Effects of regional afforestation on global climate
Ye Wang, Xiaodong Yan and Zhaomin Wang

ABSTRACT

Carbon (C) sequestration following afforestation is regarded as economically, politically, and technically feasible for fighting global warming, whereas the afforested area which will contribute more efficiently as sinks for CO₂ is still uncertain. To compare the benefits for C sequestration combined with its biogeochemical effects, an earth system model of intermediate complexity, the McGill Paleoclimate Model-2 (MPM-2) is used to identify the biogeophysical effects of regional afforestation on shaping global climate. An increase in forest in China has led to a prominent global warming during summer around 45° N. Conversely, the forest expansion in the USA causes a noticeable increase in global mean annual temperature during winter. Afforestation in the USA and China brings about a decrease in annual mean meridional oceanic heat transport, while the afforestation in low latitudes of the southern hemisphere causes an increase. These local and global impacts suggest that regional tree plantations may produce a differential effect on the Earth’s climate, and even exert an opposite effect on the annual mean meridional oceanic heat transport; they imply that its spatial variation of biogeophysical feedbacks needs to be considered when evaluating the benefits of afforestation.

Key words | climate change, land cover change, modeling, regional afforestation

INTRODUCTION

In the future, afforestation might be chosen as an option for the enhancement of terrestrial carbon sequestration for global warming. Afforestation affects global climate via biogeochemical and biogeophysical processes. The absorption of carbon from the atmosphere owing to afforestation affects climate by altering incoming solar radiation and out-going infrared (thermal) radiation that are part of Earth’s energy balance (Jackson & Masabathula 2015). Such biogeochemical effect of afforestation is quite significant. For example, Nilsson & Schopfhauser (1995) estimated that over the period from 1995 to 2095, a total of 104 Gt of carbon would be sequestered. The ongoing redistribution of mineral soil carbon (C) in the young stands and the higher soil C contents in the 200-year-old afforested stands suggests that nutrient-rich afforestation soils may become greater sinks for C in the long term (Vesterdal et al. 2002). In addition, C drawdown of afforestation can be counteracted by their biogeophysical effects and, in boreal latitudes, even overcompensated due to large albedo differences between forest canopy and snow (Bathiany et al. 2010). Due to the exclusion of carbon cycle model within our climate system model we focus here on the biogeophysical effects of the afforestation on climate.

Biogeophysical mechanisms for afforestation include the effects of changes in albedo, surface roughness, and evapotranspiration (ET) (Figure 1). In general, replacing grassland or desert with forests leads to a decrease in surface albedo and an increase in ET (Whitehead 2014). Expansion of forest increases surface roughness affecting the energy and momentum balances in the boundary layer by increasing the ability of air to mix. Various studies have simulated these effects and discussed their causes. Historical deforestation at mid-latitudes has increased surface albedo which has cooled the climate of the northern hemisphere (NH) (Betts 2001; Govindasamy et al. 2001; Bounoua et al. 2002; Bonan 2005; Browkin et al. 1999, 2006; Pitman et al. 2009).
Biogeophysical mechanisms of historical land cover changes (LCC) tend to have caused a global cooling of 0.09 to 0.22 °C since 1700, primarily by increasing surface albedo (Matthews et al. 2004). On the other hand, it is suggested that large-scale removal of tropical forests could reduce evapotranspiration rates, leading to a local reduction in precipitation and an increase in land surface temperature (Henderson-Sellers et al. 1996; Lean & Rowntree 1996; De Fries et al. 2002; Feddema et al. 2008). Claussen et al. (2004) found that tropical deforestation tends to warm the planet because of the larger effects of the increase in atmospheric CO₂ compared to the biogeophysical effects, while, in mid-high northern latitudes, biogeophysical processes win over biogeochemical processes, eventually leading to a global warming in the case of afforestation. It has been suggested that the effects of afforestation are different in the tropics and at high latitudes (Claussen et al. 2001; Pielke et al. 2002; Bala et al. 2007; Betts et al. 2007; Bonan 2008; Davin & Noblet-Ducoudré 2010). However, the effects of regional afforestation on global climate have never been studied. In addition, most of the previous land cover change experiments have been conducted using atmospheric general circulation models (AGCMs) with the absence of feedbacks between the ocean and the atmosphere. However, the responses of climate to afforestation are affected by feedbacks from sea surface temperatures (SSTs) and sea ice, even thermohaline circulation of the ocean.

To allow centurial model integrations and maintain broad-scale geographic patterns, we employ an earth system model of intermediate complexity (EMIC), McGill Paleoclimate Model-2 (MPM-2) (Fanning & Weaver 1996; Wang & Mysak 2000), to study biogeophysical effects of regional afforestation on climate. EMICs include almost all components of the Earth’s system. They are simplified but geographically explicit models capable of simulating the feedbacks and interactions among the components in climate systems. Computational efficiency of these models allows simulation of all the main processes on a long-term scale to investigate the influence of uncertainty in climatic forcings and process parameterizations on model results through many sensitivity experiments (Forest et al. 2002).

The promotion of afforestation has often been cited as a strategy to slow global warming. Afforestation absorbs CO₂ from the atmosphere, which exerts a cooling influence on the Earth’s climate. However, biogeophysical effects of afforestation, which include changes in land surface albedo, ET, and surface roughness also affect climate. Here, we focus directly on the effects of future afforestation regionally within the framework of a climate system model. We also test the response of mean meridional oceanic heat transport due to regional forest expansion. Our goal here is not to realistically simulate possible scenarios of afforestation, nor to reproduce the observed pattern for afforestation, but rather to compare the potential climate change over global scales due to differential regional afforestation.

**MATERIALS AND METHODS**

We used the MPM-2 (Wang & Mysak 2000). This model has been applied to study a number of natural and anthropogenic climate forcings (Wang & Mysak 2001; Wang et al. 2002, 2005, 2014; Wang & Yan 2013). The MPM-2 is a global climate model, which consists of an energy and moisture balance atmosphere model, a multi-basin zonally averaged dynamic ocean model, a dynamic ice sheet model, a zero-layer thermodynamic-dynamic sea-ice model and a land biosphere model. The north–south resolution is 5°, except across the equator where it is 10°. MPM-2 has been downscaled to 5° × 5° in or over North America and Eurasia. The atmospheric module of MPM-2 is represented by a relatively simple 2D (two-dimensional) energy and moisture balance model (EMBM)
(Forest et al. 2002), which has a new parameterized solar energy disposition scheme (Wang et al. 2004). The ocean module is a classic zonally averaged dynamic model based on vorticity conservation, which has nine vertical layers and a flat bottom (Wright & Stocker 1991). The MPM-2 employs a simple zero-layer thermodynamic-dynamic sea-ice model without snow, in which the ice concentration is predicted using the method of Hibler (1979). In the land surface model, the surface temperature is predicted using the energy balance equation, while the hydrological cycle is simulated using the classic bucket model (Manabe 1969). The ice sheet model in the MPM-2 is the vertically integrated dynamic part of the 2D ice sheet model of Marshall & Clarke (1997), in which the ice sheet thickness is predicted by an ice mass conservation equation, the ice flow velocity is diagnosed from the ice height, and the bedrock depression is predicted from an isostatic adjustment mode. The MPM-2 was also interactively coupled to the dynamical vegetation continuous description model. This model is based on a continuous bioclimatic classification which provides the relative cover of trees, grasses, and potential deserts for each continent and latitude (Brovkin et al. 1997). The carbon cycle feedbacks are not included in this model.

For the vegetation map in the year 1990, we employed a reconstruction of LCC by Pongratz et al. (2008). In MPM-2, vegetation cover is described in terms of fractions of tree, grass, and desert, while this dataset specifies LCC in terms of crop, C3 and C4 pasture, and 11 natural vegetation types in global coverage at 30 minute resolution. For each year, a map is provided that contains 14 fields. Each field holds the fraction the respective vegetation type covers in the total grid cell 0–1. These data are aggregated to the coarse spatial resolution of MPM-2. For one grid cell, the grass fraction is interpreted as the sum of crop, grass, and pasture fraction, the sum of the rest of the vegetation fraction is forest fraction, and the remaining fraction of the grid is desert. The forest fraction in 1990 is given in Figure 2.

To detect any remote effect of afforestation, we devised two sets of experiments (Table 1). First, we evaluated the model response to ‘true world’ land cover representations in 1990 (control experiment (EC)). Second, to evaluate the potential future impact of regional afforestation on climate, we conducted simulations with the USA, China, and low latitudes in southern hemisphere (SH) land grids 100% of forest in the 1990 map, respectively (afforestation experiment (EA)). To identify the potential influence of regional afforestation on climate, we subtract EC from EA, which are, respectively, AUSA, ACHN, and ALSH. To minimize the effect of CO2 and other variabilities, the model is run from 1990 to 2100 after a spin-up time of 5,300 years under present-day orbital forcing, constant insolation of 1,365 W/m2, absence of volcanic eruptions, and CO2 concentration of 280 ppm. In all simulations, atmosphere, ocean, sea ice, and vegetation components are interactive, accounting for both physical and chemical feedbacks. This study generally employs extreme scenarios of afforestation with the aim to illustrate climate–vegetation interactions and compare the potential impacts of regional afforestation.

RESULTS

Afforestation has become a popular instrument of climate mitigation policy, as forests are known to sequestrate large quantities of carbon. In order to reveal the biogeophysical effects of different regional afforestation, the results are described in two sections. The first section compares the three different simulations for afforestation using constant pre-industrial atmospheric conditions. Analysis for this section focuses on the differences in the surface air temperature and sea ice area. The second section of the results analyzes the differences in zonally averaged temperatures and meridional oceanic heat transport to identify some potential impacts of afforestation on climate.

Results from all simulations for future surface air temperature and sea ice area relative to the control run, are presented in Table 2. The expansion of forest causes an increase in global, NH, and SH surface air temperature. These warmings are presumably due to an increased forest cover and the positive vegetation-albedo feedback. The regional replacement of current vegetation by trees would produce an average warming of 0.03 to 0.06 °C based on different forest expansion. The increase of forest area in the USA leads to a maximum global temperature generally, although it has the smallest conversion area for afforestation. The increase in temperature due to afforestation results in a decrease in sea ice area. The larger increase in temperature causes a larger decrease in sea ice area. The
maximum decrease in NH sea ice area is in simulation AUSA and ALSH. Although the change of NH temperature in simulation ACHN is a warming of 0.05 °C, the change of NH sea ice area is increasing which is probably due to the sea ice-albedo feedback (Ganopolski et al. 2001).

Afforestation results in widespread regional warming of the near surface atmosphere (Figure 3). We find that the range of responses clearly varies depending on regional afforestation. We also have observed varying temperature increases at other latitudes, far away from the afforested areas, in response to the afforestation. It is possibly ascribed to the changes over the afforested areas which affect energy-moisture transport into other latitudes. In terms of the forest expansion in the USA, prominent warming fluctuates with the seasons with a maximum warming of 0.11 °C over 60 N annually. A maximum warming of 0.14 °C around 45 N also occurs

Table 1  | Simulations’ description

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Domain</th>
<th>Conversion</th>
<th>Simulation</th>
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</thead>
<tbody>
<tr>
<td>EC</td>
<td>Global</td>
<td></td>
<td>True land cover change in 1990</td>
</tr>
<tr>
<td>EUSA</td>
<td>74° W–130° W; 23° N–54° N</td>
<td>6.13</td>
<td>Afforestation in USA</td>
</tr>
<tr>
<td>ECHN</td>
<td>73° E–135° E; 3° N–53° N</td>
<td>7.62</td>
<td>Afforestation in China</td>
</tr>
<tr>
<td>ELSH</td>
<td>0°–30° S</td>
<td>15.9</td>
<td>Afforestation in low latitudes in SH</td>
</tr>
</tbody>
</table>

Table 2  | Changes in annual mean global, NH, SH surface air temperature and sea ice area in 2100

<table>
<thead>
<tr>
<th></th>
<th>Global-sat</th>
<th>Sat-north</th>
<th>Sat-south</th>
<th>NHmean</th>
<th>SHmean</th>
</tr>
</thead>
<tbody>
<tr>
<td>AUSA</td>
<td>0.06</td>
<td>0.08</td>
<td>0.03</td>
<td>−0.01</td>
<td>−0.03</td>
</tr>
<tr>
<td>ACHN</td>
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<td>0.05</td>
<td>0</td>
<td>0.02</td>
<td>−0.01</td>
</tr>
<tr>
<td>ALSH</td>
<td>0.05</td>
<td>0.04</td>
<td>0.06</td>
<td>−0.01</td>
<td>−0.08</td>
</tr>
</tbody>
</table>

*aGlobal, surface air temperature (°C).
*bNH surface air temperature (°C).
*cSH surface air temperature (°C).
*dNH sea ice area (10⁶ km²).
*eSH sea ice area (10⁶ km²).

*Conversion area of desert and grass to forest (10⁶ km²).
during DJF (winter). The strongest regional response of 0.22 °C occurs in simulation ACHN around 45° N during JJA (summer). The noticeable warming around 70° N during SON (winter) and DJF (winter) caused by afforestation in China is not due to direct forcing but rather because of changes in SSTs and sea ice cover (Brovkin et al. 2006). The forest expansion in low latitudes of SH leads to a significant warming around 25° S with temperature increases of up to 0.18 °C during MAM (spring). Such significant noticeable warming is probably due to the biggest conversion area for afforestation shown in Table 1. These warmings reduce the sea-ice extent in northern high latitudes, which further increases the surface air temperature. These warmings can be explained by surface albedo, evapotranspiration efficiency, and surface roughness in response to afforestation. Replacing crops or pasture with forests increases roughness length, root depth, leaf area index, etc., and decreases albedo. These tend to increase the efficiency of transpiration and canopy evaporation although increased canopy cover can decrease soil evaporation. The net change in total evaporation due to LCC is therefore uncertain and depends on complex interactions between these components and the atmosphere, exactly how each type of vegetation is parameterized in each model, and how strongly the land is coupled to the atmosphere (Pitman et al. 2009). The net effect of biogeophysical contribution to changes in global and regional temperatures is positive in response to regional afforestation.

The climate responses to afforestation are influenced by feedback with SSTs and sea ice, even thermohaline circulation of the ocean. Simulating a dynamic ocean response is important to account for feedbacks through SSTs and sea ice. To further identify the climate

Figure 3 | Zonally averaged changes in surface air temperature (°C) for different seasons in 2100.
mechanisms that are changed with regional afforestation, we analyzed the changes in mean meridional heat transport in global ocean. The responses of mean meridional oceanic heat transport to regional afforestations are shown in Figure 4. Based on our experiments, we find that the meridional oceanic heat transport experiences a pronounced decrease in response to afforestation in the USA and China but an increase in response to SH afforestation. Due to alterations in the heat transports between atmosphere and ocean caused by the differential warmings in response to afforestation, the changes in annual mean meridional oceanic heat transport are modified. The most pronounced warming in simulation AUSA leads to a most obvious decrease in annual mean meridional oceanic heat transport in most parts of the world. The decrease in annual mean meridional oceanic heat transport is largest over the equatorial region and diminishes poleward. It appears that afforestation leads globally to a surface air warming (Figure 3). This warming is associated with a pronounced increase in NH water vapor content. Warmer and wetter air in the troposphere, in turn, means more longwave radiation transmitted from the atmosphere to the ocean surface. Consequently, more energy is absorbed at the ocean surface, thus leading to an increase in SSTs. Increased SSTs cause a lower density of the surface water which leads to decreased thermohaline overturning and oceanic heat transport (Figure 5).

DISCUSSION

The afforestation in the USA leads to an increase of 0.06 °C in annual global mean temperature and a regional maximum warming of 0.14 °C around 45° N during winter. This warming in this paper in response to afforestation is in line with some other results. Betts (2000) and Govindasamy et al. (2003) show that afforestation in northern temperate and boreal regions leads to a decrease in land surface albedo and consequent warming. Bonan (1997), using land surface model (LSM1.0) with the NCAR Community Climate Model (CCM2), finds that forests’ replacement of the eastern USA with crops would lead to a cooling of 1 °C over the eastern USA and a warming of 1 °C over the western USA in spring, as well as a summer cooling of up to 2 °C over a wide region of the central USA. The spread between the results in this study and Bonan’s (1997) estimates could be that their simulations were conducted with CCM2 and LSM1.0, which neglect the water vapor feedback over the sea surface and may reverse the sign of zonally averaged temperature changes.

The forest increase in China leads to a global warming of 0.03 °C annually and a strongest warming of 0.22 °C around 50° N during summer regionally. Gao et al. (2003) obtain a mean annual cooling in the range of -1 °C to 1 °C caused by land use changes. They show that the current land use causes temperature to significantly rise in the southern portions of northeast China, the Sichuan Basin, and portions of northwest China due to an increase of sensible heat flux over these regions associated with the reduction of vegetation cover and drying of the soil. This simulation is performed using a second-generation regional climate modeling system which limits feedbacks between the land and oceans; any
increase in land temperature is constrained by the effects of the infinite heat reservoir of the oceans. Our simulation, in contrast, in this study has been performed with all the important components of atmosphere, land, ocean, ice sheet, and terrestrial biosphere interacting with each other, which amplify the direct effect of LCC by positive feedback.

The forest expansion in low latitudes of the SH causes an increase in annual global mean temperature of 0.05 °C and a strongest warming of 0.18 °C around 25°S during spring. However, in the tropics, the warming effects in this paper in response to afforestation contradicts results of some AGCM simulations, which show a net cooling effect caused by evapotranspiration (Henderson-Sellers et al. 1993; Costa & Foley 2000; De Fries et al. 2002). Betts et al. (2007) even noted that reforestation or avoided deforestation in tropical regions could exert a double cooling effect through carbon sequestration and increased evaporation and cloud cover. The spread may be because most of the simulations were conducted with AGCMs with prescribed SSTs. The use of prescribed climatological SSTs can act to dampen the model interannual variability, which could increase or decrease the magnitude of changes due to afforestation (Hasler et al. 2009). Also, prescribed SSTs neglect the water vapor feedback over the sea surface and may reverse the sign of zonally averaged temperature changes (Ganopolski et al. 2001). Owing to there being fewer land-masses than in the NH, there are few studies covering the temperature change in the SH in detail.

A maximum regional warming of 0.14–0.22 °C resulted from the biogeophysical mechanism of future afforestation being obtained. Such warming is comparable with the biogeochemical effects of forest expansion, although a full assessment of biogeochemical effects of regional afforestation is beyond the scope of this study. In the reforestation scenario of Representative Concentration Pathways 4.5, Lawrence et al. (2012) simulate a draw down of 67.3 PgC from the atmosphere into the terrestrial ecosystem and product pools. The biogeochemical effect of historical land cover change resulting from increased emissions of carbon dioxide is a global temperature change of 0.3 °C (Matthews et al. 2004). In addition, Brovkin et al. (2004) demonstrate that biogeochemical mechanisms for LCC over the last 150 years act to warm the climate by 0.18 °C. However, a further comparison analysis of biogeophysical and biogeochemical effects of afforestation is essential for assessment of climate mitigation scenarios and regional climate adaptation (Brovkin et al. 2015).

The afforestation in the NH leads to a decrease in annual mean meridional oceanic heat transport. In response to afforestation, increased SSTs associated with surface warming contribute to a lower density of the surface water, thus decreasing thermohaline overturning and oceanic heat transport. This would lead to reduced SST and thus affect the surface air temperature further. This is in line with Brovkin et al. (2005).

We recognize several limitations in our results. First, the atmosphere model here is represented by a simple EMBM without detailed descriptions about atmospheric circulations and cloud dynamics, which may damp the water cycle and precipitation in our simulations. Second, we note that the simulation in our study does not involve cloud feedback which also affects the climate response. In response to afforestation, some AGCMs simulate a decreased cloud cover which would increase forcing by increasing the effect of changes in surface albedo on net radiative fluxes at the top of the atmosphere. Due to the absence of the cloud feedback, the climate response is likely to be influenced in the simulation. Again, our simulations only focus on the biogeophysical effects of regional afforestation on climate and the biogeochemical effect of afforestation (cooling due to absorbed carbon from the atmosphere) is not taken into account.

CONCLUSIONS

Afforestation programs have become increasingly prevalent around the world as trees are considered crucial in mitigating climate change due to their carbon sequestration potential. 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 biogeophysical processes due to afforestation tend to increase it by altering evapotranspiration, roughness, and surface albedo. In this study we addressed the regional biogeophysical impact of afforestation with an Earth system model of intermediate complexity, MPM-2. We contrasted the climate of ‘true world’ land cover representations with the climate resulting from the replacement of
actual vegetable type by forest in the USA, China, and low latitudes in the SH. Our main focus here was on the surface temperature and oceanic heat transport. The overall biogeophysical contribution to changes in global and regional temperatures is positive in response to afforestation. The afforestation in the USA leads to an increase in global annual mean temperature of 0.06 °C and a regional maximum warming of 0.14 °C around 45° N during winter. China experiences a global warming of 0.05 °C and a regional warming of up to 0.22 °C around 45° N during summer. The forest expansion in low latitudes of the SH leads to an increase in global mean annual temperature of 0.05 °C and a maximum regional increase of 0.18 °C around 25° S during spring. In addition, the afforestation in the USA and China causes a decrease in annual mean meridional oceanic heat transport but the afforestation in the SH leads to an increase. These results suggest that regional afforestation causes a differential global climate and even oceanic heat transport which should be paid more attention to. Also, considerations regarding the potential to increase the positive impact of afforestation are needed when forests are planted through identifying regions that result in greater mitigation of global warming.

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