Progressive Midlatitude Afforestation: Impacts on Clouds, Global Energy Transport, and Precipitation

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(Manuscript received 23 October 2015, in final form 28 March 2016)

ABSTRACT

Vegetation influences the atmosphere in complex and nonlinear ways, such that large-scale changes in vegetation cover can drive changes in climate on both local and global scales. Large-scale land surface changes have been shown to introduce excess energy to one hemisphere, causing a shift in atmospheric circulation on a global scale. However, past work has not quantified how the climate response scales with the area of vegetation. Here, the response of climate to linearly increasing the area of forest cover in the northern midlatitudes is systematically evaluated. This study shows that the magnitude of afforestation of the northern midlatitudes determines the local climate response in a nonlinear fashion, and the authors identify a threshold in vegetation-induced cloud feedbacks—a concept not previously addressed by large-scale vegetation manipulation experiments. Small increases in tree cover drive compensating cloud feedbacks, while latent heat fluxes reach a threshold after sufficiently large increases in tree cover, causing the troposphere to warm and dry, subsequently reducing cloud cover. Increased absorption of solar radiation at the surface is driven by both surface albedo changes and cloud feedbacks. This study shows how atmospheric cross-equatorial energy transport changes as the area of afforestation is incrementally increased. The results highlight the importance of considering both local and remote climate effects of large-scale vegetation change and explore the scaling relationship between changes in vegetation cover and resulting climate impacts.

1. Introduction

Large-scale increases in forest cover, or afforestation, can drive changes in global circulation and modify atmospheric conditions both locally (Bonan 2008) and in regions far removed from the initial change in tree cover (Zhao et al. 2001; Swann et al. 2012; Devaraju et al. 2015a). Locally, changes in vegetation can drive

changes in near-surface temperature, humidity, soil moisture, and winds. At the global scale, changes in vegetation can modify the energy budget by changing the location and amount of energy absorbed by the surface (Charney et al. 1975), influencing how energy is circulated on a global scale (Swann et al. 2012). Vegetation cover can influence climate through both biogeophysical and biogeochemical pathways (Devaraju et al. 2015b); our analysis explores the biogeophysical impacts of incrementally larger areas of midlatitude afforestation to test the linearity of the response of atmospheric circulation to land-cover forcing in the midlatitudes.

The effects of changes in vegetation on global climate have been largely explored at two scales: the continental to global scale (Fraedrich et al. 1999; Kleidon et al. 2000; Gibbard et al. 2005; Findell et al. 2006; Bala et al. 2007; Swann et al. 2012) and the regional scale, often corresponding to anthropogenic land-cover change (Brovkin

Denotes Open Access content.

Supplemental information related to this paper is available at the Journals Online website: <http://dx.doi.org/10.1175/JCLI-D-15-0748.s1>.

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DOI: 10.1175/JCLI-D-15-0748.1

et al. 1999; Findell et al. 2007). In this study, we aim to bridge the spatial gap between these scales of land-cover change by systematically evaluating the climate response to incrementally increasing the area of mid-latitude forest cover.

Forests can impact the surface energy budget by modifying the surface albedo (thus the amount of shortwave energy absorbed), by modifying outgoing longwave radiation through changing surface temperatures, or by changing surface fluxes of latent heat (the evaporation of water) and sensible heat. Trees generally have a lower albedo than grasslands (Bonan 2002); with all else equal, increasing forest area darkens the surface of the mid-latitudes (reduces the albedo), resulting in an increase in absorbed shortwave solar energy by the surface. Latent heat fluxes differ between plant types and can be the dominant mechanism for affecting local climate in some regions (i.e., cooling the tropics; Bonan 2008), as well as potentially at high latitudes through increasing the greenhouse warming from atmospheric water vapor content (Swann et al. 2010). Surface roughness can influence the amount of energy lost as sensible heat; together with the energy lost as latent heat, this influences the surface temperature and outgoing longwave radiation. The potential effects of changing surface fluxes of energy and water on climate range from local changes in temperature, humidity, and cloud cover to global changes in energy transport resulting from modified latitudinal temperature and energy gradients.

The latitude of forest cover has a strong influence on the biogeophysical impact of that forest on the local climate (Bonan 2008; Li et al. 2015). The influence of mid-latitude ecosystem changes on the atmosphere is less well understood than that of boreal or tropical ecosystems (Bonan 2008). Changes to vegetation in the midlatitudes can modify the global energy balance, which impacts cloud cover, precipitation, and circulation patterns, both locally and in distant regions through atmospheric teleconnections (Swann et al. 2012). Afforestation of the boreal region, particularly by evergreen trees, has a strong warming effect on the climate resulting from the low albedo of forests, particularly during winter months when the trees mask snow-covered ground (Bonan et al. 1992; Foley et al. 1994); tropical forests act to lower surface temperatures by maintaining high water fluxes into the atmosphere (evapotranspiration) (Bonan 2008) and sequestering large stores of carbon (Claussen et al. 2001). In the midlatitudes, both the albedo and water effects of vegetation play a role in modifying the interactions between the land surface and the atmosphere.

Changes in vegetation on Earth's surface have been shown to have major implications for the global hydrological cycle, surface temperatures, and atmospheric

circulation (Fraedrich et al. 1999; Kleidon et al. 2000). Large-scale deforestation experiments performed with coupled global carbon-cycle and climate models have explored how changes in land surface albedo, evapotranspiration, cloud cover, and land-atmosphere carbon interactions affect climate (Claussen et al. 2001; Bala et al. 2007; Bathiany et al. 2010; Devaraju et al. 2015a).

Large extratropical forcings in the climate system have been shown to modify global energy transport by moving the location of the intertropical convergence zone (ITCZ; Kang et al. 2008, 2009). The ITCZ is the region of strong upward motion and heavy precipitation located between the two equatorial cells of the Hadley circulation (Hadley 1735). Several studies have shown that imposing an energy imbalance between the hemispheres in climate models of varying complexity causes a change in cross-equatorial energy flux, thus modifying the location of the ITCZ through atmospheric teleconnections (Chiang and Bitz 2005; Broccoli et al. 2006; Kang et al. 2008, 2009; Swann et al. 2012; Frierson and Hwang 2012; Devaraju et al. 2015a). Swann et al. (2012) showed that replacing all present-day grass and agricultural land in the northern midlatitudes not only resulted in increased midlatitude surface temperatures (regional effect) but also caused a northward shift in the ITCZ (remote effect). An analysis of afforestation across latitude bands suggests that adding trees in boreal and temperate regions has a larger influence on the location of the ITCZ than tropical afforestation (Devaraju et al. 2015a).

In contrast to large-scale global vegetation changes, the impact of anthropogenic land-cover change on regional scales has not been found to have a strong influence on global climate (e.g., Findell et al. 2007; Pongratz et al. 2009). However, both observational (Bonan 2001) and modeling (Bonan 1999; Oleson et al. 2004) studies in the midlatitudes have shown that the conversion of natural forest to cropland cools regional climate. In contrast, analysis of satellite observations found croplands over the United States to have higher surface temperatures than forests during most of the year (Wickham et al. 2012), while a study using local data from eddy covariance towers and weather stations found air temperatures over open land to be cooler than those over forests (Lee et al. 2011). Within the midlatitudes, the effect of forested versus nonforested land cover on surface temperatures can vary spatially; both Lee et al. (2011) and Wickham et al. (2014) found the impact of forests to be different north or south of 35° and 36°N, respectively. However, as shown by the Land Use and Climate, Identification of Robust Impacts (LUCID) experiments, the response of certain terms of the surface energy budget (particularly latent and sensible heat) to

changes in forest cover can vary in direction between models (Pitman et al. 2009; De Noblet-Ducoudré et al. 2012). Fixed sea surface temperatures used in some prior studies would strongly damp large-scale circulation changes and global effects. Here, we present simulations using the Community Earth System Model, version 1.3 (CESM1.3).

We use a global Earth system model to compare the effects of the conversion of large areas of natural and agricultural grasslands to forests in the northern midlatitudes. We explore how the coupled climate system responds to different magnitudes of midlatitude afforestation, with imposed vegetation changes ranging in area from 3.5 to $15.3 \times 10^6 \text{ km}^2$. We focus on the local impact of afforestation on midlatitude cloud cover and on the scaling response of cross-equatorial heat flux and the location of the ITCZ to increasing area of forest cover in the northern midlatitudes. Based on previous studies (e.g., Claussen et al. 2001; Swann et al. 2012), we expect that changes in surface albedo will dominate the local climate impact from an increase in midlatitude forest cover. If the amount of energy absorbed by an increase in tree cover were small enough to be dissipated locally (e.g., through a local change in temperature), we would not expect a change in global circulation. However, if the amount of energy absorbed by afforestation were very large, we would expect some of that energy to be shifted to different locations by atmospheric circulation. Studies of regional-scale vegetation change typically do not find impacts on global-scale climate; the question remains, is the global climate response from smaller-scale changes in vegetation simply overwhelmed by natural variability? Or is there a threshold of land surface forcing on climate required to initiate an atmospheric circulation response? In addition to changes in local temperature, we identify both local and remote cloud feedbacks, which further influence the hemispheric energy imbalance of the global climate system.

2. Methods

a. Experimental design

To test the response of global climate to afforestation of the northern midlatitudes, we incrementally convert what is present-day grassland or agriculture to broadleaf deciduous (BLD) trees. We conduct these experiments using CESM1.3, described below.

We completed a total of five simulations. In the control simulation, the default land-model vegetation map is used (Figs. S1a–d in the supplemental material). In each experiment, a fraction of the grassland and/or agricultural land between 30° and 60°N is replaced with

trees. The first two experiments, Grass50 and Grass100, replace 50% and 100% of the natural C3 grassland in each grid cell with broadleaf deciduous trees, respectively. The last two experiments, GrassAgr150 and GrassAgr200, probe the agricultural space; in addition to having 100% of the natural C3 grassland replaced with BLD trees, they have a further 50% and 100% of agriculture in each grid cell replaced with BLD trees, respectively. C4 grasslands (Fig. S1d) were left untouched as we wished to focus on one type of plant behavior, and C3 plants differ from C4 plants in their water usage; here, we study converting C3 grasslands to C3 forests. The total area of BLD trees added to the northern midlatitudes (30° – 60°N) in each experiment is roughly linear; the area of broadleaf deciduous forest added to the northern midlatitudes is $3.5 \times 10^6 \text{ km}^2$ in the Grass50 simulation, $7 \times 10^6 \text{ km}^2$ in the Grass100 simulation, $11.2 \times 10^6 \text{ km}^2$ in the GrassAgr150 simulation, and $15.3 \times 10^6 \text{ km}^2$ in the GrassAgr200 simulation. The land cover maps used here differ from those used by Swann et al. (2012); as such, our GrassAgr200 experiment, which adds approximately $15.3 \times 10^6 \text{ km}^2$ of forest to the northern midlatitudes, is most similar to Swann et al. (2012), although that simulation added approximately $18 \times 10^6 \text{ km}^2$ of forest.

Experiments are run for 50 years; the first 20 years of each simulation are discarded to allow for model equilibration time. We analyze the 30-yr time series taken from the end of each model simulation. The drift in the global mean temperature over the last 30 years of the control simulation is $0.00023 \text{ K yr}^{-1}$; while temperatures may have equilibrated prior to that (Fig. S2a in the supplemental material), it takes approximately 15 years to allow the leaf area index to equilibrate (Fig. S2b). As such, we discard the first 20 years of the simulation to ensure the full system has reached equilibrium.

b. Model

We conduct our experiments using CESM1.3. CESM1.3 uses the Community Atmosphere Model, version 5 (CAM5; Neale et al. 2012); the Community Land Model, version 4.5 (CLM4.5; Oleson et al. 2013); the CICE, version 4, sea ice model (Hunke et al. 2010); and a slab-ocean model with prescribed heat transport derived from a fully coupled ocean–atmosphere simulation (Haney 1979; Garwood 1979). The land model has an interactive carbon cycle, with the default nitrogen cycle modified to be constant in time following Koven et al. (2015), as nitrogen cycling and feedbacks are not the focus of this study. Atmospheric CO_2 concentration is fixed at 355 ppm. The initial grassland and agriculture distribution in CLM4.5 is determined from a combination of MODIS land cover data (Friedl et al. 2002) and global crop area data from

the year 2000 (Ramankutty et al. 2008). The model resolution used is 1.9° latitude by 2.5° longitude.

The slab ocean is a mixed layer ocean model that calculates the sea surface temperature at each point by closing the surface energy budget given the prescribed ocean heat transport and the atmospheric forcing at that location. Slab-ocean models have been used in previous studies (e.g., Chiang and Bitz 2005; Kang et al. 2008; Swann et al. 2012) when investigating phenomena expected to propagate across the globe. Unlike a fully dynamic ocean, our model setup assumes the response of ocean circulation to our perturbations does not change. Tomas et al. (2016, manuscript submitted to *J. Climate*) show that the dynamic ocean response plays an important role in the climate response to arctic sea ice loss, which, like afforestation, acts as an extratropical forcing on the global climate; use of a slab ocean here (with the same ocean heat transport in each simulation) means that all changes in heat transport we observe occur within the atmosphere. The ocean heat transport used in these simulations was derived from a fully coupled dynamic ocean–atmosphere simulation run under 1850 conditions (in order to estimate ocean heat transport when the system is in quasi equilibrium). The assumed ocean heat transport is then corrected such that the annual global mean heat uptake of the ocean is zero (see further discussion in section 3 of the supplemental material).

The grass, agriculture, and broadleaf deciduous tree plant functional types (PFTs) have different values for leaf albedo, roughness, and photosynthetic parameters controlling stomatal conductance and water resistance. The differences in these parameter values result in different surface fluxes of water and energy between plant types in the land model; they also directly influence the productivity (carbon fixation) of the plant. The land model interacts with the atmospheric model to allow the changes in the land surface to feed back on the climate system by modifying leaf area index (LAI), biomass, and stomatal conductance through time. This allows plant-driven surface properties to vary (e.g., surface albedo through changes in LAI and water fluxes through changes in LAI and stomatal transpiration). The interactive carbon cycle allows plants to take up and release carbon, thus changing leaf area; however, atmospheric CO₂ concentrations are held constant. As such, our results focus on the biogeophysical, rather than biogeochemical, effects of afforestation. Plants can increase or decrease LAI based on the environmental conditions for growth (i.e., carbon, water, and light). The total fraction of each grid cell assigned to a given PFT is fixed. However, if conditions are unfavorable for growth, the carbon-cycle model will shrink plants to a

minimum leaf area, effectively removing the plant from the system. A set of similar experiments was carried out using the CCSM3.4, a precursor to CESM1.3; results from those simulations are presented in the supplemental material.

c. Analysis

Results will be presented as the difference between two simulations; that is, for some variable ϕ , we compute $\Delta\phi = \phi_{\text{experiment}} - \phi_{\text{control}}$. To test statistical significance, we use a two-tailed Student's *t* test to compute if $\Delta\phi$ is significantly different from zero, with $p < 0.05$. The sample standard deviation is taken from the standard deviation of the 30-yr time series of each spun-up simulation. We estimate the degrees of freedom to be $\lfloor n/2 \rfloor = 15$ for $n = 30$ yr, accounting for lagged autocorrelation of up to two years. This is a conservative estimate, as calculating the degrees of freedom following Bretherton et al. (1999) results in 15 or more degrees of freedom in 99% of grid cells (for 2-m air temperature).

We analyze the total radiation (shortwave and longwave) absorbed at the surface under two sets of conditions: clear sky and full sky. In clear-sky conditions, the feedback effects of changes in cloud cover on surface radiative fluxes are ignored; fluxes of shortwave and longwave radiation at the surface are calculated as though there were no clouds present. The radiative impact of the afforestation alone (ignoring any cloud feedbacks) can be isolated using the clear-sky calculation of the surface radiative budget. In full-sky (realistic) conditions, the response of cloud cover is taken into account when calculating surface radiative fluxes. The difference between these calculations (full sky minus clear sky) allows us to isolate the impact of cloud changes on the surface radiation budget.

We are interested in how afforestation changes energy transport on a global scale. To calculate the change in cross-equatorial energy transport, we compute the change in northward energy transport at the equator, following Eq. (2.21) from Hartmann (1994):

$$\int_{-\pi/2}^{\phi} \int_0^{2\pi} R_{\text{TOA}} a^2 \cos\phi \, d\lambda \, d\phi = F_{\phi}. \quad (1)$$

In Eq. (1), R_{TOA} is the net radiation (shortwave minus longwave) at the top of the atmosphere, a is the radius of Earth, ϕ is latitude, and λ is longitude. Thus, F_{ϕ} is the total amount of energy being transported northward by the Earth system at a given latitude ϕ . The cross-equatorial energy flux is the value of F_{ϕ} at $\phi = 0^\circ$ latitude (positive northward).

The magnitude of the cross-equatorial energy flux is compared with the location (latitude) of the tropical

centroid of precipitation, used here as a proxy for the location of the ITCZ. We calculate the centroid of precipitation following Eq. (4) in Lintner et al. (2004):

$$CP = \frac{\int_0^{2\pi} \phi_{\max}(\lambda) \text{ppt}_{\max}(\lambda) d\lambda}{\int_0^{2\pi} \text{ppt}_{\max}(\lambda) d\lambda}. \quad (2)$$

The centroid of precipitation (CP) is the latitude of maximum precipitation (between 20°S and 20°N) weighted by the magnitude of precipitation occurring at each longitude in a latitudinal band. The variable $\phi_{\max}$ is the latitude (between 20°S and 20°N) of maximum precipitation at longitude λ (λ is from 0 to 2π), while $\text{ppt}_{\max}$ is the maximum precipitation at $\phi_{\max}(\lambda)$.

3. Results and discussion

a. Surface energy budget

As the total area of forest cover imposed on the northern midlatitudes increases, the total amount of absorbed shortwave radiation changes. The total change in absorbed shortwave radiation is a combination of the direct albedo change at the surface and changes in energy reaching the surface due to the response of cloud cover to vegetation change.

In the most weakly forced experiment (Grass50), there is only a small, nonsignificant change in absorbed shortwave radiation averaged over the land area between 30° and 60°N (Fig. S3 in the supplemental material). We observe a slight decrease in absorbed radiation for the Grass50 experiment under full-sky conditions and a slight increase in absorbed shortwave radiation for this experiment under clear-sky conditions. The increase in absorbed energy under clear-sky conditions is directly due to the reduced albedo of the surface. Since the total absorbed energy (full sky) decreases, we infer that there must be some compensating increase in cloud cover in the Grass50 experiment. The three more strongly forced experiments, however, all show an increase in absorbed solar radiation, both under clear-sky conditions (albedo effect) and full-sky conditions.

Energy absorbed by the land surface must be removed in the form of sensible heat, latent heat, or longwave radiation. When forests replace grasslands, the total amount of energy exported through each of these pathways changes, which has consequences for local temperatures and cloud cover (Fig. 1). Conversion of grassland to forest causes surface fluxes of latent heat to increase over the northern midlatitudes, particularly over land. Between the control, Grass50, and Grass100 simulations, latent heat fluxes increase with the area of afforestation—consistent

FIG. 1. The change in the average outgoing terms of the surface energy budget over land area between 30° and 60°N for the each simulation with latent heat flux (left solid bar), the net flux of longwave radiation at the surface (middle checkered bar), and sensible heat (right striped bar) (W m^{-2}). Error bars show 95% confidence bounds.

with the idea that forests can access water deeper in the soil. However, as forest cover is further increased in the GrassAgr150 and GrassAgr200 experiments, latent heat fluxes reach a threshold, such that increasing forest area does not result in increased water fluxed to the atmosphere. Instead, surface temperatures rise, and excess absorbed energy is dissipated to the atmosphere in the form of increased sensible heat flux and longwave radiation (Fig. 1).

The threshold behavior in the amount of water able to be fluxed out of the surface (given a roughly linear increase in absorbed energy) over the last three experiments suggests that the system transitions into a water-limited regime when midlatitude forest cover is drastically increased. We will refer to the transition between the control and Grass100 experiments as not being limited by water (in the annual mean), while the experimental space between Grass100 and GrassAgr200 falls into the water-limited regime. We expect the exact magnitude of tree cover required to reach the water-limited regime to be dependent on the representation of soil moisture in the model, in addition to the parameterizations of the water requirements of grasslands and forests. This idea of a water-limited forest state with high sensible heat fluxes is supported by observational studies, which have shown that semiarid forests with high net radiation in dry environments can have limited latent heat fluxes and sensible heat fluxes higher than those found over the Sahara Desert (Rotenberg and Yakir 2010).

While changes in surface fluxes of latent heat reach a threshold and cease to increase with increased forest area, we do not see such threshold behavior in the remaining outgoing terms of the surface energy budget

FIG. 2. The zonal mean change (experiment minus control) in relative humidity is shown for the (a) Grass50, (b) Grass100, (c) GrassAgr150, and (d) GrassAgr200 simulations, for the troposphere between 30° and 60°N. Stippled regions pass a Student's t test ($p < 0.05$).

(longwave radiation and sensible heat). In the weakly forced Grass50 experiment (where water is not limiting), there is a large drop in sensible heat fluxes, of roughly equal magnitude but opposite sign to the change in latent heat fluxes. The changes in the flux of longwave radiation at the surface in this experiment are small. The total amount of energy absorbed over the northern midlatitude land in the Grass50 experiment did not differ drastically from the control simulation (Fig. S3); we postulate that this is because even though the surface albedo was reduced (because of the increased area of dark forest), the increased flux of water to the atmosphere led to compensating cloud feedbacks. From the Grass50 experiment, we deduce that the manner in which energy is partitioned between outgoing surface terms is fundamentally different for grasslands than BLD forests. In particular, for roughly the same amount of available energy, we see that over forests, more of that available energy is balanced through the flux of latent heat, while the energy flux out of a grassland is greatly due to sensible heat (see Grass50 bars in Fig. 1).

In the water-limited regime (experiments Grass100, GrassAgr50, and GrassAgr100), the changes in the surface energy budget are markedly different from those in the Grass50 experiment—in particular, none of the outgoing fluxes of energy decrease as forest cover increases. Because of the threshold in water availability, water fluxes do not increase in this water-limited regime. As we move from the Grass100 to Grass150 experiments, latent heat fluxes do not change, and increased shortwave energy absorption is balanced by increasing sensible heat fluxes and longwave radiation, corresponding to rising land surface temperatures in this regime. The amount of water fluxed to the atmosphere in the water-limited experiments is insufficient to keep up with the temperature increase to maintain the same relative humidity observed in the control simulation, so relative humidity drops (Figs. 2b–d). Thus, in the water-limited regime, increasing forest cover effectively dries out the troposphere, leading to a decrease in clouds. More water is fluxed to the atmosphere than in the control, but the flux of water is not enough to maintain a

FIG. 3. Change in zonal mean cloud fraction for the most strongly forced experiment (GrassAgr200 minus control). Vertical dashed lines mark the region where trees were added between 30° and 60°N. Stippled regions pass a Student's t test ($p < 0.05$).

constant humidity with the substantial increase in air temperature. In comparison, the Grass50 experiment (which is not in the water-limited regime) does not dry the troposphere (Fig. 2a). These changes in relative humidity, driven by the threshold response of water fluxes in the surface energy budget, in turn drive changes in cloud cover over the northern midlatitudes (Fig. 3 and Fig. S7 in the supplemental material), which further influence the amount of solar radiation reaching the surface in this region.

b. Clouds

The impact of afforestation on the local and global energy budget is closely related to the response of clouds. The most significant cloud responses occur over two broad regions: over the northern midlatitudes, cloud cover is reduced in the water-limited regime, while in the Northern Hemisphere tropics, cloud cover increases (associated with a northward shift in the ITCZ). The largest cloud changes occur in the experiments with the largest areas of afforestation.

Temperatures in the Northern Hemisphere troposphere increase with afforestation, allowing the air to hold more water vapor. However, the relative humidity over the northern mid- and high latitudes decreases because of limited water fluxes accompanying the increased atmospheric temperatures (see section 3a). Even though water vapor increases in the lower atmosphere in the mid- to high latitudes, these increases are insufficient to maintain the relative humidity observed in the control simulation, leading to relative drying in the troposphere. This drying of the lower atmosphere leads to a reduction in cloud cover (in particular, a

decrease in low- and midlevel clouds (Fig. 3 and Fig. S7) despite an increase in latent heat flux over the northern midlatitudes. If latent heat fluxes were able to continue increasing into the water-limited regime at a rate sufficient to maintain a constant relative humidity, we would not expect a large decrease in midlatitude cloud cover. In contrast, Betts (1999) observed an increase in temperate (midlatitude) cloud cover in simulations comparing vegetated to nonvegetated land. However, these experiments are not directly comparable, as both the models and patterns of vegetation change imposed are quite different. The largest changes in midlatitude cloud cover occur over the ocean (Fig. S8 in the supplemental material).

The change in midlatitude cloud cover that we observe in our water-limited experiments is in the opposite direction of that which we would expect to result from tropical afforestation, as tropical forests are associated with increased cloud cover (Bala et al. 2007; Bonan 2008; Bathiany et al. 2010). In contrast, our most weakly forced experiment, which is not water limited, has little to no detectable impact on cloud cover. The existence of two regimes of midlatitude cloud response to afforestation of the midlatitudes shows that 1) the influence of midlatitude forests on cloud cover is dependent on the total area of forest cover and 2) we need to understand the direction of the cloud response to afforestation in order to predict the impact of midlatitude forest cover change on climate. This study highlights how the cloud responses driven by midlatitude afforestation can be fundamentally different than those associated with tropical afforestation.

In the Northern Hemisphere tropics, we see an increase in cloud cover in response to midlatitude afforestation. This change in cloud cover occurs at all vertical levels (including deep convective clouds) just north of the equator, with a corresponding decrease in clouds at all levels in the southern tropics (Fig. 3 and Fig. S7); this cloud response shows a clear signature of a northward shift of the ITCZ, which is also evident in changes in tropical precipitation (Fig. S4 in the supplemental material).

c. Global energy budget

The changes in cloud cover affect the amount of solar radiation reaching the surface. Local changes in surface albedo—because of afforestation in the midlatitudes and changes in sea ice cover in the high latitudes—influence the amount of incident solar radiation absorbed. In the zonal mean, we observe an increase in absorbed solar radiation over the northern midlatitudes, as expected from the surface albedo change (Fig. 4a). We also see changes where the surface was not directly modified,

FIG. 4. The change (experiment minus control) in absorbed shortwave radiation at the surface under (a) full-sky conditions and (b) clear-sky conditions (albedo effect). The solid line shows the change in absorbed energy for the most strongly forced case (GrassAgr200). Dashed lines show the response of the remaining experiments, with the more weakly forced experiments lying closer to the zero line. (c) The change in zonally averaged shortwave (red), longwave (blue), and net (black) cloud forcing between the GrassAgr200 experiment and the control. Shading indicates the $p < 0.05$ confidence bounds for the difference of the zonal mean between the experiment and the control.

with a decrease in absorbed solar radiation in the Northern Hemisphere tropics (because of cloud changes) and a large increase in absorbed solar radiation in the northern high latitudes (because of cloud and sea ice changes).

If all else were held equal, we would expect the afforestation experiments to absorb more solar energy at the surface over the regions where vegetation was changed, because of the lower albedo of forests compared to grasslands. However, this response is altered by changes in cloud cover. In the most strongly forced experiment (GrassAgr200), there is an increase of roughly $2.5 W m^{-2}$ in absorbed solar radiation at the surface in clear-sky conditions (Fig. 4b) over the northern midlatitudes ($4.9 W m^{-2}$ over land), which results from the albedo change due to afforestation in the northern midlatitudes (Fig. 4b). When the effect of clouds is considered

(full-sky conditions; Fig. 4a), the increase in absorbed solar radiation over the northern midlatitudes increases to $3.2 W m^{-2}$ in the most strongly forced experiment. That is, not only is the surface in the northern midlatitudes absorbing more solar energy because it is darker (surface albedo effect), but more sunlight is being allowed to reach the surface because of a decrease in cloud cover due to the decrease in relative humidity.

The increase in absorbed solar energy at the surface due to the decrease in midlatitude cloud cover ranges from 10% to 30% of the magnitude of the direct surface albedo forcing between experiments. The cloud forcing can be separated into two parts: the shortwave cloud forcing (how the clouds affect the amount of solar radiation being reflected back to space) and the longwave cloud forcing (how the clouds act to trap longwave radiation in the atmosphere) (Fig. 4c). The decrease in low cloud cover over the midlatitudes results in almost no change in the total longwave cloud forcing over that region, even in the most strongly forced experiment; low clouds have a similar temperature to the surface, so the longwave radiation emitted by low cloud cover and the surface is comparable. However, there is a warming effect from the change in shortwave cloud forcing in the northern midlatitudes—less cloud cover means more solar radiation reaching the surface and thus more energy absorbed. The net effect of changes in cloud cover over the northern midlatitudes is to warm the surface.

In contrast, the cloud response of the Northern Hemisphere tropics acts to reduce the energy coming into the Northern Hemisphere. Tropical cloud cover, particularly in the lower northern tropics (5° – $20^{\circ}N$), increases at all levels (with cloud changes in the tropics reaching higher altitudes than those in the middle and high latitudes). There is a substantial longwave warming effect resulting from cooler atmospheric emission temperatures caused by increased cloud cover. However, there is also an opposing cooling effect caused by the shortwave cloud effect (more bright clouds reflect more incoming sunlight). These competing effects largely cancel each other out (Fig. 4c, black line). The net radiative effect of changes in tropical cloud cover is a slight cooling effect around $10^{\circ}N$, where the largest changes in convective clouds occur, and a slight warming effect in the more northern tropics, similar to that observed in the midlatitudes.

Over the high latitudes, cloud cover in the lower troposphere is reduced, while high cloud cover increases. In contrast to the midlatitudes, where changes in cloud cover lead to more solar energy absorption at the surface, changes in high-latitude cloud cover actually lead to a reduction in solar energy reaching the surface, and the net cloud forcing over the Arctic is negative.

However, reduced sea ice cover results in an overall increase in the energy absorbed by the surface in the Arctic (although this increase is smaller than if cloud cover was unchanged). Temperatures in the Arctic increase dramatically in these experiments (Figs. S9 and S10 in the supplemental material), reflecting the local increase in absorbed solar radiation.

Surface temperatures increase over most of the middle and high latitudes in all our afforestation experiments (Figs. S9 and S10). The observed warming is stronger and more widespread than that found in Swann et al. (2012), who use an atmospheric model with different convection schemes; this could potentially influence both local and remote cloud responses, in turn influencing how much energy is absorbed by the surface. The changes in temperature we observe in our more weakly forced experiments are of similar magnitude and pattern (although of opposite sign) to the temperature changes found by temperate deforestation in Devaraju et al. (2015a), whose magnitude of vegetation change was approximately equal (although opposite, as this study considered deforestation rather than afforestation) to our Grass100 simulation.

d. Impact on tropical energy transport

We observe a change in atmospheric energy transport across the equator in response to the energy imbalance caused by afforestation (Fig. 5a), resulting from a shift in the Hadley circulation. The Hadley cells move energy away from the equator toward the poles. The strength and latitudinal extent of each Hadley cell depends on the season; the winter hemisphere cell straddles the equator, moving energy from the summer hemisphere into the winter hemisphere, while the summer hemisphere cell is narrower and weaker (Kraus 1977). Changes in the position of the Hadley circulation can thus act to reduce hemispheric energy imbalances. The role of the Hadley circulation in transporting heat across the equator means that the location of the ITCZ (between the two Hadley cells) is closely related to the amount of cross-equatorial atmospheric heat transport (Donohoe et al. 2013). The increase in energy absorbed by the Northern Hemisphere as a result of midlatitude afforestation drives a northward shift in the tropical Hadley circulation, with a corresponding northward shift in the annual mean location of the ITCZ (Figs. S4 and S5 in the supplemental material), and an increase in southward atmospheric energy transport across the equator (Fig. 5b).

We find a linear relationship between the magnitude of cross-equatorial energy transport and the change in the location of the centroid of tropical precipitation, with a slope of approximately 8°PW^{-1} (an 8° latitude

FIG. 5. (a) Change (experiment minus control) in southward cross-equatorial atmospheric energy transport (PW) with increasing area of afforestation (10^6 km^2). Error bars show the 95% confidence interval. (b) The change in atmospheric heat transport (PW) is plotted against the change in the latitude of the centroid of tropical precipitation for each experiment. The slope of the line shows the linear fit across experiments. Error bars show the plus and minus one standard deviation estimates for heat transport (x direction) and the location of the centroid of precipitation (y direction).

shift north per petawatt increase in southward heat transport). We note that the relationship between the change in atmospheric heat transport and the change in the location of the ITCZ is much stronger in our simulations than in those presented by Donohoe et al. (2013) and that of the seasonal cycle found by Devaraju et al. (2015a) (see section 4 of the supplemental material). The largest increases in precipitation occur along the northern edge of the band of heavy tropical precipitation, at around 10°N (Figs. S4 and S5); this increase in precipitation along the northern edge of the ITCZ slowly drags the location of the centroid of tropical precipitation [Eq. (2)] northward, with the most strongly forced experiment moving the centroid of tropical precipitation approximately 1.6° latitude northward. The northward shift in the ITCZ is most prominent over the oceans (Fig. S4) and is less prominent over land than the shift found in Swann et al. (2012), who used CCSM3.4. Vegetation change simulations conducted by Devaraju et al. (2015a) with a similar version of CESM show precipitation pattern shifts comparable to those found here.

Linearly increasing midlatitude forest cover leads to a roughly linear increase in absorbed solar energy over the northern midlatitudes (Fig. S3). The increase in absorbed solar radiation over the midlatitudes is the result both of modifying the surface energy budget in the midlatitudes and of inducing both local and remote changes in cloud cover. Midlatitude afforestation therefore induces a hemispheric energy imbalance, driven by the change in surface vegetation cover in the midlatitudes (and the associated high-latitude response). This, in turn, drives a roughly linear shift in atmospheric heat transport across the equator (Fig. 5a). The increase in cross-equatorial atmospheric heat transport is weakest in the Grass50 experiment and strongest in the GrassAgr200 experiment. The Grass100 and GrassAgr150 experiments, which have very similar midlatitude cloud responses, also have very similar shifts in cross-equatorial atmospheric heat transport. The Grass100 simulation actually has a slightly larger change in cross-equatorial energy transport than the GrassAgr150 simulation (although this difference is not statistically significant), and as such, it has a correspondingly larger shift in the centroid of tropical precipitation. Further simulations could make the linearity (or lack thereof) of cross-equatorial atmospheric heat transport to increased midlatitude afforestation more apparent. However, similar simulations conducted using CCSM3.4 did show a more linear (although less strong) response of heat transport to afforestation (Fig. S6 and section 4 of the supplemental material).

The amount of energy transported by the ocean is constant across all simulations. Thus, all the changes in energy transport between experiments occur in the atmosphere. In the real world, both the atmosphere and the ocean would be free to adjust their energy transport mechanisms in response to a hemispheric imbalance. Model simulations by Tomas et al. (2016, manuscript submitted to *J. Climate*) show a northward shift in the ITCZ in response to a high-latitude forcing in a slab-ocean experiment but an equatorward intensification of the ITCZ in a fully coupled dynamic ocean simulation, while Kay et al. (2016) find that Southern Ocean cloud brightening leads to increased cross-equatorial heat transport in the ocean but not in the atmosphere. Thus, it is possible that our results would differ if we repeated our experiments using a dynamic ocean. However, Broccoli et al. (2006) observe a southward shift in the ITCZ in response to Northern Hemisphere cooling when using both slab and fully coupled dynamic ocean models.

Without the strong arctic response, we might expect a given increase in midlatitude forest cover to drive a weaker shift in atmospheric heat transport (reducing the slope of Fig. 5a), thus resulting in a smaller shift in the ITCZ (farther to the right along the line in Fig. 5b). The

FIG. 6. (a) Total arctic sea ice area plotted against area of afforestation in each experiment. Dramatic sea ice loss occurs even under relatively small areas of midlatitude afforestation. (b) Zonal mean change in sea ice fraction over the northern middle and high latitudes with afforestation (experiment minus control). Line weighting and type indicates individual experiments identified in the legend. Gray shading shows the 95% confidence interval.

area of forest cover change needed to observe a statistically significant shift in atmospheric heat transport is on the order of several million square kilometers. Smaller changes in area of forest cover also appear to follow the same linear scaling but fail to pass a significance test because the change is not large enough to overcome the natural variability present in the 30-yr simulations.

e. Response of sea ice

While the response of tropical circulation and atmospheric energy transport to midlatitude afforestation is roughly linear, the response of arctic sea ice exhibits a more threshold-dominated pattern. Dramatic loss of sea ice occurs in the most weakly afforested experiment, while the subsequent two experiments do not yield a further loss of sea ice (Fig. 6a).

Although the land surface in the high latitudes was not directly modified, afforestation in the midlatitudes drives large changes in arctic sea ice and cloud cover (Figs. 4b and 6b), suggestive of a teleconnection. The strongest increase in temperatures with afforestation occurs nonlocally near the surface in the high latitudes (Fig. S9). Near-surface atmospheric temperatures in the

arctic mirror the nonlinear response of sea ice, despite linear increases in temperature over the northern midlatitudes (Fig. S9). This change in arctic sea ice, driven by changes in midlatitude afforestation, could be further driving changes in tropical circulation (Chiang and Bitz 2005). If the response of ice were weaker, we would still expect the energy imbalance driven by afforestation to drive a shift in the ITCZ (see section 4 of the supplemental material, where several experiments were repeated using CCSM3; Swann et al. 2012).

f. Model dependence of results

The five simulations presented here all used the same model, CESM1.3 (see section 2). As shown with the LUCID experiments (Pitman et al. 2009; De Noblet-Ducoudré et al. 2012), the response of the terms in the surface energy budget to changes in land cover varies drastically between models, and what is shown in models does not necessarily match observations (e.g., Bonan 1999 vs Wickham et al. 2012). While we repeated our simulations using an older version of the Community Earth System Model (CCSM3.4; see section 4 of the supplemental material) and found overall similar results (afforestation drives roughly linear changes in atmospheric heat transport), the magnitude of the response between the two models differed substantially. Other models could have a different response in the partitioning of the surface energy budget where vegetation is changed, which would go on to impact the cloud feedbacks we observed in our simulations. Crucially, models will also differ in the soil moisture threshold at which the system shifts between a non-water-limited and water-limited regime as a function of both plant water needs and soil moisture parameterization. We would expect the magnitude of afforestation required to transition into the water-limited regime to vary between models; for example, models with greater soil moisture available to vegetation would require a larger magnitude of afforestation to transition into the water-limited regime.

g. Water-limited regimes

Cloud feedbacks initiated by vegetation change in the tropics have been shown to be a major factor determining the total climate impact of tropical vegetation change (Bala et al. 2007); here, we show that midlatitude vegetation also plays an important role in determining midlatitude cloud cover—although in a different way than in the tropics. We have identified a threshold in the magnitude of latent heat flux from the surface as the area of tree cover over the midlatitudes is incrementally increased, leading to “water limited” and “non water limited” regimes. While the exact value of this threshold is certain to be model dependent, we hypothesize that

the existence of two regimes is likely to be found across models. The cloud feedback response to changing forest cover also depends on the water regime. Before this threshold is reached, in the non-water-limited regime, increased forest cover results in increased water fluxes from the surface; these increased water fluxes, through negative cloud feedbacks, result in very little change in the total amount of absorbed solar radiation in the midlatitudes. Once latent heat fluxes reach this threshold, in the water-limited regime, further increasing forest area does not result in larger water fluxes. Instead, excess absorbed energy resulting from afforestation is emitted through increased fluxes of longwave radiation and sensible heat, corresponding to warming surface temperatures. As temperatures over the midlatitudes rise but water fluxes are unable to increase, the result is a drying of the troposphere. The decrease in relative humidity over the midlatitudes leads to a decrease in cloud cover. Thus, in the water-limited regime of midlatitude afforestation, the midlatitudes absorb more solar energy not only because of the direct albedo effect of darkening the surface with trees but also because more solar radiation reaches the surface resulting from a decrease in cloud cover.

4. Summary and conclusions

We have systematically evaluated the response of global circulation to progressively larger areas of midlatitude forest cover. Midlatitude afforestation drives large changes in local and remote cloud cover, which have substantial impacts on the local energy budget.

The response of cross-equatorial atmospheric heat transport and the location of the tropical centroid of precipitation (associated with the ITCZ) exhibit a relatively linear relationship to the area of midlatitude afforestation. In contrast, arctic sea ice responds drastically to the first $3.5 \times 10^6 \text{ km}^2$ of afforestation and shows little variation in total ice cover until over $15 \times 10^6 \text{ km}^2$ of trees are added to the northern midlatitudes.

While cross-equatorial energy transport responds relatively linearly to changing the area of midlatitude trees, we find that different magnitudes of midlatitude afforestation have nonlinear impacts on the surface energy budget of the northern midlatitudes, which induce local cloud feedbacks. As forest cover increases, the system transitions into a water-limited regime. When water is not limiting, increased afforestation leads to increased latent heat fluxes; when water is limiting, increased afforestation leads to increased sensible heat fluxes and emitted longwave radiation at the surface, while latent heat fluxes plateau. Depending on the magnitude of afforestation, the cloud feedbacks driven

by midlatitude afforestation may compensate for or amplify the surface albedo effect, depending on whether the system is in the water-limited regime. The vegetation changes to the midlatitudes drive a response in the arctic climate, which amplifies the initial midlatitude forcing. The combined mid- and high-latitude surface changes drive an increase in atmospheric heat transport southward across the equator, resulting in a northward shift in tropical precipitation.

This work demonstrates that while some climate effects (such as energy transport) of large-scale midlatitude afforestation scale roughly linearly across a wide range of afforestation areas, others (such as the pathways of the local midlatitude energy budget response) are sensitive to the particular magnitude of midlatitude forcing.

Acknowledgments. The Community Earth System Model used for these simulations is publicly available online from the National Center for Atmospheric Research (CAM5.0 atmospheric model: http://www.cesm.ucar.edu/models/cesm1.2/cam/docs/ug5_3/index.html and CLM4.5 land model: <http://www.cesm.ucar.edu/models/cesm1.2/clm/models/lnd/clm/doc/UsersGuide/book1.html>). Model output from the experiments conducted for this study can be provided by the first author. We acknowledge high-performance computing support from Yellowstone (<http://n2t.net/ark:/85065/d7wd3xhc>) provided by NCAR's Computational and Information Systems Laboratory, sponsored by the National Science Foundation. We acknowledge National Science Foundation Award AGS-1321745 to the University of Washington, and the National Sciences and Engineering Research Council of Canada Award PGS-M444387-2013.

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