the total CO_2 sequestered in a forest of area $f_i L$ where L is the area of each land region in hectares. The biomass density of the forest is μ (tonnes per hectare) and we assume that half of the forest biomass is carbon (Myneni et al. 2001, Penman et al. 2003). Carbon content is converted to CO_2 by multiplying by 3.67, which is the ratio of the molecular weight of CO_2 to that of carbon. The scaling parameter c in formula (5) is often taken to be 6.3 (see Table 2.2 in Houghton et al. 1990), but Myhre et al. (1998) use a detailed analyses of three broadband radiative transfer models to argue for a value of $c = 5.35$. It is this latter value which we use here.

The values of μ used in our simulations below are approximate values based on available data and research. As we noted in the introduction, estimation of forest carbon stocks is a complex field in itself. Given that each plantation is influenced by location-specific factors such as soil, hydrology, and microclimate, it is almost impossible to come up with generic formulae for a particular species. Though detailed procedures and estimates for volume, biomass and carbon content of different species have been produced for IPCC reports (e.g., Annex 3A.1 and 4A.2, Penman et al. 2003), extensive field research being carried out in different parts of the world has shown varying results.

For instance, Gonzalez et al. (2010) estimated carbon densities of forests in California, USA using remote sensing technologies, including lidar and satellite imagery, with calibration by in situ measurements. They report that both lidar and satellite image analysis produce lower estimates of forest carbon density than field estimates. They conclude that lidar captures a more complete picture of areas of low tree density than the field sample, whereas satellite image analysis seemed to systematically undercount live tree density.

Many estimates of forest biomass/carbon density have been produced. Brown and Lugo (1984) surveyed existing volume estimates of tropical forests and produced an estimated weighted biomass densities for undisturbed closed and open tropical broadleaf forests of 176 and 61 tonnes per hectare, respectively. However, a later study using more varied data sources (Brown and Lugo, 1992) saw them raise their mean biomass estimate to 300 T/ha. Milne and Brown (1997) combined numerous surveys and census data to estimate forest carbon densities in Great Britain. They estimated maximum stand densities for the oldest forests at 127 and 173 tonnes carbon per hectare for coniferous and broadleaf species, respectively. However, the mean densities (averaged over age and area distribution) were reported to be 21 and 61 tonnes/ha, respectively.

Similarly, Chaturvedi et al. (2011) report that tropical forests in India show a wide distribution of carbon density based on their age and location. For a very productive site, their sample shows a carbon density of 151 tonnes/ha with a growth rate of 5.3 tonnes/ha/year while on a less productive site the carbon density is reported to be 15.6 tonnes/ha with a growth rate of 0.05 tonnes/ha/year.

For our model, we use forest biomass densities that are consistent with the above estimates. Again, precision in these values is not essential in a qualitative study such as this where we seek to compare bulk climate effects of afforestation across regions.

4. Model Climates

We establish a base climatology range by first tuning our model to reproduce a present-day climate and then finding the equilibrium response of this model climate to uniform forcing (Figure 2). By a ‘present-day’ climate we mean one with average temperature, ice area, and heat transport at 30°N near to their present-day values. Of course, this is somewhat artificial given that we initialize our model with forest fractions prescribed to be zero. That is, our base climatology is determined by an Earth with barren land surface. Nevertheless, we do not consider this a problem given that (1) we are confining our analysis to a qualitative description of model differences with respect to various afforestation regimes and (2) our model is already highly idealized and is not meant to be used for either replication of current climate or prediction of future climates.

Figure 3 shows total AHT and ice latitude across a range of climates. We note that AHT shows a generally positive relation to global temperature with a saturation, and even a possible decrease, in the low-gradient, high-temperature regime located near the +2 PW forcing value. This is consistent with AHT behavior derived from more sophisticated models, for example Caballero and Langen (2005). However, for even warmer climates (upwards of +2 PW) AHT shows a renewed and steady upwards trajectory. This could lead to some concern that our simple parametrization of AHT in equation (4) is insufficient to describe AHT response in very warm climates. However, we note that the experiments carried out in this paper will

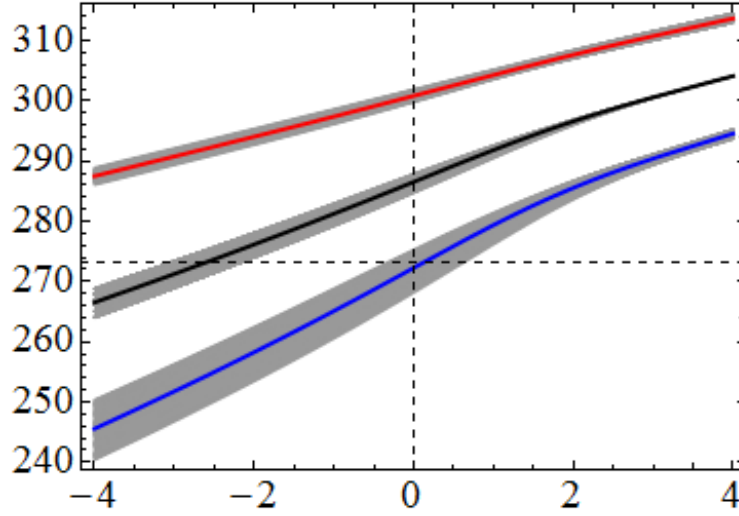


Figure 2: Model climates with $f_1 = f_2 = 0$. Horizontal axis shows TOA forcing value while vertical axis shows temperature. Model parameters are: $S_1 = 41$ PW; $S_2 = 24$ PW; $\alpha = 0.33$; $A = -49.0$ PW; $B = 0.29$ PW/K; $F_0 = 3$ PW; $\gamma_1 = 0.025$; $\gamma_2 = 0.0015$. Grey bands show the effect of 10 % variation in the values for effective ice albedo, γ_1 , and γ_2 .

never result in such high temperatures so we don't consider it a major cause for concern.

5. Afforestation Experiments

As discussed above, the model used in this study has been adapted from a model used to assess the relative roles of AHT and SAF in shaping the polar amplified response of the global climate to uniform forcing. But it is important to note that despite a more or less uniform impact on global CO_2 levels, the climate impact of afforestation is essentially non-uniform given its local effect on albedo and evapotranspiration. Nevertheless, these local effects can have large impacts across the globe due to the atmospheric circulation. Alexeev et al. (2005) investigated the effect non-uniform forcing on a 3D aquaplanet GCM. They found that even without ice-albedo feedbacks a $+4 \text{ W/m}^2$ forcing applied in the tropics resulted in a more or less uniform global response.

Figure 4 shows the outcome of our afforestation experiments. In these experiments the model was subjected to both tropical (case 'T') and extra-tropical (case 'X') afforestation and allowed to equilibrate. Forest fractions from 0 to 0.3 were prescribed in both regions in separate model runs. Three test cases were used corresponding to forest biomass densities of 100, 150, and 200 tonnes/ha. We refer to these cases as 'low,' 'medium,' and 'high' biomass density. Forest albedo was kept fixed in all model runs.

One first notices that model temperatures decrease in all regions for medium to high density forestation, but show no change, or even warming, for low density forest. Of course, there is no difference here between low, medium, and high biomass forests in terms of their effect on surface albedo or evapotranspiration. Hence, the region-specific differences between temperature response in each case is due mainly to the difference in carbon uptake.

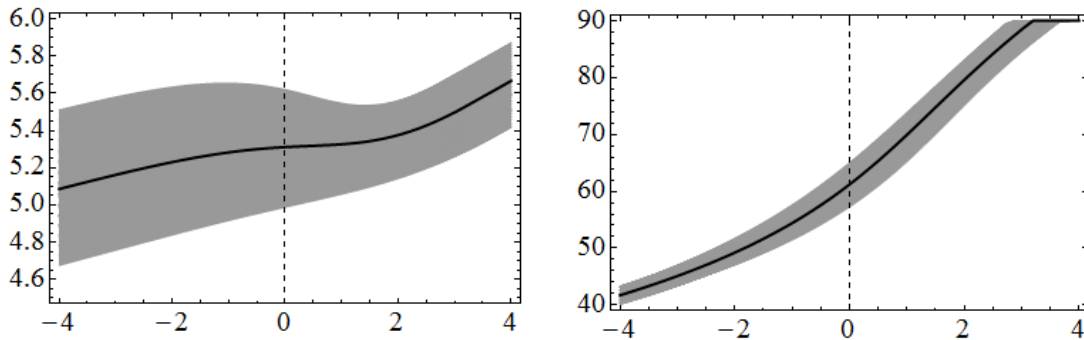


Figure 3: Total AHT in PW (left) and ice latitude in degrees (right) across a range of climates. Horizontal axis shows TOA forcing value. Grey bands show the effect of 10% variation in the values for effective ice albedo, γ_1 , and γ_2 .

Looking at tropical temperatures (Figure 4(a)) we see that the model response differs significantly depending on the region subjected to afforestation, i.e., case T or case X. In particular, afforestation in case X is seen to result in much cooler tropical temperatures when compared to equivalent afforestation in case T. Extra-tropical temperatures (Figure 4(b)) show a similar relation: afforestation in case T leads to cooler temperatures in the extra-tropics as compared to equivalent afforestation in case X. However, the magnitude of the temperature differences in the extra-tropics different afforestation regimes (T, X) are not as great under as they are in the tropics. These temperature differences are shown in Figure 4(e). We see, for instance, that for a forest fraction of 0.3 the tropics will be 2 K warmer if the afforested region is chosen to be in the tropics (case T) rather than in the extra-tropics (case X). On the other hand, the extra-tropics will be 1 K cooler in case T than in case X for the same forest fraction.

We conclude that tropical afforestation leads to increased meridional temperature gradients in our model (Figure 4(f)). This increase in gradient is accompanied by an increase in total AHT at 30°N (Figure 4(d)) which is remarkably consistent for all choices of forest density. Given that AHT acts to extract heat from the tropics, we conclude that the increased gradients in case T are due primarily to the surface albedo effect, both that of the forest and the ice cap: tropical afforestation lowers the albedo of the forested region which will lead to a significant increase in absorbed solar radiation, thus increasing the radiation budget in the tropics while, at the same time, ice albedo feedback tends to increase the meridional gradient in cooling scenarios due to the natural polar amplification exhibited by the model.

Extra-tropical afforestation will lead to decreased meridional temperature gradients, however, as can be seen in Figure 4(f). This is interesting because, again, in cooling scenarios one naturally expects to see polar amplification, and thereby an increase in gradient. This then is an artifact of the non-uniform forcing caused by regional afforestation. The natural polar amplified response of the model is overwhelmed by the local albedo effects of afforestation in case X.

It is interesting to note that the differences between case T and case X are much reduced when viewed from the perspective of global temperatures. Figure 4(c) shows the global temperature response as a function of forest fraction, while Figure 4(e) shows the difference in mean global temperature response between case T and case X. From these plots we see that in the simplest

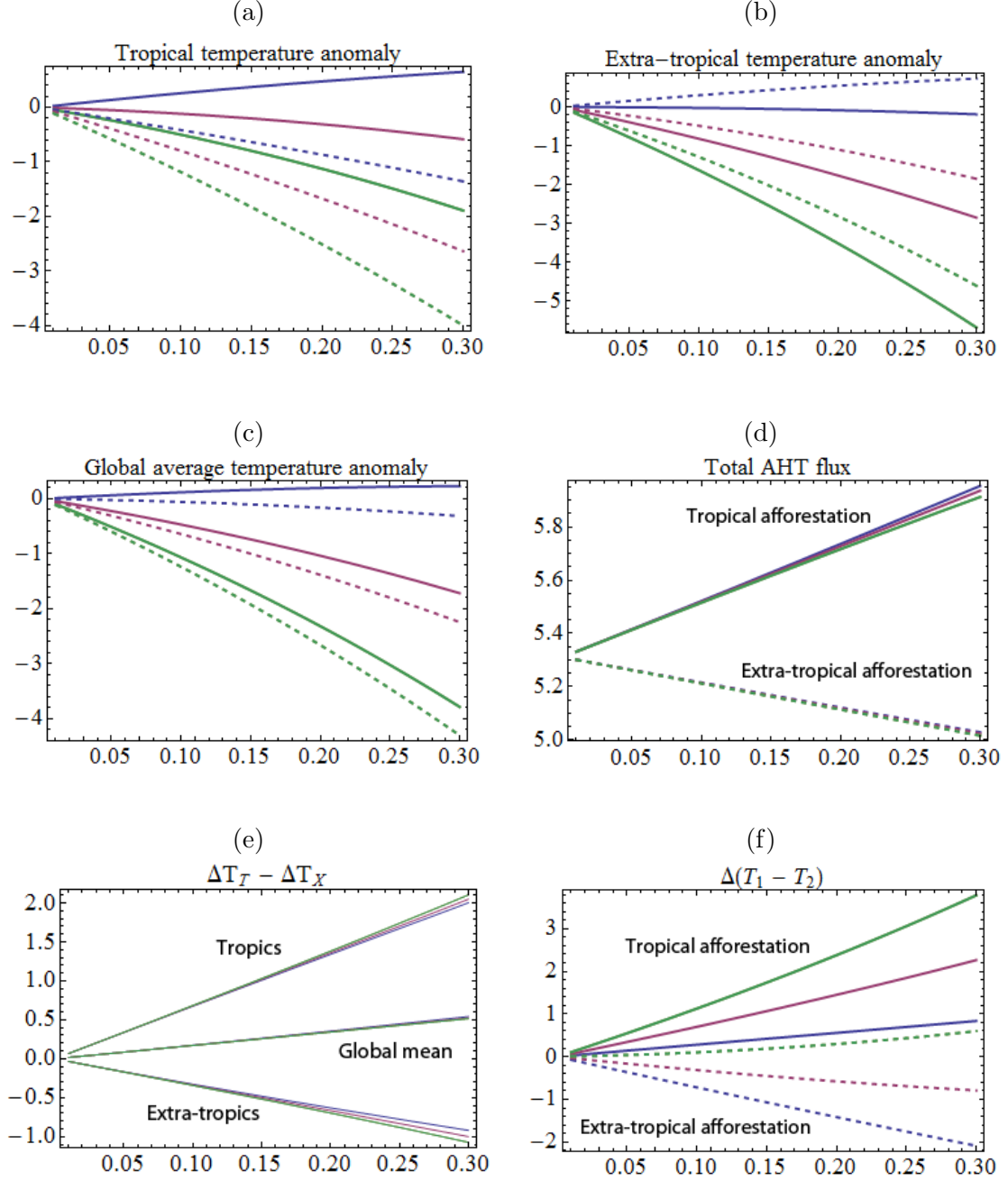


Figure 4: Results of afforestation experiments. Solid lines refer to tropical afforestation (case T), unless otherwise indicated. Dashed lines refer to extra-tropical afforestation (case X).

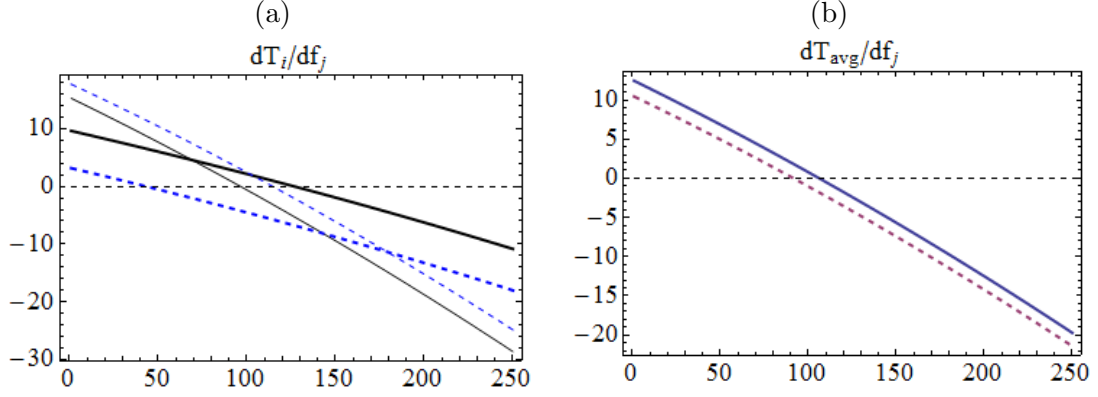


Figure 5: (a) dT_i/df_j for $i, j = 1, 2$; solid lines indicate a tropical afforestation scenario (case T) while dashed lines indicate extra-tropical afforestation (case X); thick lines correspond to tropical temperatures; dashed lines correspond to extra-tropical temperatures. (b) dT_{avg}/df_j for $j = 1, 2$; solid line indicates case T; dashed line indicates case X.

view of global temperature change, there is not much difference between afforestation in the tropics vs. the extra-tropics. Though, interestingly, what difference there is indicates that, all else being equal, a greater cooling effect will result from afforestation efforts carried out in the extra-tropics.

The fact that the response curves in Figure 4 are all linear suggests that for most model variables X , the rates dX/df_1 and dX/df_2 will depend only on the forest biomass density μ . At least, this is the case for sufficiently small forest fractions ($f_i \leq 0.3$). Figure 5 gives plots of these rates (as functions of biomass density, μ) for model temperatures, both globally and in individual boxes. In fact, nonlinearities do show up for very large forest fractions ($f_i > 0.8$) due both to nonlinear model processes (SAF, Clausius-Clayperion) as well as the logarithmic parametrization of the carbon forcing in equation (5).

The solid lines in Figure 5(a) correspond to a tropical afforestation scenario. The μ -value at which they cross (approx. 60 T/ha) is the forest biomass density for which tropical afforestation will result in uniform temperature change in both regions (warming in this case). For larger biomass densities tropical afforestation will result in increased temperature gradients and increased heat/moisture transport even in cooling scenarios (Figure 4(d)).

For extra-tropical afforestation (Figure 5(a) dashed lines), this point of intersection occurs for a much larger μ -value (approx. 180 T/ha). Hence, for most forest biomass densities in our test range (100-200 T/ha), extra-tropical afforestation results in decreased gradients and reduced heat/moisture transport.

It is worth noting once again that, as far as global temperatures are concerned, extra-tropical afforestation has the larger cooling effect in this model. This is seen in Figure 5(b) where the curve in case X crosses the axis first. The difference between mean temperatures in case T and case X are not as large as for regional temperature differences in these two scenarios. However, this difference is not insubstantial either. For instance, the μ -intercepts of the curves in Fig 5(b) differ by approximately 15 T/ha, or approximately 10% of our mean carbon density.

Such an amount could not be considered negligible when assessing the carbon sequestration potential of a given forestation project. Hence, it should not be ignored here where this difference represents the separation between overall warming and overall cooling of otherwise identical afforestation projects based on the region (tropical or extra-tropical) in which they are located.

6. Discussion and Conclusion

Carbon sequestration is only one means by which large scale afforestation projects impact the global climate system. In this study we used a simple 2-box model to illustrate the role of non-carbon processes (albedo effects, increase surface vapor flux, and atmospheric transport of latent and sensible heat) in shaping the global response to non-uniform forcing induced by afforestation.

Our model shows that tropical afforestation tends to increase meridional temperature gradients while extra-tropical afforestation tends to suppress them. Global mean temperatures in our model show a smaller dependence on the latitude of the afforested region, with high latitude plantations resulting in more global cooling than tropical plantations of the same size. This may seem somewhat surprising given that claims in the literature tend to suggest the opposite. However, tropical forests tend to have greater carbon densities and higher albedo than extra-tropical forests (Betts and Ball 1997; Culf et al. 1995). Hence, a comparative analysis such as ours which holds all parameters in common except latitude cannot treat this issue.

Complex general circulation models are growing in their predictive capacity. However, disagreement exists over the parameters needed for accurate quantification of particular afforestation activities as well as the best technologies for determining these parameters. Given this, and the high degree of variability of GCMs on relevant timescales, we feel that a qualitative approach using a simple model is a good alternative.

In any case, given that proponents of large scale afforestation have targets in the range of 30 million ha per year (Nilsson and Wolfgang 1995), it is important to consider more than just carbon when addressing the climate mitigation efficacy of afforestation.

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