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Incorporating changes in albedo in estimating the climate mitigation benefits of land use change projects

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Abstract

Some climate scientists are questioning whether the practice of converting of non-forest lands to forest land (afforestation or reforestation) is an effective climate change mitigation option. The discussion focuses particularly on areas where the new forest is primarily coniferous and there is significant amount of snow since the increased climate forcing due to the change in albedo may counteract the decreased climate forcing due to carbon dioxide removal.

In this paper, we develop a stand-based model that combines changes in surface albedo, solar radiation, latitude, cloud cover and carbon sequestration. As well, we develop a procedure to convert carbon stock changes to equivalent climatic forcing or climatic forcing to equivalent carbon stock changes. Using the model, we investigate the sensitivity of combined affects of changes in surface albedo and carbon stock changes to model parameters. The model is sensitive to amount of cloud, atmospheric absorption, timing of canopy closure, carbon sequestration rate among other factors. The sensitivity of the model is investigated at one Canadian site, and then the model is tested at numerous sites across Canada.

In general, we find that the change in albedo reduces the carbon sequestration benefits by approximately 30% over 100 years, but this is not drastic enough to suggest that one should not use afforestation or reforestation as a climate change mitigation option. This occurs because the forests grow in places where there is significant amount of cloud in winter. As well, variations in sequestration rate seem to be counterbalanced by the amount and timing of canopy closure.

We close by speculating that the effects of albedo may also be significant in locations at lower latitudes, where there are less clouds, and where there are extended dry seasons. These conditions make grasses light coloured and when irrigated crops, dark forests or other vegetation such as biofuels replace the grasses, the change in carbon stocks may not compensate for the darkening of the surface.

1 Introduction

The removal and storage of carbon dioxide from the atmosphere by photosynthesis (sequestration) as trees grow is considered a climate change mitigating activity. For this reason, there has been interest in converting non-forest lands into forests (afforestation or reforestation).

This practice has been questioned by some authors since there is often an associated change in surface albedo as a result of the land-use change. For example, Betts (2000) found that the change in surface albedo by the planting of coniferous forests in areas with snow can contribute significantly to the radiative forcing. Brovkin et al. (1999) found that cooling due to albedo change from deforestation was of the same order of magnitude as increased radiative forcing from CO_2 and solar irradiation. Bala et al. (2008) found that a global-scale deforestation event could have a net cooling influence on the Earth's climate.

Matthews et al. (2003) on the other hand suggest that the issue needs to be revisited with more realistic scenarios of transient land cover change. In a subsequent paper, Matthews et al. (2004) suggest that carbon emissions from land cover changes (deforestation) tend to exceed the cooling that results from change in surface albedo.

In this paper, we model the changes in radiative forcing due to changes in top-of-atmosphere albedo and carbon stocks over time due to a change in land-use from grasslands to forest in various locations and forest types in Canada. We present the combined impacts in terms of radiative forcing and equivalent changes in CO_2.

2 The Model

2.1 Carbon

To model the changes in carbon stocks over time we use the stand level carbon model, GORCAM (Schlamadinger and Marland, 1996). This is a simple model that tracks the flow of carbon from living above and below ground biomass to the dead wood, litter and soil pools. The model assumes that the living above ground biomass is driven by local growth curves or yield tables; the below ground living biomass is a function of the above ground biomass; and that annual non-living input is a fraction of the living biomass. The non-living biomass decays exponentially with decay rates based on temperature, rainfall and material (Moore et al., 1999; Trofymow et al., 2002). Some of the decaying material enters the soil pool, which also decays exponentially.

The carbon model, GORCAM, is similar to other carbon models CO2FIX and CBM3 in its approach, but is a simpler version that reduces the amount of parameters required for use in areas of sparse data.

2.2 Surface albedo

Albedo is a very complex function of surface and radiation characteristics, including land cover type, specifics of the vegetation, snow cover, soil moisture, incident angle, and wavelength (Henderson-Sellers and Wilson, 1983). Representations of albedo range from relatively simple functions of land cover type and snow (Henderson-Sellers and Wilson, 1983) to sophisticated models that include vegetation geometry and distribution and also multiple scattering and angle effects (Ni and Woodcock 1999, Yin, 1998).

In general, the major trends observed with surface albedo are that 1) albedo increases with snow cover, which is more reflective than foliage or soil, and that 2) forests, especially coniferous ones, have lower albedo than grass or crop covers, due to their denser canopies that trap more of the incident radiation (Robinson and Kukla, 1984; Sharratt, 1998)

For our model, we use Yin's model (Yin 1998) with modifications for variables that depend on site characteristics. Yin's model starts with a base albedo that is dependent on: stem cover; canopy cover; and life span and size of canopy leaves. The base albedo is then multiplied by factors that result from changes in: vegetation height; relative leaf age; seasonal biological stress; drought; and optical air mass. Finally the modified base albedo is blended with the snow albedo where the blending factor depends on snow presence, canopy closure and height.

We use a simplified form of Yin's full model that does not include variations in the seasonal biological stress or optical air mass. As well we base stem cover closure and height are simple function of tree biomass. Therefore:

$$W_i = \left(\frac{B_i \left[\mathrm{t\,ha}^{-1} \right]}{B_{\mathrm{closure}} \left[\mathrm{t\,ha}^{-1} \right]} \right)^{0.5} \quad for B \leq B_{\mathrm{closure}} \tag{1}$$

And

$$H_i\,[m] = \frac{H_{\max}\,[m]\,B_i \left[\mathrm{t\,ha}^{-1} \right]}{B_{\max} \left[\mathrm{tha}^{-1} \right]} \tag{2}$$

Where W_i is the stem cover in year i, B_i is the tree biomass in year i, B_{closure} is the tree biomass when closure is reached, H_i is the height of trees in year i, $H_{\max}$ is the maximum tree height. As well, Yin's snow factor is calculated using a snow presence as the fraction of the month that the site has snow depths greater than 5 cm. The surface albedo from January–May for a coniferous forest in Prince George, Canada as a function of age is shown in Fig. 1.

2.3 Top-of-atmosphere albedo

The climate forcing effect of changes in surface albedo is masked by clouds. As a result, the model must also include annual variations in cloud cover. We propose the following simplified model for the energy interaction of the surface with a cloud layer in between (Fig. 2).

The incident radiation, R, is partially reflected by the cloud layer by the amount α_c, partially absorbed by the amount A_b and an amount, τ_c is transmitted.

BGD

5, 1511–1543, 2008

At the surface, the transmitted radiation is partially reflected by the amount α_c. This value is the albedo of the vegetation.

On the return path, the energy from the surface is partially reflected back towards the surface by the clouds and partially transmitted by the same amounts as from above (α_c, τ_c). The re-reflected energy is reflected off the vegetation again and so on.

In general, the transmission and reflection coefficients will be frequency dependent, but for the remaining of the paper we will not include a frequency variable for ease of presentation.

The total reflected energy, E, is the sum of all returning energy:

$$E = R\left[\alpha_c + \tau_c\alpha_v\tau_c + \tau_c\alpha_v\alpha_c\alpha_v\tau_c + \tau_c\alpha_v\alpha_c\alpha_v\alpha_c\alpha_v\tau_c + \ldots\right] \quad (3)$$

Collecting terms:

$$E = R\left[\alpha_c + \frac{\tau_c^2}{\alpha_c}\left(\alpha_v\alpha_c + \alpha_v^2\alpha_c^2 + \alpha_v^3\alpha_c^3 + \ldots\right)\right] \quad (4)$$

Since

$\frac{1}{1-x} = 1 + x + x^2 + x^3 + \ldots + x^n$ for $|x| < 1$

The reflected energy becomes:

$$E = R\left[\alpha_c + \frac{\tau_c^2\alpha_v}{1-\alpha_c\alpha_v}\right] \quad (5)$$

And the effective albedo is

$$\alpha_{\text{eff}} = \frac{E}{R} = \left[\alpha_c + \frac{\tau_c^2\alpha_v}{1-\alpha_c\alpha_v}\right] \quad (6)$$

If we assume there is absorption by the atmosphere then $\tau_c = 1 - \alpha_c - A_b$.

BGD

5, 1511–1543, 2008

Albedo and sequestration in LULUCF projects

D. N. Bird et al.

The percentage of possible sunshine is a readily available measurement (for example; IWMI 2007). If we assume that cloudiness is related to the lack of sunshine and also that the reflectivity is proportional to cloudiness then:

$$\alpha_c = k_c(1 - S) \tag{7}$$

Where S=percentage of possible sunshine and k_c is a factor that controls the opacity or reflectivity of the clouds such that

$$0 \le k_c \le \frac{(1 - A_b)}{(1 - S)} \tag{8}$$

If $k_c = 0$ then the clouds are transparent. Substituting into Eq. (6):

$$\alpha_{\text{eff}} = \frac{k_c\,(1 - S)\,(1 - 2\alpha_v\,(1 - A_b)) + \alpha_v\,(1 - A_b)^2}{1 - k_c\alpha_v\,(1 - S)} \tag{9}$$

For typical values of sunnyness, $S\sim 0.4$, $A_b\sim 0.2$ (Valero et al, 2000) and $\tau_c\sim 0.52$ (Haigh, 2007), then k_c~0.5. Based on a least squares best-fit of the model to surface radiation measurements at eights sites across the United States, we find values of A_b and k_c of 0.25 and 0.65, respectively.

It is generally known that A_b varies with elevation, solar angle, air pollution, water vapour and dust particles. These factors tend to increase absorption and scattering by the atmosphere, thus decreasing the albedo further. For this discussion, we will assume that it is constant and investigate the sensitivity of the results to this assumption.

The components of the albedo for a 75 year old Lodgepole pine stand at Prince George, British Columbia are shown in Fig. 3. The vegetation changes little during the year. With snow in January, February, March, November and December, the surface albedo (vegetation + snow) is a little higher. Finally, when the effect of clouds is added, the effective albedo is increases for all months. The increase is greatest in the winter months when Prince George receives less than 25% of possible sunshine.

BGD

5, 1511–1543, 2008

The effective albedo of grass is also shown in this figure. A grassy field has on average albedo only 0.05 higher than a forest stand. The difference in albedo is greatest (0.08) in the winter due to the presence of snow on the grassy field.

A comparison of the surface and effective albedo of a forest with age in January in Prince George is given in Fig. 3.

It should be noted that Prince George has only 22% sunshine in the January (CCN, 2007).

3 Radiative forcing

3.1 Change in CO_2 concentration

In Betts (2000), he presents the annual radiative forcing as:

$$F_{CO_2}^{Ann}\left[Wm^{-2}\right] = B\left[Wm^{-2}\right]\ln\left(1+\frac{\Delta CO_2\,[ppmv]}{pCO_{2,ref}\,[ppmv]}\right) \tag{10}$$

Where,$pCO_{2,ref}$ = a reference partial pressure (383 ppmv), and from Betts (2000)

$$\Delta CO_2\,[ppmv] = \frac{\Delta CO_2\,[g]}{M_{CO_2}\left[gmole^{-1}\right]}\left(\frac{M_{air}\left[gmole^{-1}\right]}{m_{air}\,[g]}\right) * 1.0x10^6\,[ppmv] \tag{11}$$

Because the volume of a mole of gas is independent of the gas constitution. M_{CO2}=molecular mass of carbon (44.0095 g/mol), M_{air}=molecular mass of dry air (28.95 g/mol), m_{air}=mass of the atmosphere (5.148×10^{15} Mg) and ΔCO_2=the change in CO_2 (in grams) as a result of the reforestation project.

The constant, B, is calculated as

$$B\left[Wm^{-2}\right] = \frac{\Delta F_{2X}\left[Wm^{-2}\right]}{\ln(2)} \tag{12}$$

BGD

5, 1511–1543, 2008

Albedo and sequestration in LULUCF projects

D. N. Bird et al.

Where ΔF_{2X}=the radiative forcing per CO_2 doubling ($3.7\,Wm^{-2}$) (IPCC).

Combining equations:

$$F_{CO_2}^{Ann}\left[Wm^{-2}\right] = \frac{\Delta F_{2X}\left[Wm^{-2}\right]}{\ln(2)} \ln\left(1+\frac{1.0x10^6\,[ppmv]\,\Delta CO_2\,[g]\,M_{air}\left[gmole^{-1}\right]}{pCO_{2,ref}\,[ppmv]\,M_{CO_2}\left[gmole^{-1}\right]1.0x10^6 m_{air}\,[Mg]}\right) \quad (13)$$

As Caldiera and McCraken (2004) point out, Betts does not include the decay of CO_2 in his paper. Betts instead assumes that half the CO_2 remains airborne over forest growth time scales. The decay of CO_2 is given by:

$$\Delta pCO_2(t) = \Delta pCO_{2,init}\left(a_0 + \sum_{i=1}^{4} a_i e^{-t/\tau_i}\right) \quad (14)$$

Where $\Delta pCO_{2,init}$=the initial perturbation of atmospheric CO_2 content and a_i and τ_i are decay constants (Archer et al., 1997; Maier-Reimer and Hasselmann, 1987).

Including this term and assuming making a small scale approximation to ln(1+x),

$$F_{CO_2}^{Ann}(t)\left[Wm^{-2}\right] \approx \frac{\Delta F_{2X}\left[Wm^{-2}\right]}{\ln(2)} \left(\frac{1.0x10^6\,[ppmv]\,\Delta CO_2\,[g]\,M_{air}\left[gmole^{-1}\right]}{pCO_{2,ref}\,[ppmv]\,M_{CO_2}\left[gmole^{-1}\right]1.0x10^6 m_{air}\,[Mg]}\right)\left(a_0 + \sum_{i=1}^{4} a_i e^{-t/\tau_i}\right) \quad (15)$$

And for a project that removes CO_2 annually:

$$F_{CO_2}^{Ann}(t)\left[Wm^{-2}\right] \approx \frac{\Delta F_{2X}\left[Wm^{-2}\right]}{\ln(2)} \quad (16)$$

BGD

5, 1511–1543, 2008

Title Page
Abstract Introduction
Conclusions References
Tables Figures
◀| |▶
◀ ▶
Back Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

$$\left(\frac{1.0x10^{6}\,[\text{ppmv}]\,\Delta CO_2(t)\,[\text{g}]\,M_{\text{air}}\left[\text{gmole}^{-1}\right]}{pCO_{2,\text{ref}}\,[\text{ppmv}]\,M_{CO_2}\left[\text{gmole}^{-1}\right]\,1.0x10^{6}\,m_{\text{air}}\,[\text{Mg}]}\right)\otimes \text{Decay}^{\text{Ann}}_{CO_2}(t) \quad (17)$$

Where $\otimes$represents the convolution operation and $\text{Decay}^{\text{Ann}}_{CO_2}(t)$ is given by:

$$\text{Decay}^{\text{Ann}}_{CO_2}(t) = a_0 + \sum_{i=1}^{4} a_i e^{-t/\tau_i} \quad (18)$$

3.2 Change in albedo

The annual change in forcing due to a change in albedo at the top of the atmosphere (TOA) as:

$$F^{\text{Ann}}_{\alpha}(t)\left[\text{Wm}^{-2}\right] = \frac{-R\left[\text{Wm}^{-2}\right]A\left[\text{m}^2\right]\Delta\alpha(t)}{A_{\text{Earth}}\left[\text{m}^2\right]} \quad (19)$$

R=impinging radiation A=area of albedo change, A_{Earth}=the area of the earth (5.1×10^{14} m^2) and $\Delta\alpha$ albedo change.

This is a rather simplistic model since the albedo change and radiation both vary with season. With this consideration:

$$F^{\text{Ann}}_{\alpha}(t)\left[\text{Wm}^{-2}\right] = -\frac{A\left[\text{m}^2\right]}{12A_{\text{Earth}}\left[\text{m}^2\right]}\sum_{m=1}^{12} R_m\left[\text{Wm}^{-2}\right]\Delta\alpha_m(t) \quad (20)$$

Where m=month. It is important to note that one should be consistent. Either the radiation and albedo are both measured at the top of atmosphere (i.e. above the clouds), or the radiation and albedo are measured at the surface (i.e. below the clouds).

$$F^{\text{Ann}}_{\alpha}(t)\left[\text{Wm}^{-2}\right] = -\frac{A\left[\text{m}^2\right]}{12A_{\text{Earth}}\left[\text{m}^2\right]}\sum_{m=1}^{12} R_{\text{TOA},m}\left[\text{Wm}^{-2}\right]\Delta\alpha_{\text{eff,m}}(t) \quad (21)$$

Where $\Delta\alpha_{\mathrm{eff,m}}$ is the effective albedo change taking into account cloud cover and $R_{\mathrm{TOA,m}}$ is the monthly top-of-atmosphere radiation.

One can estimate $R_{\mathrm{TOA,m}}$ geometrical considerations, since the Earth receives 1360 W/m^2 of mean annual solar radiation, but this radiation varies with latitude.

The cross sectional area at a given latitude, λ, and longitude, ϕ, ignoring inclination of the Earth's axis is given by:

$$dA = (r\cos\lambda d\lambda)(r\cos\phi d\phi\cos\lambda) \tag{22}$$

Therefore the average radiation at given latitude is

$$\overline{R}\left[\mathrm{Wm^{-2}}\right] = \frac{\int_{\frac{-\pi}{2}}^{\frac{\pi}{2}} R_{\mathrm{Ann}}\left[\mathrm{Wm^{-2}}\right] r^2\left[\mathrm{m^2}\right](\cos\lambda)^2 d\lambda\cos\phi d\phi}{\int_{\pi}^{\pi} r^2\left[\mathrm{m^2}\right]\cos\lambda d\lambda d\phi} \tag{23}$$

Note: the radiation integral is evaluated from only over the sunlit regions.

$$\overline{R}\left[\mathrm{Wm^{-2}}\right] = \frac{R_{\mathrm{Ann}}\left[\mathrm{Wm^{-2}}\right]}{\pi}\cos\lambda \tag{24}$$

When, η, inclination of the earth's axis of rotation of the Earth's axis is included:

$$\lambda = \lambda + \eta\cos\left[\frac{(6.5-m)\pi}{6}\right] \quad 24 \tag{25}$$

$$R_{TOA,\lambda,m}\left[\mathrm{Wm^{-2}}\right] = \frac{R_{\mathrm{Ann}}\left[\mathrm{Wm^{-2}}\right]}{\pi}\cos\left(\lambda + \eta\cos\left[\frac{(6.5-m)\pi}{6}\right]\right) \tag{26}$$

Where m=month number

3.3 Total forcing

Finally, the change in forcing due to a change in land use is given by the sum of the individual forcing components above;

$$F^{\text{Ann}}_{\text{Total}}(t)\left[{}^{\text{W}}\!/\!_{\text{m}^2}\right] = F^{\text{Ann}}_{\text{CO}_2}(t)\left[{}^{\text{W}}\!/\!_{\text{m}^2}\right] + F^{\text{Ann}}_{\alpha}(t)\left[{}^{\text{W}}\!/\!_{\text{m}^2}\right] \tag{27}$$

3.4 CO_2 Equivalence

Reducing radiative forcing is not the item negotiated directly as part of the UNFCCC and the Kyoto Protocol. The climate change mitigation community uses CO_2-equivalence as its indicator. Fortunately, with a little mathematical reorganization of Eq. (14), one can also express changes in radiative forcing in terms of CO_2 equivalence.

$$\Delta \text{CO}_2 eq_{\alpha}(t)[g] \approx \frac{F^{\text{Ann}}_{\text{CO}_2}(t)\left[\text{Wm}^{-2}\right]\ln(2)}{\Delta F_{2X}\left[\text{Wm}^{-2}\right]} \left(\frac{\text{pCO}_{2,\text{ref}}[\text{ppmv}]\, M_{\text{CO}_2}\left[\text{gmole}^{-1}\right] 1.0x10^6 m_{\text{air}}[\text{Mg}]}{1.0x10^6[\text{ppmv}]\, M_{\text{air}}\left[\text{gmole}^{-1}\right]}\right) \otimes \text{InvDecay}^{\text{Ann}}_{\text{CO}_2}(t) \tag{28}$$

Where $\text{InvDecay}^{\text{Ann}}_{\text{CO}_2}(t)$ is the inverse-filter of $\text{Decay}^{\text{Ann}}_{\text{CO}_2}(t)$ so that

$$\text{Decay}^{\text{Ann}}_{\text{CO}_2}(t) \otimes \text{InvDecay}^{\text{Ann}}_{\text{CO}_2}(t) = 1 \tag{29}$$

$\text{InvDecay}^{\text{Ann}}_{\text{CO}_2}(t)$ can be calculated analytically since we are modelling only changes in albedo (i.e., no change before the start of the project).

The form of $\text{Decay}^{\text{Ann}}_{\text{CO}_2}(t)$ and $\text{InvDecay}^{\text{Ann}}_{\text{CO}_2}(t)$ is shown in Fig. 5

BGD

5, 1511–1543, 2008

Albedo and sequestration in LULUCF projects

D. N. Bird et al.

4 An example from Prince George, Canada

The result for a Lodgepole Pine stand (site index=20) in Prince George, B.C., Canada is shown in Fig. 6. The growth model is from TIPSY growth and yield model used by the Government of British Columbia (Ministry of Forests and Range, 2007). Climatic information (% of possible sunshine, depth of snow) is taken from Environment Canada (2006). Initially, during the first 35 years, the albedo change creates a positive forcing, but after this stage the carbon sequestration exceeds the albedo effect by at maximum a factor of 3. As a result, the change in albedo reduces the combined forcing roughly 30%. It should be noted that Prince George is a region with very cloudy winters, some snow and good tree growth for Canada, so that an average stand grows with a Mean Average Increment of 3.3 t/ha/year.

As shown in Fig. 7, the change in albedo is equivalent to an emission of roughly 39 t CO_2e during the first year. After this time, the carbon sequestration removes more CO_2 than the equivalent emission caused by the change in albedo, but the later continues to create an equivalent annual emission of around 2 t CO_2e.

The annual emissions are a complicated way to view the effects caused by the change in albedo. One can also display the cumulative emissions (Fig. 8). Here one clearly sees that the change in albedo causes equivalent emissions that are compensated by the sequestration after 40 years. As well, the albedo equivalent emissions reduce the sequestration by 45% after 200 years.

4.1 Sensitivity

We investigate the sensitivity of the cumulative emissions to atmospheric (atmospheric absorption, cloud opacity) and tree parameters (growth and canopy closure) in the model. Atmospheric parameters control the amount of energy reaching the surface while the tree parameters control the amount of sequestration and energy absorbed by the surface.

Figure 9 shows the sensitivity of the modelled results to cloud opacity. When the cloud are transparent, the opacity=0 and the change in surface albedo, because of the darker surface in winter, has a large effect. The albedo change dominates until the trees have a chance to grow rigourously. As the clouds become more reflective, the effect of the change in albedo is less and the sequestration is the dominating factor. For reasonable value of cloud opacity, there is a period of small positive emissions at the onset of the reforestation, but in the long term the reforestation creates a net emission removal of approximately 800 t CO_2e/ha.

Figure 10 displays the sensitivity of the modelled results to atmospheric absorption. Without absorption, the albedo change has the most impact since more energy reaches the surface. As more energy is absorbed by the atmosphere, less energy reaches the surface and the benefit of carbon sequestration increases.

In Fig. 11, the sensitivity to tree dynamics are shown. Tree dynamics control, the rate of sequestration, the total amount of sequestration and the age at which canopy closure is reached. The last factor affects the surface albedo. Using the TIPSY model, which produces both biomass and canopy closure data, we have modelled a range of stand qualities (SI=site index=tree height at 50 years). We see that in all cases except for the poorest stands (SI=5), there is a net benefit of planting trees. In the poorest stand, there are only small net emissions. This result is because at lower site indexes, the canopy closure never closes. The trees sequester less but they are also not so dark.

5 Extension to various regions of Canada

We have used the model to evaluate the combined surface albedo with correction and carbon sequestration for various sites across Canada. The models are based on provincial growth curves for the dominant forest species, and long-term measurements of amount of sunshine at various sites across Canada. Unfortunately, most growth curves do not include an estimate of canopy, so we have assumed canopy closure occurs when the annual biomass increment begins to decrease.

BGD
5, 1511–1543, 2008

Title Page
Abstract Introduction
Conclusions References
Tables Figures
◂ ▸
Back Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

The net cumulative emissions from the modelled stands after 50 years and 100 years are shown in Fig. 12 and Fig. 13, respectively. The variation at 50 years age is caused by the variation in average growth rate at young ages. By 100 years, all stands have roughly the same cumulative net emissions (−400 t CO_2e/ha). Sites with higher albedo component, due to a sunnier climate and/or more southerly latitude, are also sites with increased sequestration.

A model of a birch forest in south eastern Quebec is also included. The site has the lowest albedo component, as one would expect since the there are no leaves in winter, but also the slowest growth over 100 years.

6 Conclusions

In conclusion, stand scale modelling suggests that the change in combined greenhouse gas mitigation benefit due to afforestation/reforestation projects is reduced due to the increase in albedo during months with substantial snow. This result is in agreement with other authors (Betts, 2000; Bala et al., 2008).

Where we differ from other authors is in the suggestion that afforestation / reforestation in areas with snow for many months has a positive radiative forcing or an equivalent greenhouse gas emission. As shown in Fig. 12 and Fig. 13 all Canadian sites analyzed showed negative net cumulative emissions as a result of an afforestation/reforestation project. The change in surface albedo is counterbalanced by two factors, sun angle and clouds. The change in albedo occurs at times of the year where the sun angle is lowest, but the dominant factor that reduces the effectiveness of the change in albedo is the inclusion of clouds. In many cases, the snowy time of the year coincides with the cloudiest time of the year. As a result, even though a snow covered field has a higher albedo than forest there is less direct sunlight that reaches the surface. It is the effective albedo at the top of atmosphere that is significant.

BGD

5, 1511–1543, 2008

7 Discussion

This work is an attempt to model the climate change effects of afforestation / reforestation projects. There are still many areas of research required to properly understand the affects of changing albedo. These include:

1. An improved top of atmosphere albedo model that includes cloud cover (based on MODIS data);
2. Albedo effects due to incident angle; and
3. Timing of crown closure at maturity.

As well, this is only a partial model since it is well known that increasing tree cover increases evapotranspiration. This will dissipate the increases in energy caused by a decrease in albedo. Modelling of these effects is beyond the scope of a locally based model. Nevertheless, a locally-based model shows that the change in albedo reduces the climate mitigation benefit of a land-use change that increases carbon sequestration. But, in the example shown, the decrease is not sufficient to suggest that afforestation / reforestation should not be considered as a climate change mitigation activity. As a corollary to this observation, it would be interesting to model the net cumulative emissions in low latitude areas with lots of sun. These areas tend to be savannas and grasslands which can have high albedo in periods of drought. We speculate that afforestation with pines or other dark vegetation and potentially irrigation of cropland in areas that were savannas and grasslands should also be investigated for their net cumulative emission profile (both sequestration and forcing due to albedo change). This has also been suggested by Field et al. (2008). For some types of vegetation such as Jatropha, the darkening of the land surface may not be compensated by the increases in sequestration.

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BGD

5, 1511–1543, 2008

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BGD
5, 1511–1543, 2008

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
◀| |▶
◀ ▶
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

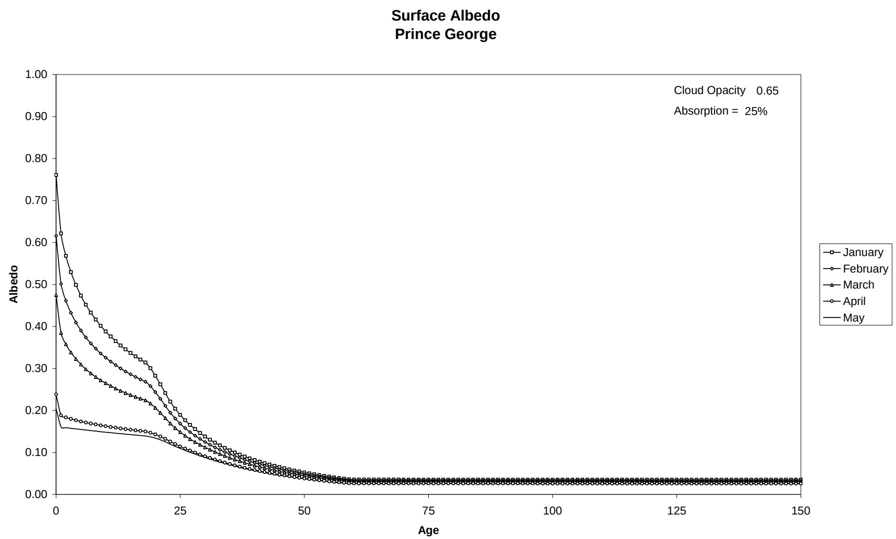

Fig. 1. Surface albedo by month and age–Prince George, Canada. Notes: The change in slope of the albedo curves at 20 years is caused by a change in growth rate of the trees as predicted by the yield tables. We have assumed a constant annual biomass increment to 20 years age.

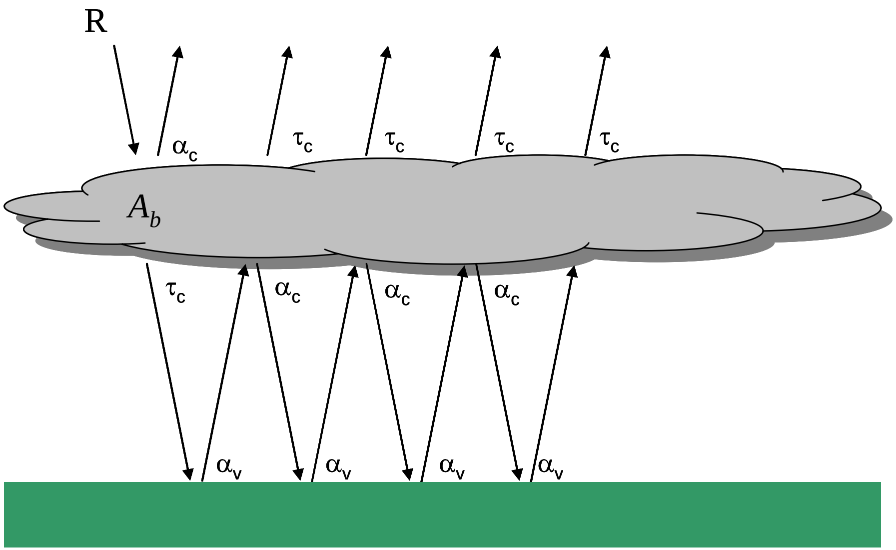

Fig. 2. Schematic of cloud scattering.

BGD
5, 1511–1543, 2008

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
◀◀ | ▶▶
◀ | ▶
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

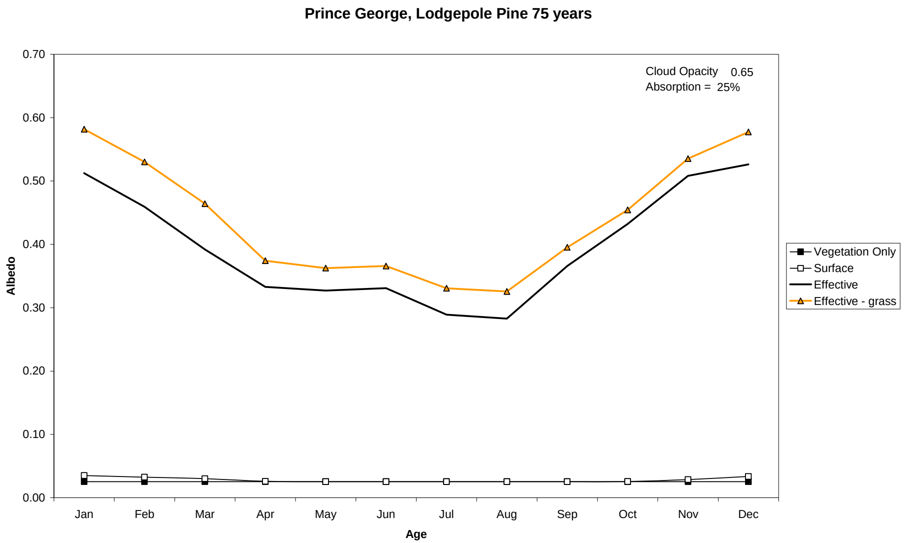

Fig. 3. Components of albedo–75 year old Lodgepole Pine Stand (Si=20), Prince George, B.C., Canada.

BGD
5, 1511–1543, 2008

Title Page
Abstract
Introduction
Conclusions
References
Tables
Figures
◀I
▶I
◀
▶
Back
Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

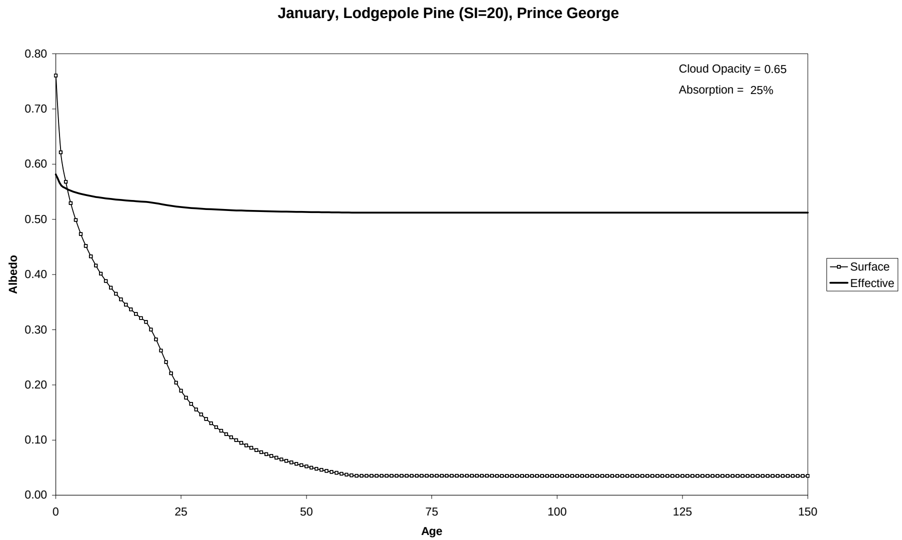

Fig. 4. A comparison of surface and effective albedo with age–January, Prince George , B.C., Canada.

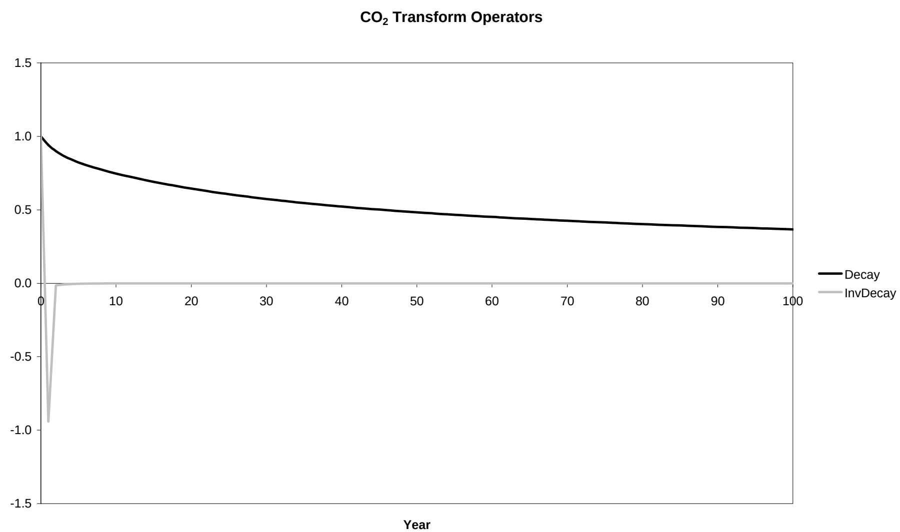

Fig. 5. The CO_2 decay and inverse operators.

BGD
5, 1511–1543, 2008

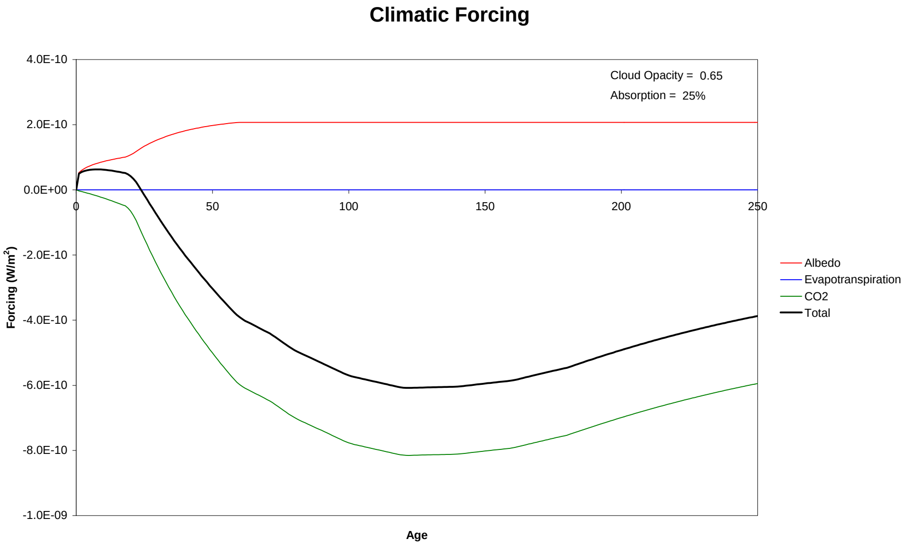

Fig. 6. Climatic forcing–Lodgepole Pine stand (SI=20), Prince George, B.C., Canada.

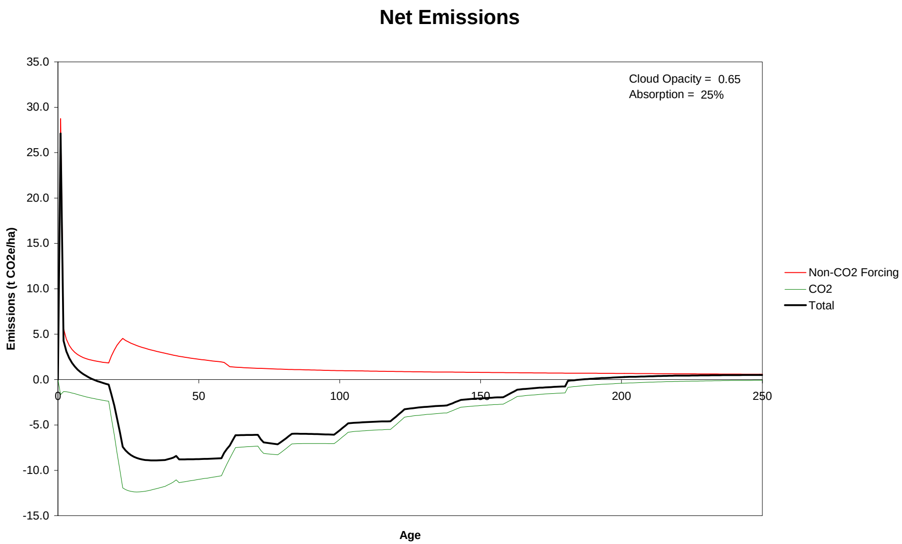

Fig. 7. Net emissions including CO_2e from albedo change–Lodgepole Pine stand (SI=20), Prince George, B.C., Canada.

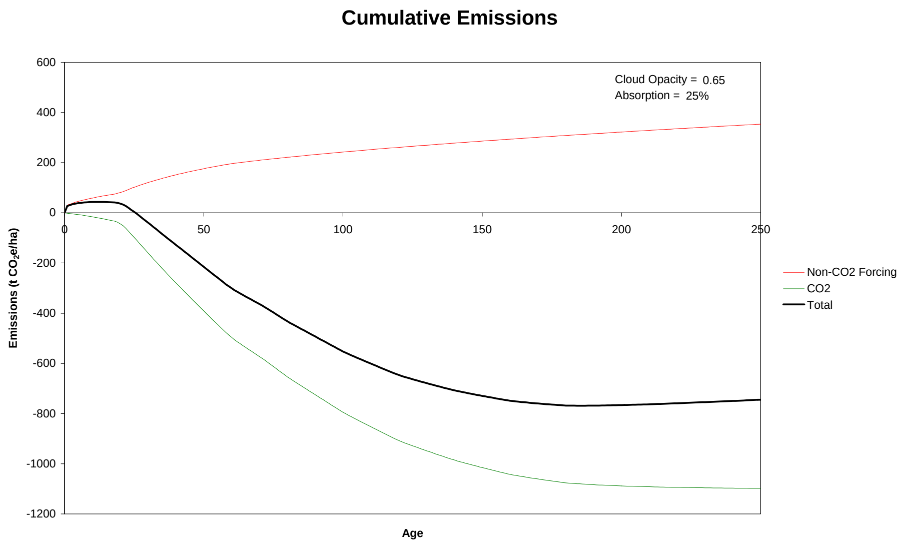

Fig. 8. Cumulative net emissions including CO_2e from albedo change–Lodgepole Pine stand (SI=20), Prince George, B.C., Canada.

BGD

5, 1511–1543, 2008

Albedo and sequestration in LULUCF projects

D. N. Bird et al.

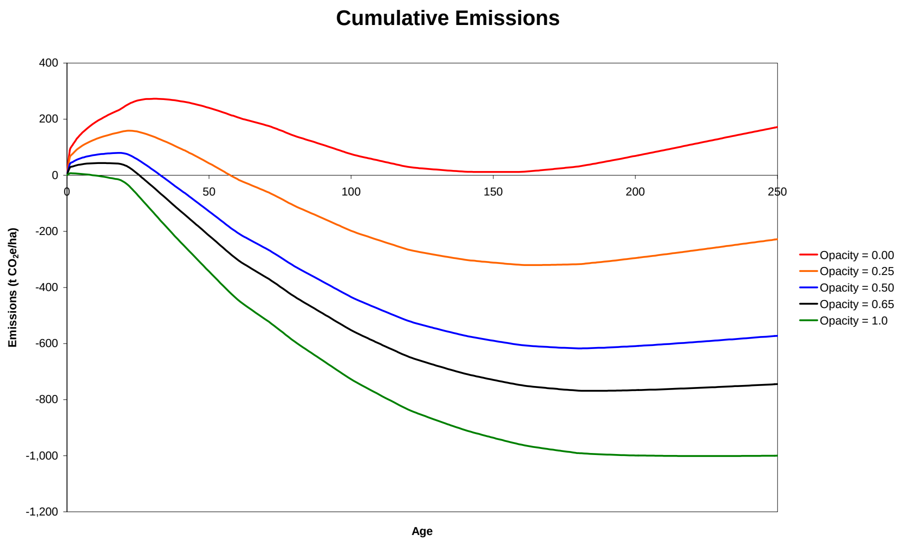

Fig. 9. Sensitivity of cloud opacity on cumulated net emissions–Lodgepole Pine stand (SI=20), Prince George, B.C., Canada.

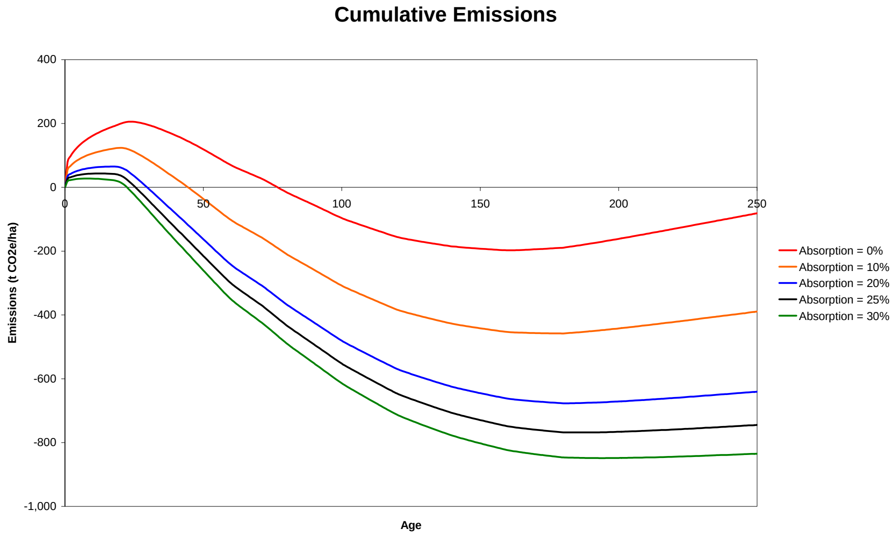

Fig. 10. Sensitivity of absorption on cumulated net emissions–Lodgepole Pine stand (SI=20), Prince George, B.C., Canada.

BGD

5, 1511–1543, 2008

Albedo and sequestration in LULUCF projects

D. N. Bird et al.

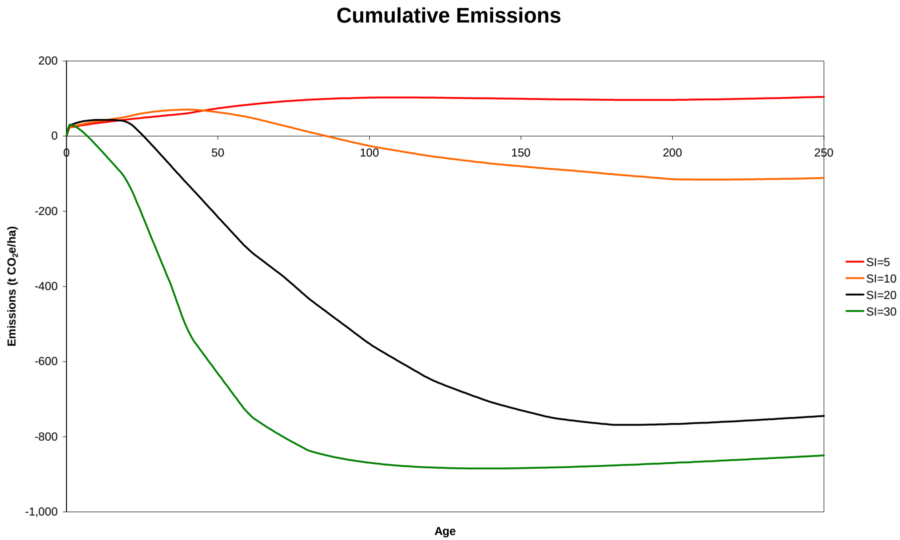

Fig. 11. Sensitivity of tree dynamics on cumulated net emissions–Lodgepole Pine stand (opacity=0.65, absorption=25%), Prince George, B.C., Canada

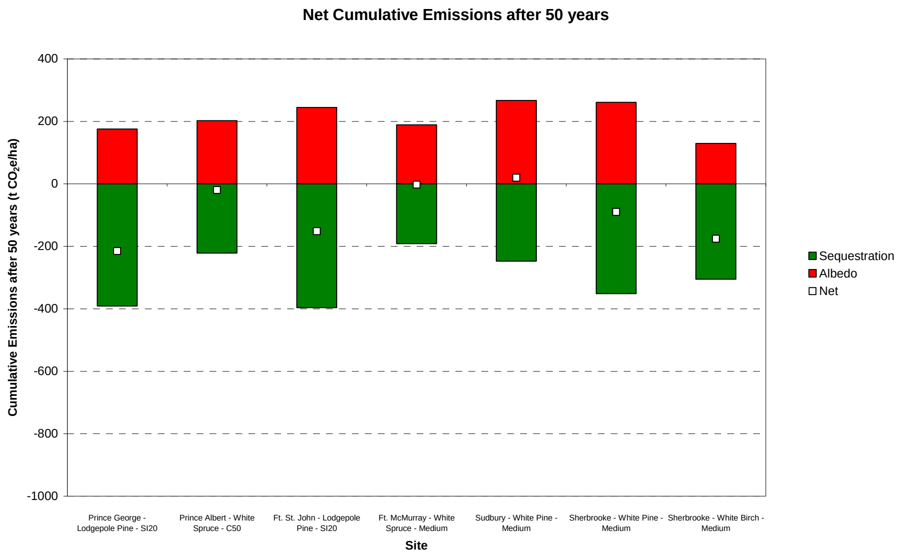

Fig. 12. Examples from across Canada–After 50 years.

BGD

5, 1511–1543, 2008

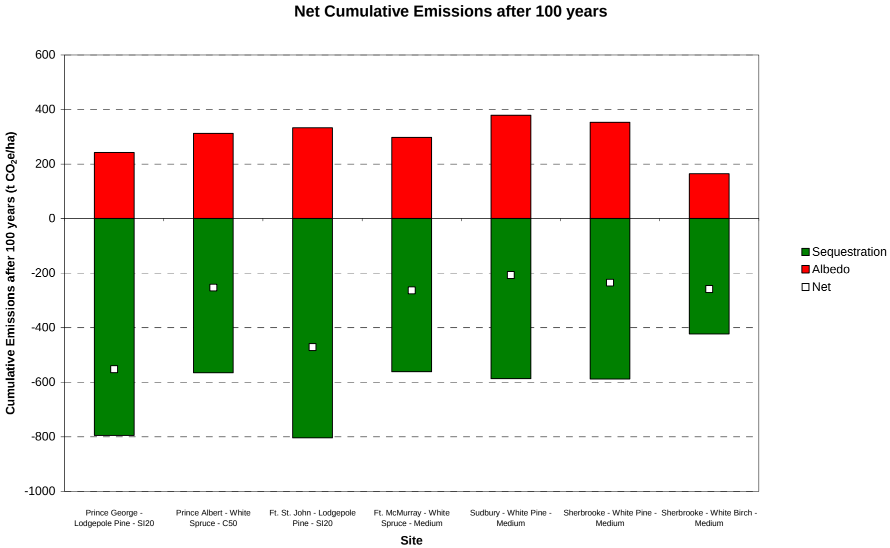

Fig. 13. Examples from across Canada–After 100 years.