Arctic greening can cause earlier seasonality of Arctic amplification

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3 key points

1. Arctic greening amplifies Arctic warming mostly in boreal summer

2. Arctic greening has little impact in boreal winter because of little insolation

3. Polar amplification may appear earlier due to Arctic greening
Abstracts

As global temperatures rise, vegetation types will change, particularly in the northern high latitudes. Under a warming scenario, shrub and grasslands over the Arctic are expected to shift to boreal forests. This study compares the impact of such a change in Arctic vegetation type with that of CO₂ doubling on the seasonality of Arctic warming. Even though vegetation is changed throughout the year, the effect of the surface albedo change is maximum in boreal summer when the incoming solar radiation is largest. Evapotranspiration changes are also maximized in the summer, when the photosynthesis rate is highest. As a result, when Arctic vegetation change is considered in addition to doubled CO₂, Arctic amplification is maximized earlier in the annual cycle.

1. Introduction

Since the beginning of the twentieth century, Arctic surface temperature has increased at twice the rate of global mean temperature [IPCC AR4, 2007]. This feature, referred to as Arctic amplification, is ubiquitous in climate models and observations [e.g. Manabe and Stouffer 1980, Serreze and Barry 2011]. In particular, based on CMIP3 models, Winton [2006] finds a mean annual Arctic (60°-90°N) warming that is 1.9 times greater than the global mean warming at the time of CO₂ doubling. One of the key features of amplified Arctic warming is a pronounced seasonality as winter warming exceeds summer warming by at least a factor of 4 [Bintanja and van der Linden 2013].

In response to Arctic warming, high latitude biome changes are characterized by habitat migration [Sturm et al., 2001], a longer growing season [Jeong et al., 2011a], and enhanced photosynthetic activity [Xu et al., 2013]. These are generally considered as
evidence of Arctic “greening”. In fact, over the past century, ground observations show that vegetation has been increasing, especially in the Arctic tundra where an extension of shrub area is prominent [Tape et al., 2006; Bunn et al., 2006; Sturm et al., 2001; Potter et al., 2013]. Pearson et al. [2013] reported that the Arctic has experienced significant changes in vegetation, including a northward migration of the Arctic tree-line, during the last several decades. In line with observed relationships between warming and biome changes, models also project circumpolar greening with a northward expansion of boreal forests and enhanced greenness of tundra regions with increasing CO₂ [Notaro et al., 2007; Jeong et al., 2011b (hereafter JSJ11)].

Greening in the Arctic in a warmer climate can amplify Arctic warming by two to seven times due to decreased surface albedo [Levis et al., 2000; Chapin et al., 2005; Bonan 2008]. Transpiration of water vapor and feedbacks from the ocean and sea-ice also contribute to amplifying high-latitude warming [Swann et al., 2010 (hereafter ALS10); Loranty et al., 2011]. These studies confined their analysis to the annual mean. Although vegetation changes throughout the year, its effect on Arctic climate may exhibit large seasonality as there is no sunlight to reflect during the polar night. Therefore, we study how changes in Arctic vegetation affect the seasonality of Arctic amplification.

JSJ11 show that the land surface temperature response to vegetation change is sensitive to its latitudinal position. In midlatitudes, increasing vegetation decreases warming due to increased evapotranspiration, whereas in the high latitudes it causes additional warming due to the reduced surface albedo. To avoid complications from these contrasting effects, we only consider changes in vegetation poleward of 60°N. In contrast to most previous studies that focused on the impact of global-scale vegetation changes, this study allows us to identify and isolate the effect of Arctic vegetation changes on local and remote climate.
2. Model and Experiments

2.1 Model

The global climate model (GCM) used in this study is the National Center for Atmospheric Research (NCAR) Community Atmospheric Model version 3 (CAM3; Collins et al. 2004). We use spectral T42 resolution (2.875° x 2.875°) horizontally and 26 hybrid-sigma levels in the vertical. The model is coupled to a mixed layer ocean, with spatially varying mixed layer depth where the sea surface temperatures (SSTs) freely evolve to ensure that the surface energy budget is closed. Land-surface processes in CAM3 are calculated with the Community Land Model version 3 (CLM3; Oleson et al. 2004) that calculates the heat, moisture, and momentum fluxes between land surfaces and atmosphere as well as the thermal and hydrologic processes at the surface and the interior of near-surface soil layer.

Coupled with CAM3 at T42 horizontal resolution, CLM3 is comprised of 3,799 grid points, each a collection of sub-grid elements of four primary land cover types: glacier, lake, wetland, and vegetation. The vegetated portion of the grid cell is represented by the fractional coverage of plant functional types (PFTs). The model uses seven primary PFTs: namely, needle-leaf evergreen or deciduous trees, broadleaf evergreen or deciduous trees, deciduous shrubs, grass, and crops. These seven primary PFTs are further refined to tropical, temperate, and boreal deciduous or evergreen trees, C3 and C4 grasses, and evergreen and deciduous. In each PFT, leaf phenology in the CLM3 is prescribed, and the seasonal course of leaf area index (LAI) for each PFT is derived through interpolating the monthly PFT-specific LAI from National Oceanic and Atmospheric Administration (NOAA) Advanced Very High
Resolution Radiometer (AVHRR) data. We integrate the model for 100 years, and use the mean of the last 80 years as an equilibrium state.

2.2 Experiments

The control integration, denoted as CNT, uses present-day CO\textsubscript{2} concentrations (335ppmv) and vegetation types set to the current climate distribution. The PFT in CNT shown in Figure 1a indicates that shrubs and grass are the dominant vegetation types poleward of 60°N. JSJ11 show that northern high latitudes with grass and shrubs will likely be replaced mostly with boreal forests in response to doubled CO\textsubscript{2} concentrations. Hence, we study the effect of poleward expansion of forest under global warming by replacing grass/shrub with boreal forest north of 60°N in CNT as shown in Figure 1b. We use the extreme case of all grassland and shrubland being replaced in order to study the maximum impact of vegetation change on Arctic amplification. For instance, the imposed changes in LAI in June-July-August are 25% larger than the simulated response to doubling of CO\textsubscript{2} by the dynamic global vegetation model presented in JSJ11. In the study of JSJ11, the model is run with fixed SST and sea ice concentrations (SICs) (to that derived from the ocean-atmosphere coupled model under either present-day or doubled CO\textsubscript{2} condition), whereas in our study the SST and SICs are allowed to adjust. Monthly LAI in Figure 1c indicates that there is a distinct growing season of shrub and grass types from May to September in CNT (blue), but under doubled CO\textsubscript{2} (red), vegetation exists all year long, especially increasing from June to November.

In order to differentiate the response to direct CO\textsubscript{2} radiative forcing from that of Arctic vegetation change, we perform three perturbation experiments: (1) with present-day
CO₂ concentrations and altered vegetation types poleward of 60°N (denoted as VEG), (2) with CO₂ doubled, but vegetation fixed to present-day conditions (denoted as 2CO2), and (3) with CO₂ doubled and altered vegetation types poleward of 60°N (VEG+2CO2). We analyze the differences between each perturbation experiment and the control integration.

3. Results

Figure 2a shows the simulated annual- and zonal-mean change in temperature at 950 hPa (T₉₅₀) in all experiments. We note that the following discussion qualitatively holds for all vertical levels below 800 hPa. In response to CO₂ doubling (red), the polar regions warm twice as much as the tropics, with a similar magnitude of warming at both poles. In response to Arctic vegetation change (blue), the annual-mean temperature increases by 2.3 K poleward of 60°N, with a peak at 65°N where land albedo is changed most. The spatial pattern of warming is shown in Figures 2b-2d. In a study by ALS10 where the effect of Arctic vegetation changes is investigated by adding deciduous forest on bare ground at high northern latitudes, the near surface Arctic warms by only 1 K in the annual-mean, because the total converted area is 3.3 times larger in our experiment setup compared to that in ALS10. The differences in experimental setup cause twice as much reduction of surface albedo in April and 5 times as much reduction in July as compared with ALS10. There is 1.5 times the increase in transpiration in April, and 20 times the increase in July in our study compared to ALS10, which induces a larger water vapor feedback. The sum of the responses in VEG and 2CO2 (black dashed) almost coincides with the response in VEG+2CO2 (black) for latitudes equatorward of 65°N. This linearity holds at every month with high accuracy (Figure 3a), allowing us to decompose the response in VEG+2CO2 into the response from CO₂ doubling and from Arctic vegetation change. One third of the annual-mean T₉₅₀ increase poleward of
60°N in VEG+2CO2 results from VEG, but its relative contribution varies with month. The contribution from VEG reaches 55% during the growing season and falls to 11% in boreal winter.

Figure 3a shows the monthly changes in T950 averaged poleward of 60°N in all experiments. In 2CO2 (red), the warming response over the northern high latitudes is largest in November, and smallest in June, consistent with the CMIP5 simulated response to increased CO2 [Pithan and Mauritsen, 2014]. The maximized warming occurs in fall and early winter in part because most incoming solar radiation in summer is used to melt ice and warm areas of open ocean, whereas during late fall and winter, the ocean transfers heat to the atmosphere [Screen and Simmonds, 2010]. The remainder of the winter preference in Arctic amplification is due to the seasonality of the lapse rate feedback [Pithan and Mauritsen, 2014]. In VEG (blue), during the growing season, Arctic warming is intensified by 3.5 K, 20% larger than the response in 2CO2. The response is much larger than that shown in Jeong et al. [2012] of 1 K, possibly because of the larger vegetation changes and the positive feedback from SST and SICs in our model. The predominant temperature increase during the growing season is due to reduced surface albedo from the change in vegetation type (Figure 3b). In fact, there are two peaks in surface albedo reduction, one in May associated with land greening and the other in October associated with sea ice melt (Figures 3b and 3c). However, since little sunlight reaches the high latitudes in late fall and winter, the reduced albedo in late fall does not lead to more absorbed solar radiation (Figure 4). Hence, although vegetation is changed throughout the year, the amount of insolation reaching the Arctic surface is most enhanced in June. Moreover, as ALS10 suggested, the greenhouse warming by additional atmospheric water vapor content resulting from enhanced evapotranspiration in the summer (dashed pink) can also contribute to a maximized warming during the growing season. In
ALS10, the radiative forcing of enhanced water vapor is the same order of magnitude as the shortwave forcing from albedo reduction. However, in our experiment setup with imposed large changes in albedo, we expect the effect of surface albedo to overwhelm the effect of an increase in water flux from transpiration, as is implied in Figure 4.

The large seasonality in incoming solar radiation over the Arctic causes the large seasonal variation both in net shortwave flux at the surface and in evapotranspiration. As a result, when vegetation changes are considered in addition to doubling of CO$_2$ (black in Figure 3a), Arctic amplification is maximized earlier than when only CO$_2$ doubling is accounted for, leading to a weaker seasonality in temperature change.

4. Conclusions

We study the effect of Arctic greening on the seasonality of polar amplification using CAM3 coupled to mixed layer ocean and CLM3 with prescribed vegetation type. We compare the effect of CO$_2$ doubling with the effect of changes in Arctic vegetation type. In particular, we change shrublands and grasslands poleward of 60°N to boreal forest, which is a potential extreme scenario under doubled CO$_2$ suggested by JS/11. Even though vegetation is changed over a relatively small area, its effect on polar temperature is comparable to CO$_2$ doubling. In the annual-mean, the lower tropospheric temperature response poleward of 60°N to Arctic vegetation type change is 60% of the warming from CO$_2$ doubling. Hence, the Arctic vegetation type change will amplify regional warming due to CO$_2$ increases, although, as stated in Section 3, the exact magnitude of the response would be sensitive to the experiment setup. However, the qualitative effect of Arctic greening on the phase shift of peak Arctic warming, which is the focus of our study, is expected to be a robust feature. It is
important to understand the factors that affect the seasonal cycle of Arctic warming because it affects Arctic ecosystems, economic activities such as shipping and fisheries, and the mass balance of glaciers and ice caps that affects sea level.

Even though vegetation is changed all year long, its ability to change absorbed shortwave radiation is maximized in June when the incoming solar radiation reaches a maximum over the Arctic. Also, evapotranspiration is enhanced the most in the summer, which has the highest photosynthesis rate and the largest change in radiation, leading to the largest increase in water vapor and consequently the greenhouse effect. Conversely, from November to January, little solar radiation reaches the Arctic, so vegetation changes have little impact on the Arctic temperature. Hence, although vegetation is increased throughout the year, its ability to warm the Arctic is confined to boreal summer because of the large seasonality of insolation. Therefore, the peak of Arctic warming that appears in November when a doubling of CO$_2$ is prescribed in isolation shifts to August when Arctic vegetation change is additionally taken into account, leading to a weaker seasonality in Arctic amplification.

Although there is uncertainty in the potential vegetation change, we expect details of vegetation will primarily affect the magnitude of the response while holding the qualitative results the same. The present study emphasizes the importance of improving vegetation parameterizations to raise the precision of climate change projections.

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References


Figure 1. Plant Functional Types (PFTs) in (a) CNT and (b) VEG, with green and yellow corresponding to boreal forest and shrub/grasslands, respectively. (c) Monthly Leaf Area Index (LAI) averaged poleward of 60°N in CNT (blue) and in VEG (red).

Figure 2. (a) The response of annual-mean zonal-mean 950hPa temperature to vegetation change (blue), CO₂ doubling (red), and both vegetation and CO₂ doubling (black). The sum of blue and red is shown with the black dashed line. The shading indicates plus/minus one standard deviation. The map of annual-mean 950hPa temperature change to (b) vegetation, (c) CO₂ doubling, and (d) both vegetation and CO₂ doubling.

Figure 3. Monthly response averaged poleward of 60°N of (a) 950hPa temperature, (b) surface albedo, and (c) sea ice fraction, to vegetation change (blue), doubling of CO₂ (red), and both vegetation and CO₂ doubling (black). The sum of blue and red is shown with the black dashed line. The shading indicates plus/minus one standard deviation.

Figure 4. Monthly response averaged poleward of 60°N in VEG of net shortwave radiation (SW, red), net longwave radiation (LW, blue), sensible heat fluxes (SH, cyan), latent heat fluxes (LH, pink), and evapotranspiration (ET, dashed pink). Positive indicates downward fluxes.
Figure 1  Plant Functional Types (PFTs) in (a) CNT and (b) VEG, with green and yellow corresponding to boreal forest and shrub/grasslands, respectively. (c) Monthly Leaf Area Index (LAI) averaged poleward of 60°N in CNT (blue) and in VEG (red).
Figure 2 (a) The response of annual-mean zonal-mean 950hPa temperature to vegetation change (blue), CO2 doubling (red), and both vegetation and CO2 doubling (black). The sum of blue and red is shown with the black dashed line. The shading indicates plus/minus one standard deviation. The map of annual-mean 950hPa temperature change to (b) vegetation, (c) CO2 doubling, and (d) both vegetation and CO2 doubling.
Figure 3 Monthly response averaged poleward of 60°N of (a) 950hPa temperature, (b) surface albedo, and (c) sea ice fraction, to vegetation change (blue), doubling of CO₂ (red), and both vegetation and CO₂ doubling (black). The sum of blue and red is shown with the black dashed line. The shading indicates plus/minus one standard deviation.
Figure 4 Monthly response averaged poleward of 60°N in VEG of net shortwave radiation (SW, red), net longwave radiation (LW, blue), sensible heat fluxes (SH, cyan), latent heat fluxes (LH, pink), and evapotranspiration (ET, dashed pink). Positive indicates downward fluxes.