Assessing the influence of canopy snow parameterizations on snow albedo feedback in boreal forest regions

by

Chad William Thackeray

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AUTHOR'S DECLARATION

I hereby declare that I am the sole author of this thesis. This is a true copy of the thesis, including any required final revisions, as accepted by my examiners.

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Abstract

Variation in snow albedo feedback (SAF) among CMIP5 climate models has been shown to explain much of the variation in projected 21st Century warming over Northern Hemisphere land. Prior studies using observations and models have demonstrated both considerable spread in the albedo, and a weak bias in the simulated strength of SAF, over snow-covered boreal forests. Boreal evergreen needleleaf forests are capable of intercepting snowfall throughout the snow season, which has a significant impact on seasonal albedo. Two satellite data products and tower-based observations of albedo are compared with simulations from multiple configurations of the Community Climate System Model (CCSM4) to investigate the causes of weak simulated SAF over the boreal forest. The largest bias occurs in April-May when simulated SAF is one-half the strength of SAF in observations. This is traced to two canopy snow parameterizations in the land model. First, there is no mechanism for the dynamic removal of snow from the canopy when temperatures are below freezing, which results in albedo values in midwinter that are biased high. Second, when temperatures do rise above freezing, all snow on the canopy is melted instantaneously, which results in an unrealistically early transition from a snow-covered to a snow-free canopy. These processes combine to produce large differences between simulated and observed monthly albedo, and are the sources of the weak bias in SAF. This analysis highlights the importance of canopy snow parameterizations for simulating the hemispheric scale climate response to surface albedo perturbations. A number of new experiments are described as recommendations for future work.

Acknowledgements

First, I would like to thank Dr. Chris Fletcher and Dr. Chris Derksen for their continuous support of my research and for giving me the opportunity to work with them. I look forward to continuing on with our collaboration over the next few years. I would also like to acknowledge my thesis readers Dr. Richard Kelly and Dr. Claude Duguay for readily agreeing to participate and provide insightful feedback. Lastly, I would like to express the deepest appreciation to my Mom, Dad and siblings (Jen and Ryan) for their encouragement and unwavering support of my academic pursuits. Thank you Jen and Ryan for setting a wonderful example throughout the years with your own educational success.

Preface

This thesis contains a submitted journal article that investigates the impact of canopy parameterizations on snow albedo feedback. This paper has been submitted to *The Journal of Geophysical Research: Atmospheres*. Collaboration with colleagues helped in the submission of the paper, as credited below. The article (Chapter 2) is the result of collaboration with Dr. Christopher Fletcher and Dr. Chris Derksen. Both of whom provided a great deal of support throughout the study. My contribution was to carry out all analysis and write the first edition of the manuscript, and editing subsequent drafts from the comments and writing of Dr. Christopher Fletcher and Dr. Chris Derksen.

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Chapter 1

Introduction

1.1 Background

Seasonal snow in the Northern Hemisphere has a strong influence on hydrological and climate processes. The reflective nature of snow in the visible spectrum (400 nm – 700 nm) means that it can have a strong influence on the surface energy budget. Freshly fallen snow has an extremely high albedo (reflectivity of solar radiation), reaching up to 0.90, and slowly becoming less reflective with the aging process (time after snowfall) to about 0.45-0.50 during melt (Wiscombe and Warren, 1980). Much of this albedo loss can be attributed to the rapid changes in snow grain size and shape that occur as a result of temperature and liquid water content variations, also called snow metamorphism (Colbeck, 1982). Its low thermal conductivity also makes snow an efficient insulator, which moderates soil temperatures and influences permafrost extent (Lawrence and Slater, 2010; Brutel-Vuilmet et al., 2013; Vaughan et al., 2013).

Snow also has a great hydrological importance, making up a significant portion (17%) of terrestrial water storage in nonpolar cold climate regions (Gunthner et al., 2007). The snowpack that remains throughout the winter stores a large amount of water, as much as a third of annual precipitation, and this is mainly released during the spring melt (Bartlett et al., 2006). In open environments, the buildup of a snowpack on the ground is strongly tied to aspect, or the direction a slope faces, due to its direct linkages with incoming solar radiation (Varhola et al., 2010). More snow tends collect on northerly aspects because of less melting

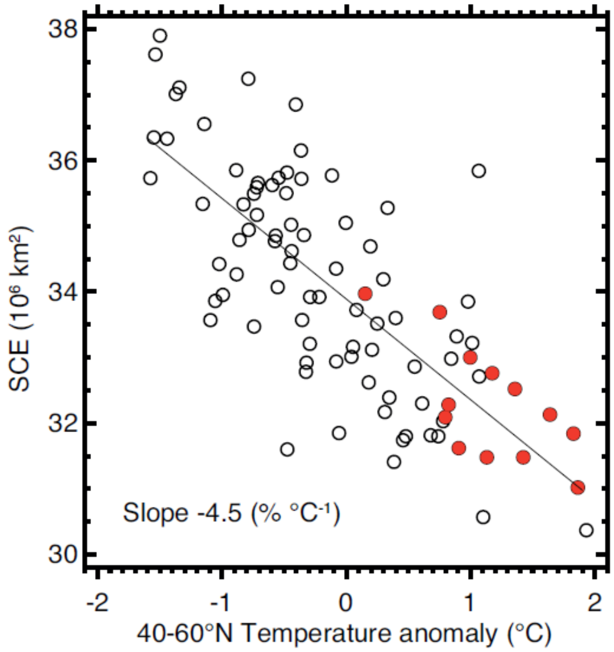

Figure 1-1: Relationship between NH April SCE and corresponding land air temperature anomalies over 40 to 60 from CRUTEM4. Red circles indicate 2000-2012. Updated from Brown and Robinson (2011), (Vaughan et al., 2013).

and sublimation as a result of decreased exposure to incident energy (Golding and Swanson, 1986). Snowfall accumulation patterns naturally vary from year-to-year depending on climatic conditions, but on average approximately 40% of Northern Hemisphere land is snow covered at its peak extent (Hall, 1988; Robinson and Frei, 2000; Qu and Hall, 2005). However, with recent warming, a sharp decrease has been shown in snow cover extent (SCE) during spring (Brown et al., 2010; Brown and Robinson, 2011; Figure 1-1). These snow cover losses over the observational record have been shown to be more drastic at higher latitudes because of stronger albedo feedbacks (Dery and Brown, 2007; Vaughan et al., 2013). There are also changes expected to affect snow season length, which is projected to shorten through both a later fall accumulation and earlier spring snow melt (Lawrence and Slater, 2010).

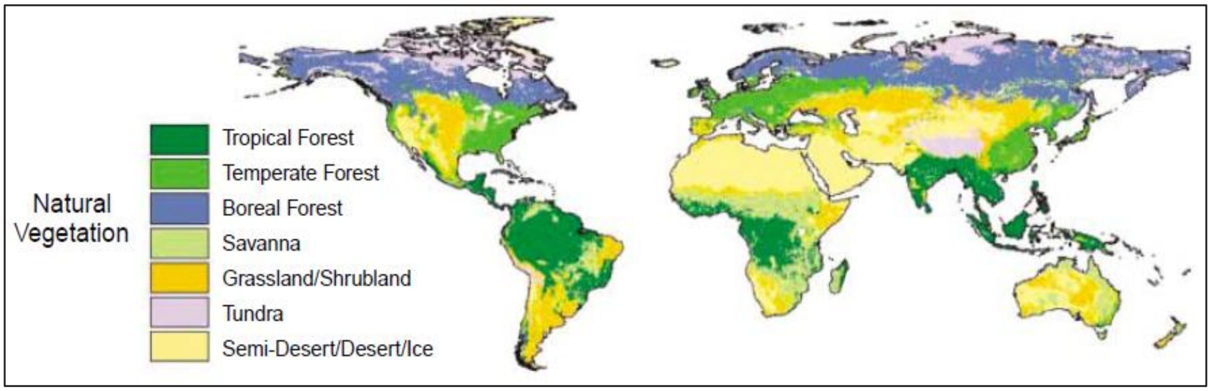

Figure 1-2: Global natural vegetation map (Foley et al., 2005).

A significant portion of the Northern Hemisphere snow-covered area is forested, creating an environment where complex interactions occur. The boreal forest occupies 11.6 x 10^6 km^2 (Bonan et al., 1992; Figure 1-2), and for much of the region, snow is on the ground for more than six months annually. This landscape is dominated by three forest types, needleleaf evergreen, needleleaf deciduous, and broadleaf deciduous. The region is crucial in masking snow albedo, and northward movement of the tree line into tundra may produce further warming (Bonan et al., 1992). It is through this masking of reflective fresh snow that more incoming radiation is absorbed at the surface, resulting in a warming effect on climate. A circumpolar band of forest can be clearly identified during winter by its low albedo (~0.30; Thomas and Rowntree, 1992; Barlage et al., 2005; Essery, 2013; Figure 1-3). The boreal biome has been shown to have the greatest biogeophysical impact on global temperatures of any landcover type (Snyder et al., 2004; Bonan, 2008). Snyder et al. (2004) found that with the removal of this land classification (converted to bare ground) there was a large increase in simulated annual-average surface albedo (0.26 or 156%), decreases in net radiation (14 Wm^{-2} or 27%), and in turn an annual-average temperature decrease of 2.8 °C over the region.

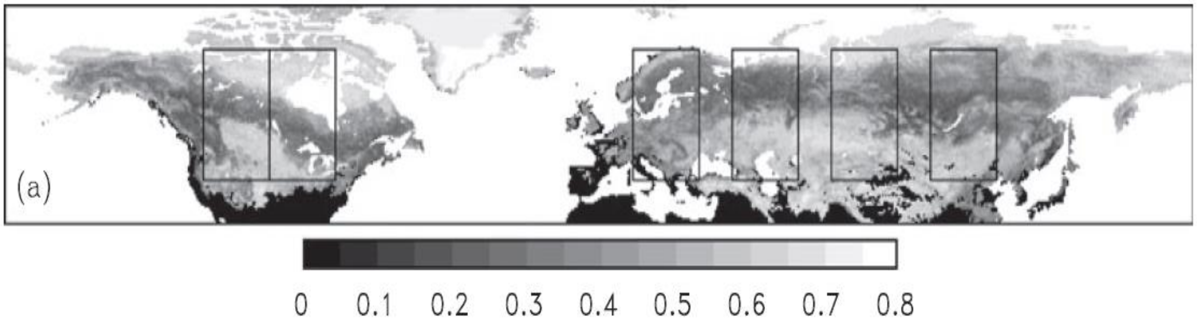

Figure 1-3: Average albedo for land with snow cover from MODIS. Black pixels have missing data or no observed snow cover for 2006-2010 (Essery, 2013).

1.1.1 Forest Snow Processes

Snowfall over forest stands is divided into canopy interception and throughfall to the surface (Hedstrom and Pomeroy, 1998; Storck et al., 2002; Rutter et al., 2009). Sturm et al. (1995) classified this unique forest snowpack as Taiga, and described it as a 30-120 cm deep, low-density cold snow cover. By spring, the pack is typically about 50-80% depth hoar with fresh snow on top, and more than 15 layers in total (Sturm et al., 1995). Since evergreen species retain their needles year-round, they can efficiently intercept snow. Prior research has found that between 40-60% of cumulative snowfall is intercepted over the boreal forest during mid-winter (Pomeroy and Schmidt, 1993; Storck et al., 2002). It has also been suggested that land beneath canopies only receives approximately 60% as much snow as that of open grasslands at the Boreal Ecosystem Research and Monitoring Sites (BERMS; Hardy et al., 1997). Thus, forest cover can drastically alter snow accumulation and ablation processes, changing how snow processes are occurring at the surface (Pomeroy et al., 1998b; Andreadis et al., 2009; Essery et al., 2012).

1.1.1.1 Canopy Interception

The interception of snowfall by forest canopy is one of the primary processes affecting the snow regime in this environment. Interception rates are controlled by a number of factors including canopy morphology, air temperature and wind speed (Miller, 1964). In general, snow accumulation at the surface decreases with an increasing forest cover because of increased canopy interception (Essery et al., 2003; Varhola et al., 2010). Once snow is intercepted, it can either sublimate, melt within the canopy, or unload from the canopy (Pomeroy et al., 1998a; Rutter et al., 2009; Andreadis et al., 2009).

It is fairly difficult to measure the latent, radiative and sensible heat fluxes in a snow-covered canopy because of the large contrast of surface temperature, albedo and roughness of snow and the conifer canopy (Pomeroy et al., 1998a). The complex system of a boreal forest in winter contains advective flows of energy as a result of the strong energy difference between the ground snow and an unloaded canopy (Pomeroy et al., 1998b). Furthermore, depending on the structure of the forest, there are various degrees of snow interception possible. Pomeroy et al. (1998b) found that snow interception by black spruce species is greater than that of both a pine forest and a mixed spruce/aspen stand in the southern boreal forest (Figure 1-4). The seasonal evolution of the snow interception as a percentage of cumulative snowfall also varies by type. The decrease in mid-winter is the result of the canopy reaching its finite holding capacity for snow, allowing for more snowfall to directly reach the surface, while the percentage at the end of the winter is equal to the seasonal sublimation loss (Pomeroy et al., 1998b)

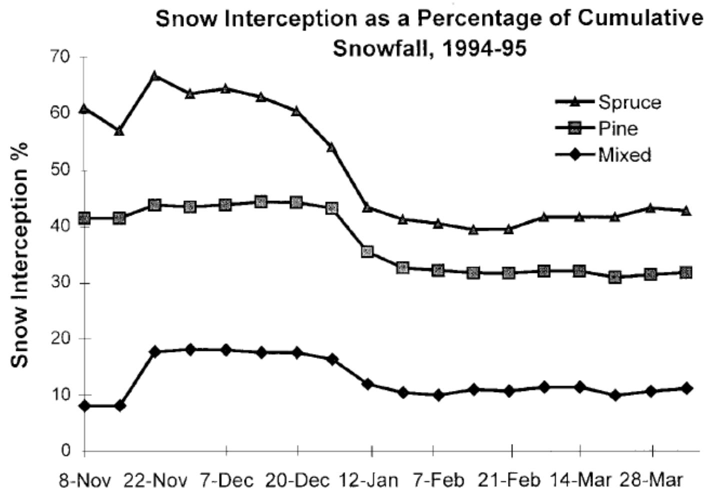

Figure 1-4: Snow interception at three forest stands as a percentage of cumulative snowfall in the southern boreal forest. Interception is estimated as above-canopy – below-canopy snowfall (Pomeroy et al., 1998b).

Snow that is intercepted by the canopy does not usually stay bonded to branches for extended periods because of strong winds and large incoming shortwave radiation (Betts and Ball, 1997). The duration of snow on canopy following a precipitation event becomes shorter as spring progresses. By April, snow only covers branches sporadically and tends to disappear quickly after snowfall (Kuusinen et al., 2012). However, under certain conditions high canopy snow retention is likely; they include well-below freezing temperatures, weak winds, and cloud coverage (Kuusinen et al., 2012). It has also been suggested that numerous small precipitation events can lead to greater interception than one large snowfall event in which a large proportion of snow falls to the surface because the canopy cannot hold any more (Varhola et al., 2010).

On the other hand, the proportion of snow that does make it to the forest floor is very different from snowpacks that can be found in grasslands for example. Snow beneath a canopy is sheltered from most wind and incoming solar radiation, however, it may be exposed to greater longwave radiation when the canopy is snow free (Rutter et al., 2009). This increased thermal radiation from the canopy helps to partially offset the lack of turbulence and shortwave radiation influencing the pack (Harding et al., 2001; Sicart et al., 2004). Despite this increased longwave flux, it has been shown that the rate of snowmelt within a forest can be almost 70% lower than that of an open grassland because of shading (Boon, 2007; Teti, 2008). All of these factors throughout the system impact the energy and mass balances at ground level. Thus, the proper representation of a forest canopy within a model is critical for accurate estimates of snow growth and loss (Pomeroy et al., 2002; Talbot et al., 2006). The snow under these forested stands is also much more dense because of compaction as a result of mass release from the canopy along with melt water drip, further exemplifying the impact of interception (Lundberg and Halldin, 2001; Bartlett et al., 2006).

1.1.1.2 Canopy Sublimation

The processes by which snow is directly converted to water vapor have been measured and modelled to gain a better understanding of their complex nature. Canopy snow has a large exposed surface area, and a significant portion of annual snowfall over boreal forests sublimates before ever reaching the forest floor (Essery et al., 2003; Schmidt and Troendle, 1992; Pomeroy and Gray, 1995; Lundberg and Halldin, 2001). Losses to sublimation tend to peak at around 30-40% of annual snowfall in conifer forests (Pomeroy and Schmidt, 1993). These substantial sublimation rates are the result of forests being aerodynamically rough, meaning they support turbulent exchange of mass and energy (Pomeroy et al., 1998b; Bartlett et al., 2006). This process is largely driven by the net available energy, with net losses typical of forested areas having a tendency to reduce the energy available for melting (Essery et al., 2008; Varhola et al., 2010).

There are numerous approaches to measuring snow ablation processes, a common technique used to compute interception rates and sublimation is to weigh trees after snowfall (Lundberg, 1993; Montesi et al., 2004). However, Molotch et al. (2007) pointed out that it is difficult to apply a measurement of individual trees to the stand scale. Also, there is some subjective analysis required to differentiate between mass release of snow and sublimation, while trace snowfall can introduce more uncertainty. Measurements and estimations of maximum sublimation rates from a snow-covered canopy have been quantified as 0.25-0.3 mm/hr, with a long-term average of about 0.1 mm/hr (Lundberg and Halldin, 1994; Harding and Pomeroy, 1996; Nakai et al., 1999; Pomeroy et al., 1998b). More recently, Lundberg and Halldin (2001) found sublimation rates of between 1.3 and 3.9 mm/day. This sublimation from the canopy is closely tied to the length of time snow remains in the canopy. Sublimation from beneath the canopy was previously thought to be minimal because of the sheltering effects of a forest stand (low wind speeds and shading). However, when the canopy is warm and snow-free there is the potential for large longwave radiation fluxes (Molotch et al., 2007; Woo and Giesbrecht, 2000). This sub-canopy sublimation was found to account for much more of the total sublimation rates after successive days without snowfall, because the canopy would have lost a majority of its intercepted snow. However, over the snow season, sublimation from the canopy is the primary contributor to total sublimation (Molotch et al., 2007; Figure 1-5).

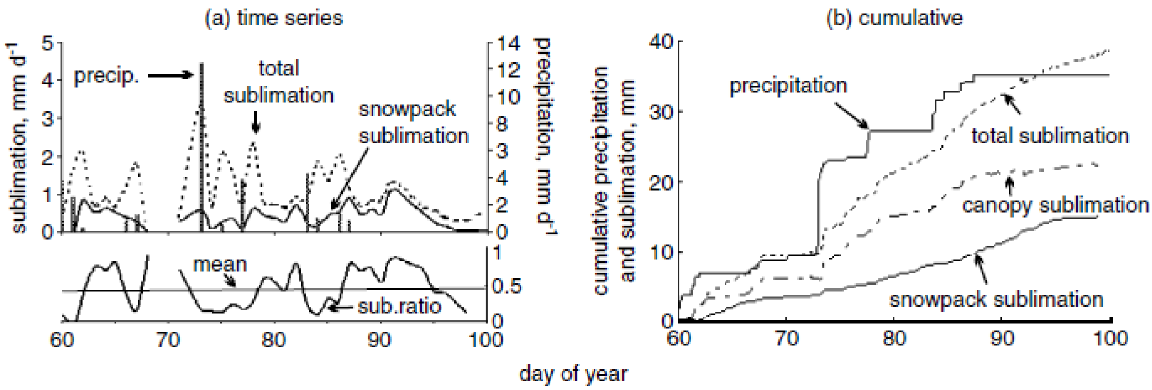

Figure 1-5: a) Time series of daily average sublimation measurements. Precipitation and the ratio of snowpack sublimation to total sublimation are shown. b) Cumulative sublimation from the snowpack, intercepted snow (canopy), total sublimation and precipitation throughout the study period (Molotch et al., 2007).

Less common is the mass release of snow from a forest canopy. This process is the result of a combination of wind perturbation and melting influences (Andreadis et al., 2009). As the adhesiveness of snow to tree branches becomes greater, as is common during snowfall and at temperatures slightly below freezing, removal of snow due to wind becomes less likely. When melt occurs, the bonds between snow and canopy are broken, allowing for mass release of snow (Andreadis et al., 2009). Increased temperature also causes decreased branch stiffness, contributing to the likelihood of snow unloading (Schmidt and Pomeroy, 1990).

1.1.1.3 Snow properties and forest albedo

Boreal evergreen forests are characterized by high seasonal energy flux variability as a result of their northerly location. In the snow-free season, approximately 90% of incident radiation is absorbed at the surface or the canopy level over this land cover type, whereas in winter, the presence of snow causes the albedo to vary greatly (Moody et al., 2007; Kuusinen et al., 2012). Snow albedo is influenced by a number of variables, but primarily properties related to snowpack structure. Snow grain size, liquid water content, snow depth, and snow impurities all have been shown to impact albedo (Wiscombe and Warren, 1980; Colbeck, 1982; Warren, 1984; Doherty et al., 2010). New snow grains tend to have a radius of approximately 50-100 µm, whereas old melting snow has a radius of about 1000 µm, with this metamorphism proceeding slowly unless a large temperature gradient between the surface and atmosphere is present (Colbeck, 1982; Doherty et al., 2010). This increased grain size makes it more likely that incident radiation can become absorbed by particles because of the longer ice grain path that must be crossed (Warren et al., 1998).

The dirtying of snow can occur rapidly in forested complexes because of leaf litter deposition onto the pack (Wiscombe and Warren, 1980). Other contaminants such as dust and soot settle on snow packs as well, thus darkening the upper snow layer. These impurities to the snow surface absorb radiation and warm the surrounding snow, thus increasing grain size and liquid water content, while decreasing albedo (Warren, 1984; Flanner et al., 2007).

Forested landscapes also tend to be less reflective than non-forested land in winter because much of a forest's dark green vegetation remains exposed, while bare land becomes completely covered by white snow. A snow-free canopy is also less reflective than when snow is present in the canopy layer. However, even when large amounts of snow reside on the canopy, scattering of incident radiation within the canopy vegetation reduces albedo significantly from the value of fresh snow on a non-vegetated ground surface (Betts, 2000; Harding and Pomeroy, 1996). Kuusinen et al. (2012) measured the albedo over a Pine stand in Finland, and found that the influence of canopy snow on the albedo was fairly large. During January-March when the canopy was snow free, there was an average daily albedo of

0.188, whereas the mean daily albedo was 0.314 when the canopy was snow covered. This quantifies the impact of intercepted snow as a 67% increase in albedo for this particular forest (Kuusinen et al., 2012). Prior research using satellite observations has revealed a wide range of values for snow-covered albedo over boreal evergreen forests throughout the snow season, from a Nov-Dec-Jan (NDJ) mean of 0.21 (Jin et al., 2002) to a peak value of between 0.26 and 0.34 (Barlage et al., 2005; Essery, 2013; Kuusinen et al., 2013), before dropping to approximately 0.23 in April (Loranty et al., 2014). Although albedo values at the upper end of this range result from contributions of snow on the ground through canopy gaps (Barlage et al., 2005). This marks a significant decrease from the albedo of fresh snow, even when a large amount of snow is held in the canopy (Essery et al., 2013).

1.1.2 Modelling of snow

The modelling of snow processes has been the subject of much attention in recent years, due to a better understanding of their importance in climate change through feedbacks (Flato et al., 2013). There is a hierarchy of general circulation models with varying degrees of complexity, which simulate natural processes in different ways. When used in a multi-model ensemble framework they allow for more reliable projections than using a single model (Bohn et al., 2010). In order to compute mass and energy fluxes, models need parameterizations of mechanisms determining the albedo, thermal conductivity, snow cover fraction, density of snow and physical exchanges between the ground and atmosphere (Essery et al., 2013). Many of the same parameterizations are commonly used in different models allowing them to be grouped, and implying that they are not all independent (Essery et al., 2013). Despite this fact, various model intercomparison studies have shown that simulations of snow vary greatly in their accuracy by model (maximum normalized root mean squared errors ranged from 2.1 to 6.9 for SWE over forested sites; Rutter et al., 2009; Essery et al., 2013). There are usually a few different ways in which various snow processes and properties are represented in snow models. Some surface schemes represent a snowpack as one or more layers above the soil surface (explicit; Oleson et al., 2010), whereas other models alter the characteristics of soil at the surface to reflect the properties of snow cover (composite; Boone et al., 2004; Bartlett et al., 2006). There are also differences in how models prescribe the albedo of snow, with some permitting albedo to evolve with density variations, while others keep the albedo constant regardless of snowfall totals (Slater et al., 2001; Bartlett et al., 2006; Essery et al., 2009). The compaction of snow over time due to metamorphism is critical for calculation of snow depth from snow mass (Essery et al., 2013). This process can be represented through a physical compaction parameterization (CLM4), an empirical parameterization (CLASS), or through the use of a constant snow density that is somewhere between that of fresh and compacted snow (Bartlett et al., 2006; Oleson et al., 2010; Essery et al., 2013). Snow cover fraction (SCF) parameterizations tend to be empirical or conceptual, often fitted by snow cover depletion curves that relate a given snow depth to SCF (Essery et al., 2013).

Parameterizations of forest snow processes on the other hand are fairly basic in many models because of our limited understanding of both physical exchanges within the canopy and between different layers (ground, canopy, and atmosphere). Notably, there are many different ways in which the vegetation masking effect on surface albedo can be parameterized; classified by Qu and Hall (2007) as four types among the CMIP3 models with varying complexity. Type 1 models utilize a full canopy radiative transfer model, with ground albedo calculated as the weighted mean of soil and snow albedo as determined by snow cover fraction. Type 2 schemes prescribe vegetation albedo for each plant functional type (PFT) and then modify albedo with the presence of snow on the canopy. In these models, surface albedo is calculated with weights determined by the vegetation gap fraction. The simpler type 3 and type 4 parameterization schemes do not separate canopy and ground albedo, but rather calculate it based on the ratio of snow-covered albedo to snow-free land albedo. The only difference between these two being that for type 3 schemes, snow albedo varies by vegetation type. Snow albedo parameterizations also vary among the CMIP3 models, with snow albedo metamorphism dependent on either snow age or temperature (Qu and Hall, 2007). Simulations of snow depth or SWE throughout a winter can diverge between models depending on how each model deals with canopy interception, and unloading processes. As well as how the melt and sublimation rates are computed and how this water is then drained from the snowpack (discharged or absorbed and refrozen to the pack; Rutter et al., 2009).

Various snow model intercomparisons have concluded that there is no one model that consistently outperforms the rest, especially when multiple locations are taken into consideration (Etchevers et al., 2004; Rutter et al., 2009; Essery et al., 2012). Results from these studies of snow models have shown that intermodel variance becomes greater during melt conditions, implying that simulating snow is more difficult for locations where temperatures tend to hover around 0°C for extended periods (Etchevers et al., 2004; Rutter et al., 2009; Figure 1-6).

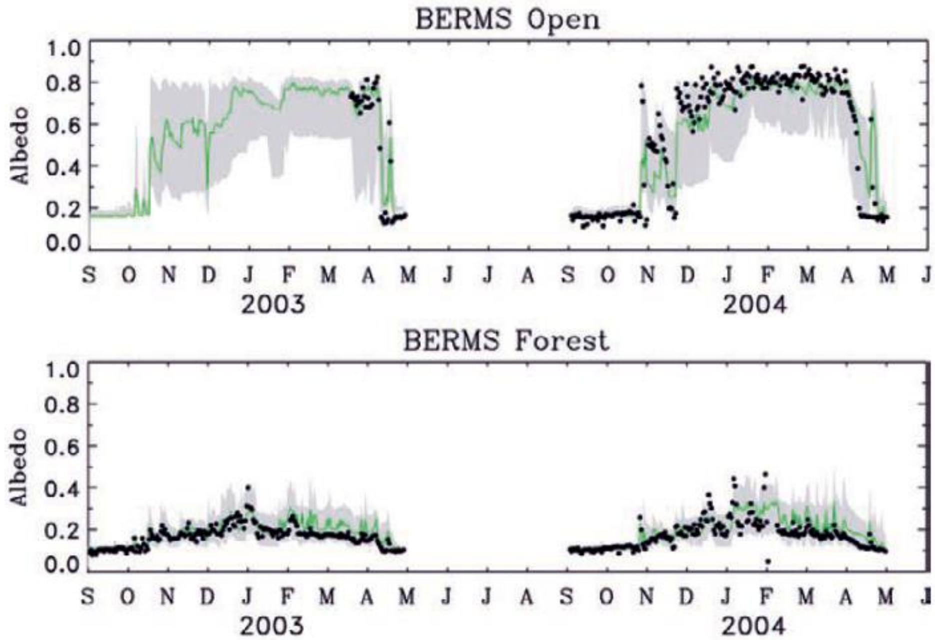

Figure 1-6 Observed and modeled albedo at an open site and a forested site. Observations (black dots), model medians (green lines), and interquartile ranges (gray bands) are shown (Essery et al., 2009).

Rutter et al. (2009) found that warming events when temperatures were above freezing for more than two days resulted in the greatest divergence between models. These difficulties are also associated with mixed precipitation events at these temperatures because the models have different techniques for partitioning between snow and rain (Rutter et al., 2009; Essery et al., 2012).

The model in focus here is the Community Climate System Model, version 4.0 (CCSM4), and in particular its land component, the Community Land Model, version 4.0 (CLM4). CLM4 uses the SNICAR (Snow and Ice Aerosol Radiation) model to simulate snow albedo (Flanner and Zender, 2006; Oleson et al., 2010). A two-stream radiative transfer solution produces radiative fluxes at each snow layer, with the albedo of this snow reliant on solar zenith angle, the underlying surface, snow depth, concentration of impurities, and the

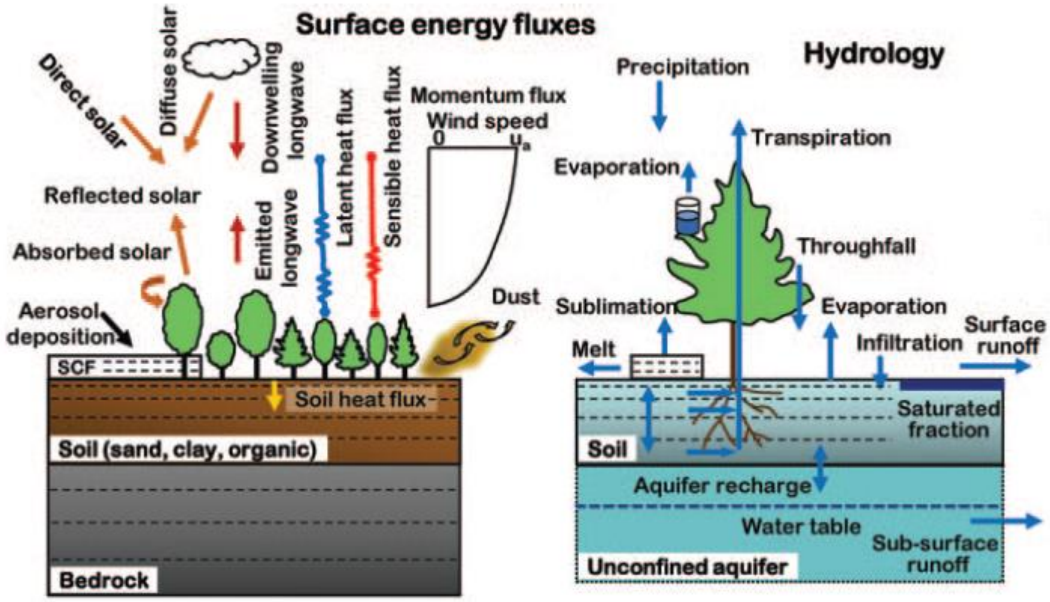

Figure 1-7 Schematic representation of the primary processes and functionality in the CLM4 (Lawrence et al., 2011).

effective grain size, which evolves with a snow aging parameterization (Lawrence et al., 2011). It is a multi-layer representation of snow that increases in segments as the snow depth grows, up to a maximum of five layers (Essery et al., 2012; Figure 1-7). Further details on this model and its parameterizations will be discussed in Section 2.3.1.

There is difficulty in attempting to accurately model certain aspects of this complex system, due to a shortage of reliable observations, and an imperfect knowledge of arctic and sub-arctic climate processes (MacKay et al., 2006; Bartlett et al., 2006). It is not an easy task to simulate snowpack dynamics, because what starts as a single layer snowfall of consistent grain size and density, becomes a multi-layered object containing ice lenses, wind crusts, and great variability in grain size (MacKay et al., 2006). Progress in the understanding of these processes, and interactions between varying levels of a forest are essential for future development of the models, as this is an area where discrepancy between models is greatest

(Rutter et al., 2009). The Coupled Model Intercomparison Project phase 5 (CMIP5) models have been shown to underestimate the recent negative trend in spring snow cover (Derksen and Brown, 2012), largely due to a cool bias over the boreal land surface (Brutel-Vuilmet et al., 2013). One factor that is not commonly represented in GCMs is the transport of snow as a result of wind, which has a tendency to remove a significant portion of snow cover from open land (Essery et al., 2003). Even in the most complex land surface schemes, there are difficulties associated with representing leaf litter on snow, and melt wells around trees, which act to decrease albedo (Bartlett et al., 2006; Woo and Giesbrecht, 2000). However, dramatic improvements in the modelling of these environments have occurred over the past two decades (Rutter et al., 2009), making it evident that as our understanding of the complex processes at work gets better, modelling improvements will follow. One such example is that seasonal biases in the representation of snow cover have been improved by a revised parameterization in CLM4 (Niu and Yang, 2007). The bias in annual mean snow cover area was reduced by ~ four million km^2 from CLM3 to CLM4 (19.6 million km^2 NH average; Lawrence et al., 2012). The accurate representation of albedo and snow cover is crucial because of mechanisms that allow for land alterations to affect global climate sensitivity, such as snow albedo feedback.

1.1.3 Snow Albedo Feedback (SAF)

It has been suggested that snow albedo changes during the last 20 years of the last century contribute to a large amount of the warming seen over the Northern Hemisphere (Groisman et al., 1994; Qu and Hall, 2006). SAF is a positive feedback climate mechanism, whereby an increasing air temperature causes snow to retreat and reveal a more absorbent land surface,

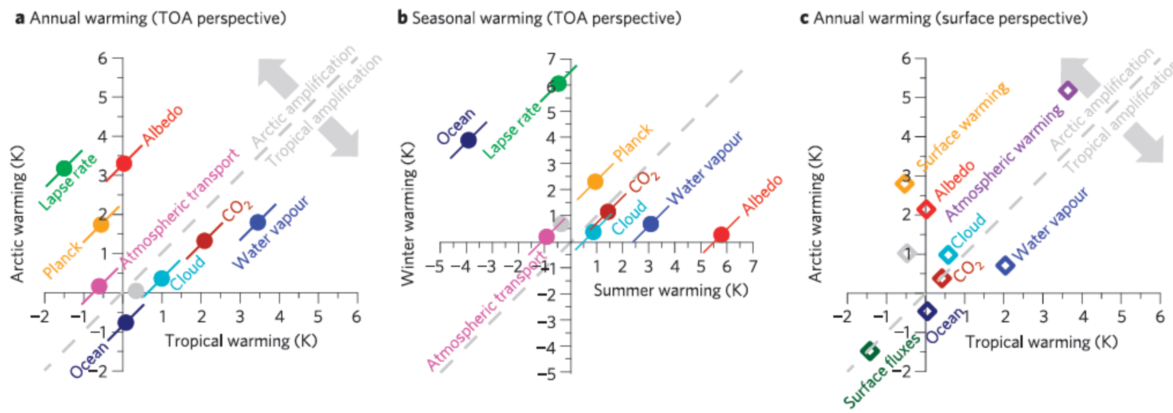

Figure 1-8: Warming contributions of individual feedback mechanisms. (a) Arctic versus tropical warming from a TOA perspective (b) Arctic winter versus summer warming (c) Arctic versus tropical warming from a surface perspective. For a,c, the 1:1 line shows whether the feedback contributes to or opposes Arctic amplification. Grey is the residual error of the decomposition. 'Ocean' includes the effect of ocean transport changes and ocean heat uptake (Pithan and Mauritsen, 2014).

thus resulting in further warming. It is one of many climate feedbacks in action within the climate system, and it has been shown to have a major impact on Arctic warming (Pithan and Mauritsen, 2014). The contribution of surface albedo feedback (made up of both SAF and ice-albedo-feedback) has been calculated in a decomposition of recent warming as +5.7 K over the Arctic in summer (Pithan and Mauritsen, 2014), thus making it the second largest contributor to Arctic amplification after the lapse rate feedback (Figure 1-8).

The strength of this feedback is calculated as the change in net shortwave radiation (or percent change in surface albedo) per degree of surface temperature change (Cess and Potter, 1988; Fletcher et al., 2012). There are however, large variations in the observed strength of this feedback amongst previous studies (Cess et al., 1991; Holland and Bitz, 2003; Flanner et al., 2011). A spread in total SAF strength also exists between models, with a range among CMIP5 models of -0.63 $Wm^{-2}K^{-1}$ to -1.52 $Wm^{-2}K^{-1}$ and an ensemble-mean of -1.08 $Wm^{-2}K^{-1}$ (Andrews et al., 2012). The variation between models is largely a result of each land scheme parameterizing snow processes differently. This large spread in SAF strength amongst CMIP5 models has been shown to account for much of the intermodel spread for future projections of warming over Northern Hemisphere land areas (Qu and Hall, 2013).

In general, SAF has been found to be dominated by the decrease in surface albedo that occurs with snow cover loss (SNC), while albedo decreases due to snow metamorphism (TEM) as a result of warming contributes as a secondary effect (Qu and Hall, 2006; Fletcher et al., 2012). Out of the 17 CMIP3 models, only 3 show the TEM component as the dominant contributor to SAF (Lawrence et al., 2012; Fletcher et al., 2012). It is somewhat expected that the TEM term would have a smaller contribution to NET SAF because snow albedo change through metamorphism is fairly minimal (0.80 to 0.50) compared to the change from a snow-covered to snow-free surface (0.80 to 0.10). Qu and Hall (2007) demonstrated that snow metamorphism is a nonlinear function of temperature; increasing from 250 to 270 K, where its maximum is reached. Fletcher et al. (2012) used a test of additivity to see if these components accounted for the bulk of the variance in total SAF (NET).

In order to more realistically represent springtime mean climate, improvements to the land surface scheme in CCSM4 have been extensive (Lawrence et al., 2011). These changes have resulted in a reduction in the bias of albedo for fully snow-covered surfaces from CCSM3 to CCSM4 (Lawrence et al, 2012) and go a long way to more accurately representing SAF during the melt season. SAF is usually calculated for the winter-to-spring transition period because it is at its strongest during this time frame, as this is a time when both snow cover and incoming solar radiation are fairly large (Hall, 2004; Qu and Hall, 2006; Figure 1-9).

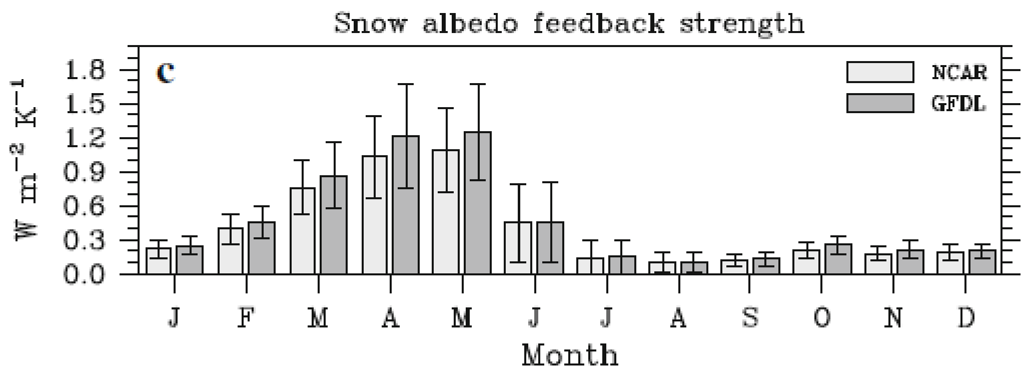

Figure 1-9: Seasonal cycle of the ensemble-mean of feedback strength over NH extratropical land masses in 25 CMIP5 models (Qu and Hall, 2013).

SAF over the boreal forest is expected to be weaker than other biomes, because of its low snow-covered albedo values as a result of vegetation masking the underlying snow surface (Qu and Hall, 2013). The way in which canopy vegetation is represented in climate models has been shown to have a strong impact on the SAF. Models that treat it as a simple surface (no real handling of canopy structure) generated an albedo that was biased high compared to observations and thus a stronger SAF, whereas models with a specific parameterization for various canopy types tend to underestimate the albedo (Qu and Hall, 2007; Kuusinen et al., 2012). This discrepancy in the latter models could be the result of vegetation masking too much ground snow, or the canopy albedo being underestimated (Qu and Hall, 2007).

1.2 Motivation for Research

Modelling of snow processes is of utmost importance because it plays a key role in the climate system and other methods of observation have proven unsatisfactory (Rutter et al., 2009). This includes a lack of both reliable ground-based measurements and accurate remotely sensed observations from sensors like MODIS over northern forest regions (Hall and Riggs, 2007). Understanding of forest snow processes will become even more vital under future climate scenarios, where forest coverage is likely to advance northward into more snow covered areas (Denman et al., 2007).

A large spread in SAF strength among the CMIP5 models explains much of the variation in projections of warming over the Northern Hemisphere (Qu and Hall, 2013; Essery, 2013). It has been suggested that winter observations over the boreal forest, should be used to constrain modeled albedo because of the vastly different parameterizations for the masking effect of vegetation on snow cover among CMIP5 models (Qu and Hall, 2013). Because of this uncertainty in SAF strength between observations and simulations, further comparison and diagnosis at multiple scales using multiple observational datasets and sets of model simulations are therefore needed. We include the use of a second satellite-based product (MODIS), as suggested by Fletcher et al. (2012) in order to reduce the observational uncertainty in SAF. A number of prior studies have looked into the relationship between observed and simulated albedo of snow (Essery et al., 2013; Loranty et al., 2014), but none identified the link between albedo biases over the boreal forest and hemispheric-scale climate feedbacks. Also, in much of the literature there is no discussion of why such albedo biases may exist in the model. The biases between observed and simulated albedo likely have a direct impact on total SAF strength. Prior research has shown signs that simulated SAF in CCSM4 may be weaker than observed over the boreal forest region, but no reason was given for this discrepancy (Fletcher et al., 2012; Figure 1-10).

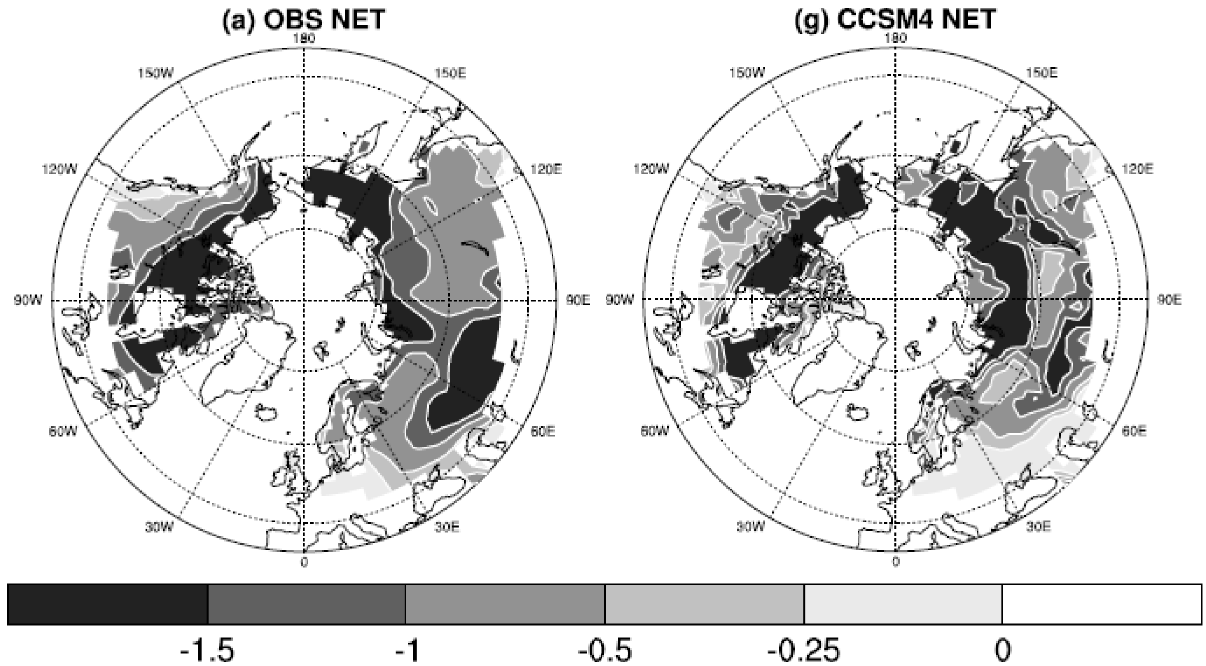

Figure 1-10: Northern Hemisphere MAMJ mean NET feedback (units % K^{-1}) for OBS and CCSM4. Unshaded areas south of ~45°N are excluded from the analysis because S < 0.1 in all months (Fletcher et al., 2012).

1.3 Research Objectives

The overall objectives of this research are to investigate the comparatively weak simulated boreal SAF found in Fletcher et al. (2012), while making use of multiple observational datasets and diagnosing the underlying cause of differences between them. We hope to decompose the previously used methodology of SAF strength over northern hemisphere land (Fernandes et al., 2009; Fletcher et al., 2012) to just look at the feedback strength over evergreen forests. In doing so we aim to answer the following questions:

- How do the model parameterizations compute snow properties and in turn albedo?

- Why do the models underestimate SAF over the boreal region?
- What are the primary mechanisms driving seasonal albedo change over the boreal forest canopy?
- How does the seasonal evolution of simulated and observed albedo and snow vary?
- What is the relationship between vegetation densities in these forested regions and the SAF bias?

This research looks at how a previously observed discrepancy over forested regions may influence climate at a much larger scale. In the calculation of SAF, we investigate the impact of changing only the albedo on its total strength. Both fully coupled and uncoupled climate model simulations are assessed to determine if an interactive climate would affect SAF, and the parameters relevant to it.

1.4 Structure of Thesis

The structure of this thesis is broken into three chapters, the first of which describes necessary background information and provides motivation for this research. Background content includes discussion of prior research into forest snow processes (canopy interception and sublimation), the modelling of snow, and snow albedo feedback. Chapter 2 encompasses the body of this thesis, which is a manuscript. The manuscript (The influence of canopy snow parameterizations on snow albedo feedback in boreal forest regions) provides a thorough evaluation of model simulations from CCSM4 and how its parameterizations impact snow and albedo processes throughout the boreal forest. Chapter 3 summarizes the key findings of this study, provides some limitations to our approach and a number of recommendations for future work.

Chapter 2

The influence of canopy snow parameterizations on snow albedo feedback in boreal forest regions

2.1 Overview

Variation in snow albedo feedback (SAF) among CMIP5 climate models has been shown to explain much of the variation in projected 21st Century warming over Northern Hemisphere land. Prior studies using observations and models have demonstrated both considerable spread in the albedo, and a weak bias in the simulated strength of SAF, over snow-covered boreal forests. Boreal evergreen needleleaf forests are capable of intercepting snowfall throughout the winter, which has a significant impact on seasonal albedo. Two satellite data products and tower-based observations of albedo are compared with simulations from multiple configurations of the Community Climate System Model (CCSM4) to investigate the causes of weak simulated SAF over the boreal forest. The largest bias occurs in April-May, when simulated SAF is one-half the strength of SAF in observations. This is traced to two canopy snow parameterizations in the land model. First, there is no mechanism for the dynamic removal of snow from the canopy when temperatures are below freezing, which results in albedo values in midwinter that are biased high. Second, when temperatures do rise above freezing, all snow on the canopy is melted instantaneously, which results in an unrealistically early transition from a snow-covered to a snow-free canopy. These processes combine to produce large differences between simulated and observed monthly albedo, and are the source of the weak bias in SAF. This analysis highlights the importance of canopy snow parameterizations for simulating the hemispheric scale climate response to surface albedo perturbations.

2.2 Introduction

Seasonal snow cover plays a very important role in the surface energy budget and water balance for much of the Northern Hemisphere, and it has a strong influence on land surface albedo, a key parameter in climate models. Annually, the maximum snow cover extent is approximately 47 million km^2 over the Northern Hemisphere, or approximately 40% of the total land area (Hall, 1988; Robinson and Frei, 2000; Qu and Hall, 2005). A large portion of this snow-covered region is composed of boreal evergreen needleleaf forests, which make up approximately 19% of the snow-covered area (Pomeroy et al., 1998b; Essery et al., 2003; Rutter et al., 2009). These coniferous species can efficiently intercept falling snow, capturing up to 60% of annual snowfall over boreal forests (Pomeroy and Schmidt, 1993; Storck et al., 2002). For much of the boreal forest, snow is on the ground for more than six months annually, so variability in snow albedo during the accumulation and melt seasons has important implications on the energy budget by altering the amount of incident energy absorbed at the surface. There is also a strong linkage between land cover type and maximum surface albedo (Betts and Ball, 1997; Jin et al., 2002), which in turn impacts snow albedo feedback strength.

Snow albedo feedback (SAF) is a positive feedback climate mechanism that has a strong impact on directing shortwave forcing in a changing climate and is the second largest contributor to Arctic amplification (Pithan and Mauritsen; 2014). Qu and Hall (2007) divided SAF into two components; the first related to the albedo contrast between snow-covered and snow-free land surfaces (SNC), and the second as the temperature dependency of snow albedo for a constant snow cover extent and albedo contrast between snow-covered and snow-free land (TEM). It is expected that total feedback strength (NET) will equal the sum of these two terms in an idealized setting, where additivity should be satisfied if there are no other processes contributing to NET (Fletcher et al., 2012). A large spread in SAF strength among the Coupled Model Intercomparison Project phase 5 (CMIP5) models explains much of the variation in predictions of warming over the Northern Hemisphere (Qu and Hall, 2013). Qu and Hall (2013) suggested that winter observations over heavily vegetated surfaces such as the boreal forest, should be used to constrain modeled albedo because of the vastly different parameterizations employed in the suite of CMIP5 models for the masking of vegetation by snow cover. Previous work has shown that modeled SAF strength is significantly weaker than observed over the boreal forest during the spring melt period (Fletcher et al., 2012). This is slightly counterintuitive when considering that other studies have found climate models to overestimate snow albedo over the boreal forest, implying a greater snow-covered to snow-free albedo change in models (Essery et al., 2013; Loranty et al., 2014). Implying that there are other unknown factors at work influencing the simulated spring surface albedo.

The partitioning between SAF components (i.e., into SNC and TEM terms) has highlighted inconsistencies between observational studies, even when analyzing the same data. Fernandes et al. (2009) used Extended AVHRR Polar Pathfinder; APP-x satellite observations of albedo (Wang and Key, 2005) to calculate the SNC and NET terms, with the TEM component as the residual TEM = NET – SNC. Their results showed that TEM and

SNC contributed approximately equally to NET; however, more recently Fletcher et al. (2012) used the same data to show NET was comprised of 69% SNC and only 31% TEM. The partitioning between the two terms in climate models has remained largely unchanged between CMIP3 and CMIP5 (~63% SNC, 37% TEM) (Fletcher et al., 2012; Qu and Hall, 2013).

A lack of ground-based measurements and reliable satellite retrievals over northern forests (Hall and Riggs, 2007) causes uncertainty in our understanding of canopy snow interactions and their effect on surface albedo. Sellers et al. (1995) suggested that an accuracy of measured albedo in the range 0.02-0.05 (dimensionless) is required for albedo characterization at the necessary precision for climate research. Thus, climate models require a consistently accurate global dataset of albedo in order to allow for proper analysis into the sensitivity of climate to certain forcing (Lawrence and Chase, 2007; Liu et al., 2009).

Because of the uncertainties in SAF strength between observations and simulations, further comparison and diagnosis at multiple scales using multiple observational datasets and multiple sets of model simulations are therefore needed. The primary goal of this work is to investigate the causes of the weak bias in simulated SAF over boreal forests, using multiple satellite datasets along with tower-based observations. We also aim to diagnose the underlying cause of differences in observed versus simulated SAF using the Community Land Model Version 4.0 in offline mode and in coupled simulations. The model simulations and observational products are introduced in Section 2. Also included is pertinent information on how SAF is calculated, and how the boreal study area was derived. In Section 3, we present an evaluation of SAF strength from the multiple data sources during the winter to spring transition, and diagnose the causes of SAF biases in model simulations. Section 4 highlights the key findings of this work, and discusses potential improvements to model parameterizations of canopy snow processes.

2.3 Data and Methods

2.3.1 CCSM4 model simulations

We calculated SAF using albedo, temperature and snow cover fraction over land areas polewards of 30°N for the period of 2000-2004 from a set of historical (1850-2005) simulations from the Community Climate System Model, version 4.0 (CCSM4) produced for CMIP5. These simulations are described in detail in Gent et al. (2011), and are the same set used for SAF analysis by Fletcher et al. (2012) and Lawrence et al. (2012). These simulations were run with coupled dynamic ocean, land and atmosphere components, driven by observed historical boundary forcing. The land component is the Community Land Model Version 4.0 (CLM4) (Oleson et al., 2010), which utilizes the Snow, Ice, and Aerosol Radiative (SNICAR) model to compute snow albedo and absorption for snow-covered areas (Flanner and Zender, 2005; Flanner and Zender, 2006; Lawrence et al., 2012).

CLM4 calculates surface albedo in forested regions by first computing radiative interactions within vegetation stands using a two-stream approximation (Bonan, 1996). In the boreal forest ecozone there are areas where a portion of the ground surface is visible from above, this gap fraction is calculated using prescribed leaf area index, stem area index, and canopy height for each plant functional type (PFT) that is derived from MODIS data (Lawrence and Chase, 2007). Albedo is calculated as the weighted combination of surface and canopy albedo, and is dependent on the gap and snow cover fractions (Oleson et al., 2010). The model uses a coefficient of interception to determine how much snowfall will be intercepted by canopy vegetation. This coefficient has a maximum value of 0.25, decreasing with lower leaf area index (LAI) and stem area index (SAI). In order to determine if snow is still present on the canopy after a snowfall event, the model evaluates whether the vegetation temperature is less than 0°C. The optical properties of intercepted snow are set to reflect 50% of the diffuse and direct beam incoming radiation (Oleson et al., 2010). Vegetation free of snow on the other hand has an albedo computed based on its leaf and stem reflectance (Lawrence et al., 2012). Snow on the ground in CLM4 is represented with up to five layers (that include liquid water and solid ice) depending on the total depth (Jin and Miller, 2011). Snow albedo from SNICAR depends on a number of variables, including solar zenith angle, snow depth, albedo of the surface beneath a snowpack, deposition of aerosols, and ice effective grain size (Oleson et al., 2010). Snow effective grain size is simulated using a snow aging parameterization that is a combination of changes that result from dry snow metamorphism, refreezing of liquid water, and accumulation of freshly-fallen snow (Flanner and Zender, 2006). Snow aging occurs fastest when snow is warm, with a large internal temperature gradient, and low density, while cold snow limits the aging process dramatically (Oleson et al., 2010). CLM4 uses a snow cover fraction parameterization that was developed by Niu and Yang (2007), in which SCF rises with increasing snow depth until it reaches complete coverage at depths between 0.2 and 0.3 m depending on ground surface roughness, and snow density or time of year.

An offline simulation from CLM4 forced by the observational dataset from Qian et al. (2006), was available for the period 1982-2004; we selected the years 2000-2004 for comparison with observational products. This type of simulation is useful for evaluating a particular process or parameterization for verification with observations, however, there are limitations to this approach due to the one-way coupling from the atmosphere to the land surface (Dutra et al., 2010).

2.3.2 Observation-based products

2.3.2.1 Satellite data products

We use satellite retrievals from the Moderate-resolution Imaging Spectroradiometer (MODIS) for both surface albedo (α_{sfc}) and snow cover fraction (S_f). The white-sky albedo of the MCD43C3 product, which contains 16 days of observations within each time step at a 0.05 degree resolution, is used to evaluate the modeled surface albedo for the period September 1, 2000 – December 31, 2004. The 16-day sampling period attempts to maximize the number of cloud-free images, while maintaining sub-monthly temporal resolution; however, data gaps associated with persistent winter cloud cover remain a limitation (Schaaf et al., 2002; Barlage et al., 2005). The number of clear looks at the surface essentially determines albedo quality. However, when there are fewer than seven days of cloud-free observations, albedo estimates are acquired from a backup model that uses the available observations to adjust prior knowledge of the surface (Strugnell and Lucht, 2001; Jin et al., 2002). The amount of data infilled by the model varies with the number of cloud-free scenes over the 16 day period, from 25% or less (Albedo Quality flag 2) to 50% or more (Albedo Quality flag 4) (Schaaf et al., 2002).

To mitigate against sampling biases caused by data quality, only MODIS albedo quality grade 2 data or better were included, which yielded sufficient data of adequate quality during mid-winter over the boreal region. Uncertainty in the albedo retrievals also increases when the solar zenith angle (SZA) exceeds 70 degrees, as they typically do for much of the northern extratropics during midwinter (Schaaf et al., 2002; Wang and Zender, 2010). At high solar zenith angles, especially where canopy vegetation is dense, shadows become more prominent and can impact the ability to accurately detect the presence of snow (Hall et al., 1995). This vegetation shading also affects albedo through an apparent darkening of the surface (Jin et al., 2002). Although we do include the MODIS data for December, January and February, when SZAs are large (>75), there is greater uncertainty during this time period. Our analysis is focused on the March through June time period, so uncertainty in the albedo retrievals due to SZA is not expected to affect our results.

To provide an independent observational estimate of SAF for comparison with MODIS and the climate models, we use surface albedo retrievals from Advanced Very High Resolution Radiometer (AVHRR) Polar Pathfinder extended (APP-x) project (Wang and Key, 2005). In this study we use a version of the APP-x albedo product that was extended to 2004 at a spatial resolution of 25 km.

Snow cover fraction was taken from the MODIS MOD10C1 daily product, on the same 0.05 degree grid as the albedo data (Hall et al., 1995; Hall et al., 2002). This product is produced by aggregating the native 500 m resolution MODIS product to a grid more suitable for comparison with climate model output (Schaaf et al., 2002).

Snow water equivalent (SWE) is taken from the GlobSnow data record (Takala et al., 2011) to provide an observational estimate of snow depth and a secondary estimate of snow cover fraction to compare with MODIS MOD10C1. SWE estimates are converted to both

SCF and snow depth through two processing steps in order to allow for comparison with output from CLM4 and CCSM4. Snow depth is calculated from SWE using an assumed constant density of 0.24 g/cm^3, which exactly matches the fixed density values in the GlobSnow algorithm, and represents a reasonable average value across the boreal forest (Sturm et al., 2010; Takala et al., 2011). Snow cover fraction (SCF) is calculated using a threshold of 60 kg/m^2 of SWE, where a value greater than this is assigned a fully snow-covered designation (1) and less than the threshold is divided by the full snow cover value to give a fractional return (2).

$$\text{SWE} \geq 60\ \text{kg/m}^2\text{: SCF} = 1 \tag{1}$$

$$\text{SWE} < 60\text{kg/m}^2\text{: SCF} = \text{SWE} / 60\ \text{kg/m}^2 \tag{2}$$

Bilinear interpolation is used to remap all satellite-derived data products to the spatial grid of the CCSM4 model (0.94° latitude by 1.25° longitude).

2.3.2.2 Meteorological tower data

Tower measurements of albedo are taken from the Boreal Ecosystem Research and Monitoring Sites (BERMS), Old Jack Pine location in central Saskatchewan. These data are employed to evaluate how snow and albedo processes interact at the point scale in a dense canopy environment. The Old Jack Pine (OJP) site (53.91634 N, -104.69203 W) is a mature forest stand in central Saskatchewan that is representative of the landscape of the southern boreal forest (Amiro et al., 2005; Neumann et al., 2006). The site has a mean canopy height of approximately 13 m, and is considered an open canopy as snow can fall unobstructed through gaps in tree cover (Neumann et al, 2006). Snow depth measurements are taken from a small clearing between trees, while the tower measurements of albedo are taken above the canopy at 28 m. Surface albedo is calculated at local solar noon as the ratio of upwelling to downwelling shortwave radiation.

To diagnose the underlying causes of disagreement between satellite estimates, simulated albedo and SAF, the tower data are used to evaluate biases in the model output. This approach provides a comparison between the climate models and observations at daily frequency, which is not possible with the optical satellite retrievals, but is required to fully resolve the complex canopy interception, storage, and melt processes affecting the subseasonal evolution of boreal albedo. The comparison between the model output and point observations is appropriate in this context because the boreal forest land cover is sufficiently homogeneous that the OJP tower site is spatially representative of the region as a whole, and the best match with the evergreen needleleaf PFT (Roman et al., 2009).

2.3.3 SAF Calculation

We use a similar approach to previous work by Fletcher et al. (2012), with SAF characterized using the month-to-month change of climatological (2000-2004; despite the short period, the results are comparable with the 1982-1999 period used by Fletcher et al. (2012) monthly mean surface albedo, snow cover fraction, and temperature during the winter to spring transition. Three monthly transitions are examined: March to April (Mar-Apr), April to May (Apr-May), and May to June (May-Jun; months will henceforth be abbreviated by their first three letters). These were also averaged together to represent the seasonal mean snow albedo feedback (MAMJ). We use the seasonal mean because peak SAF is determined by the shifting snow pack and is an important driver of regional climate changes (Fletcher et al., 2012).

We assume that the surface albedo (α_{sfc}) in a grid cell can be represented by a linear combination of the reflectance from snow-covered and snow-free surfaces:

$$\alpha_{sfc} = \alpha_{snow} * S_f + \alpha_{land} * (1 - S_f) \tag{3}$$

where α_{snow} represents the albedo of snow, S_f represents the snow cover fraction, and α_{land} is the snow-free albedo, as represented by the average α_{sfc} for the two months following complete snow melt (Fernandes et al., 2009; Qu and Hall, 2007).

Finite differences are utilized to calculate estimates of the idealized SAF components for each pixel:

$$NET = \Delta\alpha_{sfc} / \langle \Delta T \rangle \tag{4}$$

where deltas represent the difference over a monthly transition, and the angle brackets around the change in surface temperature show that it has been averaged over the study region north of 30°N. Only land areas that have a snow cover fraction greater than 0.1 in the first month of each winter-spring transition are included in the analysis, while those with a lesser fraction have a SAF set to zero. These resulting SAF component strengths are then averaged over the monthly transition phases in the melt period to create MAMJ means. Hemispheric mean SAF values and maps showing the geographic distribution of this climate feedback are produced in an attempt to show where and how greatly the observations and models differ.

It is important to note that the NET SAF calculations for CLM4, APP-x and MODIS use snow cover and temperature data from the land model, so that α_{sfc}, α_{snow} and α_{land} are the only variables changing in equations 1 through 4. This allows the impact of albedo variability on calculated SAF strength to be isolated.

2.3.4 Determining the boreal study area

The boreal forest is primarily composed of needleleaf evergreen trees (Bonan et al. 2002), and in this study we define the "boreal region" as all CLM grid cells with at least 75 percent of the boreal needleleaf evergreen plant functional type (PFT). This threshold value was chosen to obtain a representative sample of grid cells, while also limiting the influence from non-boreal PFTs on the albedo and snow calculations. The remaining land cover in boreal grid cells is typically comprised of small percentages (5-20%) of shrub and grasslands. The boreal region is a unique ecozone because the year-round canopy presence acts to partially mask surface snow cover (Bonan, 2008). This masking of surface snow makes the boreal forest easily identifiable during mid-winter from satellite observations because its peak albedo (~0.3-0.35) is much lower than that of freshly snow-covered grassland to the south and tundra to the north (~0.80-0.85) (Jin et al., 2002; Barlage et al., 2005; Kuusinen et al., 2013).

2.4 Results and Discussion

2.4.1 SAF strength during the winter to spring transition

We first evaluate how well the CLM4 model represents SAF over boreal forest areas by comparing SAF calculated using satellite retrieved albedo from MODIS and APP-x with simulated albedo from CCSM4 and CLM4. Fig. 2-1 shows that melt period (MAMJ) SAF in CCSM4 exhibits a low bias. The simulated feedback strengths for CCSM4 and CLM4 are -0.60 $\%K^{-1}$ and -0.67 $\%K^{-1}$ respectively, about 30-40% weaker than that calculated from the satellite data (-1.06 $\%K^{-1}$ for APP-x, -0.87 $\%K^{-1}$ for MODIS).

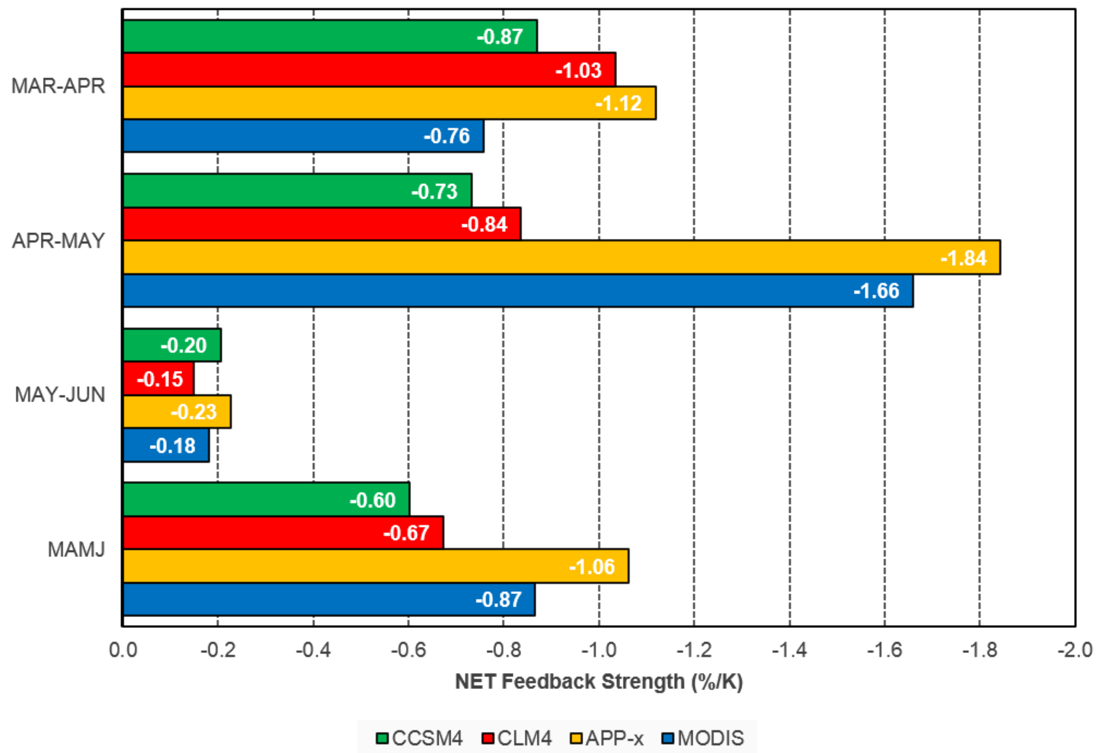

Figure 2-1: Total (NET) snow albedo feedback strength (%/K) over the boreal forest (>75%) for CCSM4 (green), CLM4 (red), APP-x (yellow), and MODIS (blue).

We can trace the majority of this seasonal difference to a single pair of months, Apr-May, during which SAF in MODIS and APP-x is approximately twice the strength of the simulated SAF. The low bias is present in both CCSM4 and CLM4, suggesting that it is not a direct result of biases in model simulated snow cover or temperature.

CCSM4 underestimates NET SAF over a majority of the hemisphere, with the bias between observed and modeled SAF largest and of uniform sign over regions with high fractions of boreal forest cover. This bias is most pronounced over the boreal region during Apr-May (Fig. 2-2), while it peaks earlier at locations further south, and later for locations to the north because that is when SAF is locally strongest (not shown).

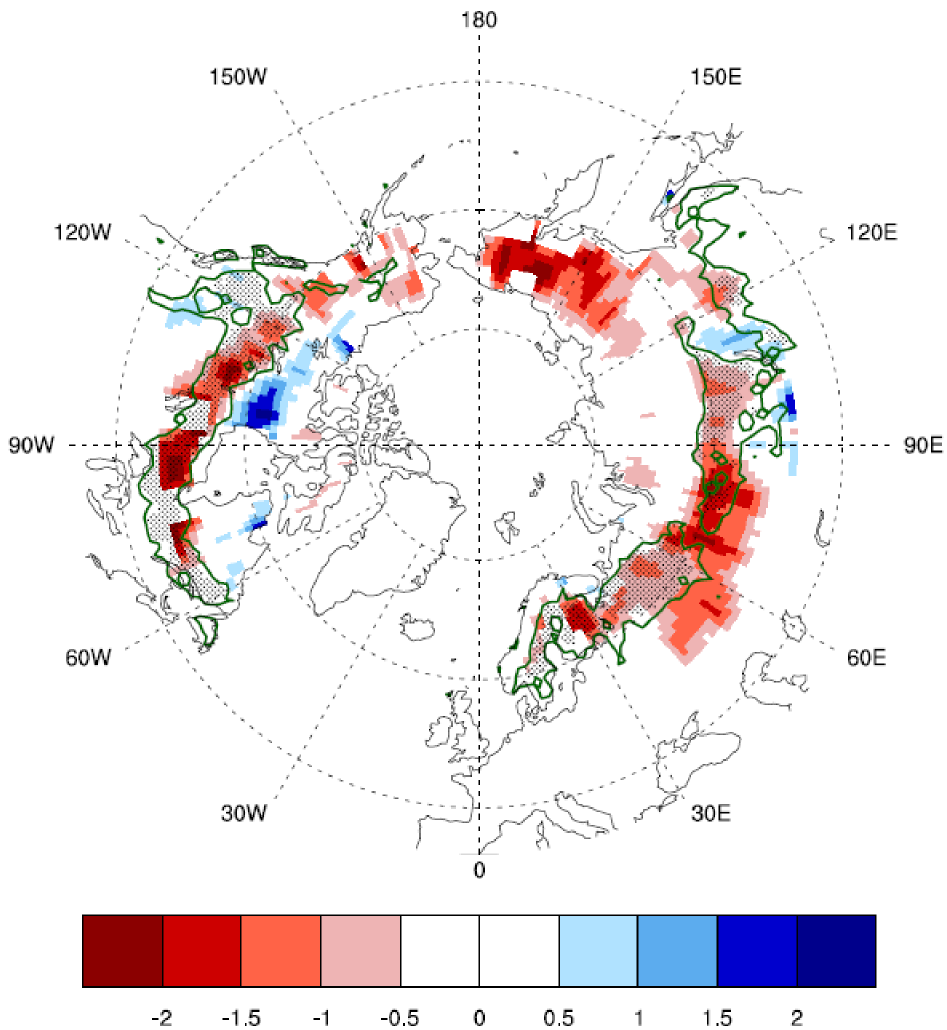

Figure 2-2: Map of NET snow albedo feedback bias (MODIS – CCSM4) for April-May (units % K^{-1}). Mean NET has a negative sign, so areas shown in red (blue) indicate that MODIS is stronger (weaker) than CCSM4. The dark green line shows the 50% boreal region, while black stippling indicates the >75% region.

There are also large biases in a limited number of areas with low fractions of boreal forest cover, for example over NE Siberia (negative SAF bias in simulations) and the Canadian tundra (positive SAF bias in simulations). The SAF biases in these regions do not show a consistent sign, which suggests they are unlikely to be a result of the same process or parameterization. In addition, both regions primarily consist of Arctic grass and bare land PFTs, with little boreal forest cover, and so they fall outside the scope of this study.

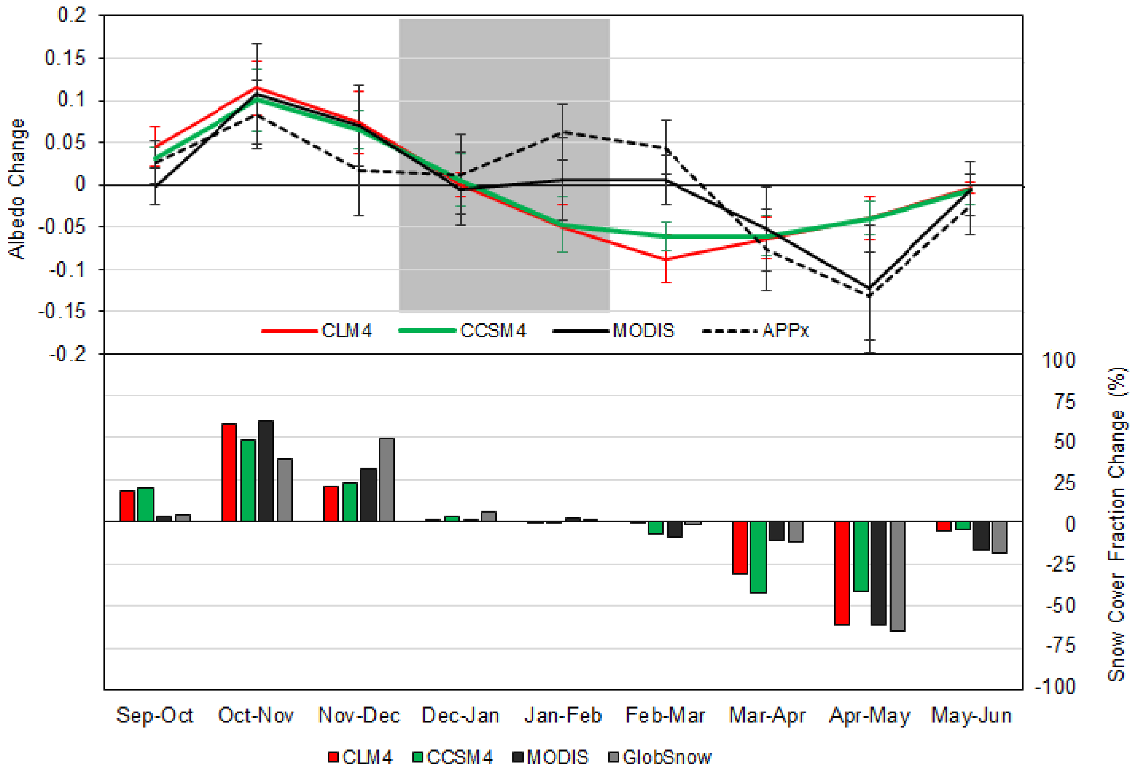

Figure 2-3: Monthly change in (a) albedo and (b) snow cover fraction (SCF) for boreal forest (>75%). Monthly changes are climatologies over the 2000-2004 period for CLM4, MODIS, and APP-x. Snow products used include CLM4, CCSM4, MODIS, and GlobSnow. The grey shaded region indicates months of the year when observational uncertainty is high due to large solar zenith angles (>75 degrees).

In order to further investigate the SAF bias over the boreal forest during the melt period, it is instructive to evaluate the temporal evolution of snow albedo prior to melt. Fig. 2-3a shows that in satellite observations, snow albedo increases weakly, or remains constant from Dec until Mar-Apr, when it decreases sharply before reaching its seasonal minimum in May-Jun. By contrast, the models exhibit relatively large albedo decreases during Jan-Mar, but only weak decreases during Apr-Jun. In Jan-Feb and Feb-Mar, there is a difference in the sign of monthly albedo change for the observations (positive month-to-month albedo change) versus the models (negative month to month albedo change; Fig. 2-3a). Apr-May albedo change in the models is less than 50% of the observed change. Our interpretation of this result is that the simulated albedo during melt can only decrease by a small amount because it already decreased substantially during Jan through Mar.

One plausible explanation for the simulated late winter decrease in albedo is that the models are losing SCF sooner than the observations; however, Fig. 2-3b demonstrates that this is not the case. Simulated SCF remains stable at its maximum during mid-winter, agreeing well with observational datasets from MODIS and GlobSnow. Thus, the albedo in the models is decreasing during Dec to Feb under conditions of 100% snow cover, whereas the observational albedo decrease in Apr and May coincides with SCF loss (Fig. 2-3b). Another possible explanation for the early decrease in albedo could be that simulated snow depth is much lower than observations, allowing for the underlying ground surface to influence albedo through a thin snowpack. In fact, we find the models overestimate snow depth for the boreal region relative to the GlobSnow retrievals (bias of +13 cm pre-melt maximum depth or 33% greater than GlobSnow max; Table 2-1).

	CLM4	GlobSnow	Bias (Model-Obs)
90^{th} Percentile Depth (cm)	48.8	37.6	11.2
Max Depth (cm)	53.0	39.9	13.1

Table 2-1: Mean snow depth biases from CLM4 in relation to GlobSnow over the boreal region. GlobSnow depth is derived from SWE data using a constant density.

2.4.2 Diagnosing the causes of weak SAF in model simulations

For the remainder of the paper we focus on the CLM4 offline simulations (henceforth, CLM4-OFF) as a diagnostic tool to evaluate the biases in the CCSM4 simulations. The primary benefits of this approach are (i) the offline simulations have observation-based forcing, allowing for a more direct comparison with the satellite datasets, and (ii) daily snow cover and albedo were archived for CLM4-OFF, but not for CCSM4. However, we emphasize that the SAF and albedo biases that form our main focus are present in both CLM4-OFF and CCSM4 (Figs. 2-1 and 2-3), which strongly suggests that the primary cause of the biases lies within the land model (CLM4), rather than in the boundary forcing (for CLM4-OFF) or in the non-land model components (for CCSM4).

The daily average snow depth and albedo for the 2000-2004 period were first plotted against each other and color-coded as a function of month (Figs. 2-4a,b). This figure reveals the two key differences that lead to an explanation of why CLM4-OFF albedo evolves differently through the snow cover season relative to the observations. First, the peak albedo in the model (~0.33) is consistently higher than observations (~0.26), stays at its peak value for longer (~ 2 months), and shows much less variability during the winter than the observations. We attribute the difference in variability to comparing point data with grid-box average data (It is found to be a secondary effect, and so will not discuss it further here). The peak albedo tends to occur when snow depth exceeds ~0.2 m in the model, whereas for the tower site peak albedo occurs when snow depth exceeds ~0.1 m. Second, during Feb and Mar there is a strong decline in CLM4-OFF albedo despite snow depths continuously in excess of 0.5 m; this behavior is completely absent in the observations. We will demonstrate below that these characteristics of the simulated seasonal evolution of albedo are directly related to two separate physical parameterizations in the snow module of CLM4.

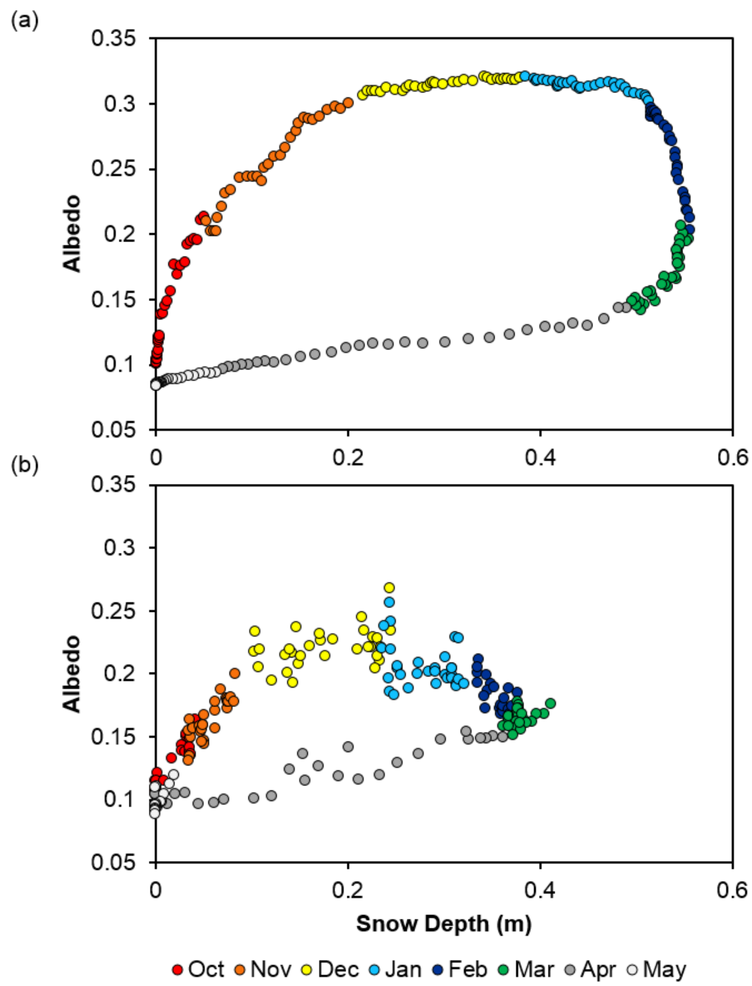

Figure 2-4: Daily climatological snow depth and albedo scatterplots for (a) CLM4 over Pure Boreal Forest (>90%) grid cell and (b) Old Jack Pine

The consistently high bias in CLM4-OFF albedo between Dec and Mar, is very likely related to the absence of a mechanism in the model to offload intercepted snow from the canopy when temperatures remain below freezing and the sun angle is low (Oleson et al., 2010). As a result, CLM4 exhibits a tendency for snow to persist in the canopy for an unrealistically long time during mid-winter. This effect is demonstrated by the very stable albedo values depicted by the yellow (Dec) and blue (Jan) points in Fig. 2-4a.

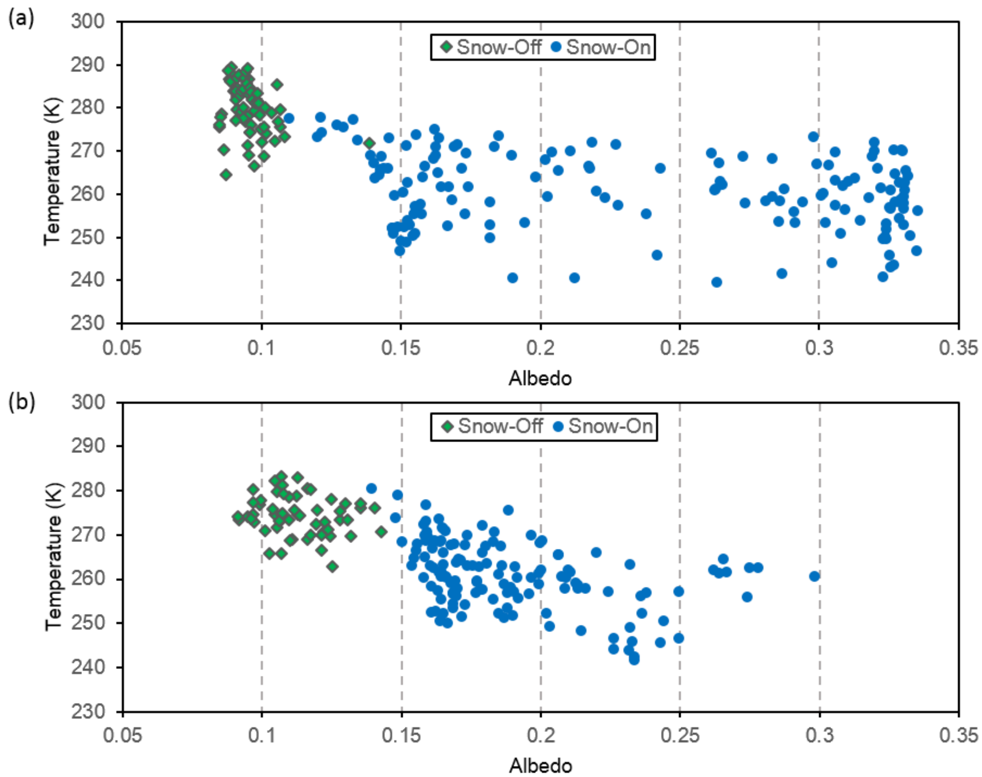

Figure 2-5: Relationship between daily albedo and mean surface temperature for (a) a boreal forest grid cell in CLM4 for Oct 1, 2001 – Apr 30, 2002 (b) Old Jack Pine 01-02. Presence of snow on the ground is denoted by snow-on (blue circles) and snow-off (green diamonds).

Figure 2-5 shows the relationship between albedo and surface temperature through one accumulation and melt season (including snow and non-snow-covered cases). Snow-on and snow-off refer to the presence of snow on the ground as determined by daily snow depth measurements at the ground surface. The cluster of points to the right of Fig. 2-5a show that albedo exceeds 0.3, up to a well-defined maximum value at around 0.33, in more than one quarter of days with non-zero snow cover and mean temperatures below 270 K. By contrast, there are no days in the OJP data where albedo exceeds 0.3 (and only one day where albedo exceeds 0.27) and, more importantly, there is no clustering of points at a maximum value indicative of persistent snow storage in the canopy (Fig. 2-5b). Rather, the clustering in OJP tends to occur at albedo values around 0.17, characteristic of snow lying on the ground but not in the canopy, as is the case for the majority of the winter season in a typical boreal conifer forest setting (Betts and Ball, 1997).

Next, we address the strong decline in simulated albedo in Feb-Mar that occurs in the presence of snow depths exceeding 0.5 m. The cause of this decline is that the canopy snow parameterization in CLM4 immediately removes all snow from the canopy when temperatures exceed 0°C at any time during the day. As early as mid-Feb, higher sun angles can induce brief warming events where temperatures rise slightly above freezing. Even if such events persist for a single model time step CLM4 still melts all snow stored on the canopy, revealing a green (snow-free) canopy with an albedo that is a weighted combination of its leaf (0.07) and stem (0.16) albedo, dependent on the LAI and SAI, respectively. However, because the duration of periods of above-freezing temperatures are short at this time of year, only limited melting can occur in the surface snow pack beneath the canopy (typically having greater than 0.5 m depth), and so the ground beneath the canopy remains snow-covered long after the canopy becomes snow-free.

Canopy snow (referred to as water on the canopy where temperatures are below 0 ^{o}C in CLM4) has a major influence on the simulated albedo of a grid cell. Canopy water suddenly increases during Oct and Nov with the arrival of snow across the boreal forest, until the canopy’s finite snow capacity is reached in Dec and Jan (Fig. 2-6). There is a sharp decrease in boreal canopy snow storage (as a result of warming events) during Feb and Mar that drives the albedo decrease shown in Fig. 2-3a over the boreal forest during that same period.

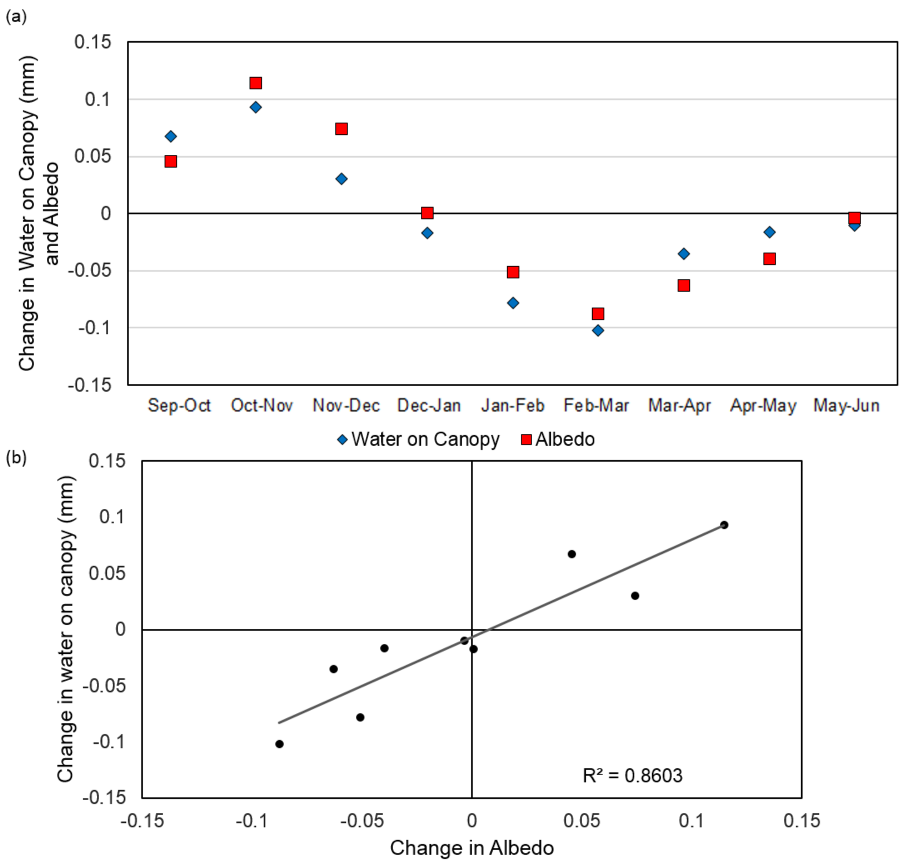

Figure 2-6: Monthly climatological change in water on canopy (blue diamonds) and albedo (red squares) from CLM4 Northern Hemisphere boreal forest (>75%) (b) Same relationship shown as a scatterplot.

Canopy snow gradually decreases back to a point of no monthly change over the remainder of the melt period (similar to surface albedo), until all snow is removed from the canopy. These two variables are strongly correlated over the boreal region, both reaching their greatest negative change in Feb-Mar, with an R-squared value of 0.86 over the snow season (Fig. 2-6b). Therefore it is the canopy albedo that is driving seasonal albedo changes, while surface variations are muted.

Although we do not have a direct observational measure of canopy snow to compare with the simulated canopy interception, prior studies involving boreal forest locations have found intercepted snow load and the efficiency of canopy interception to vary greatly throughout the snow season, with no clear seasonal pattern (Hedstrom and Pomeroy, 1998; Garvelmann et al., 2013) because interception efficiency is dependent on numerous factors including snowfall amount, snow density, temperature, leaf area index, and wind speed (Hedstrom and Pomeroy, 1998; Marsh, 1999).

The importance of canopy snow to the surface albedo and in turn SAF strength in CLM4 cannot be understated. Areas where the SAF bias shown in Fig. 2-2 is greatest tend to have a high exposed leaf area index (LAI), suggesting that it is in denser canopies where the bias is more likely to be large (Fig. 2-7). We are not suggesting there is any fitted relationship between LAI and SAF bias, rather that the grid cells with the largest SAF bias are clustered around the highest LAI values for pixels that are made up of at least 75% boreal evergreen needleleaf forest. This means that in locations where the canopy has a greater impact on albedo, there is likely to be a larger discrepancy between SAF calculated from observations versus simulations, because canopy snow loss during Feb and Mar plays a bigger role. As a knock-on effect of this Feb and Mar canopy melt, the albedo over the boreal region is already well below its peak value by the time the spring melt period begins (Figs. 2-3a, 2-4a), thus drastically weakening the maximum possible change in surface albedo (numerator in Eq. 4) and as a result, the NET feedback strength for MAMJ. This explains why the NET SAF is much weaker in simulations versus the satellite observations, which do not see a major change in α_{sfc} until the spring melt period.

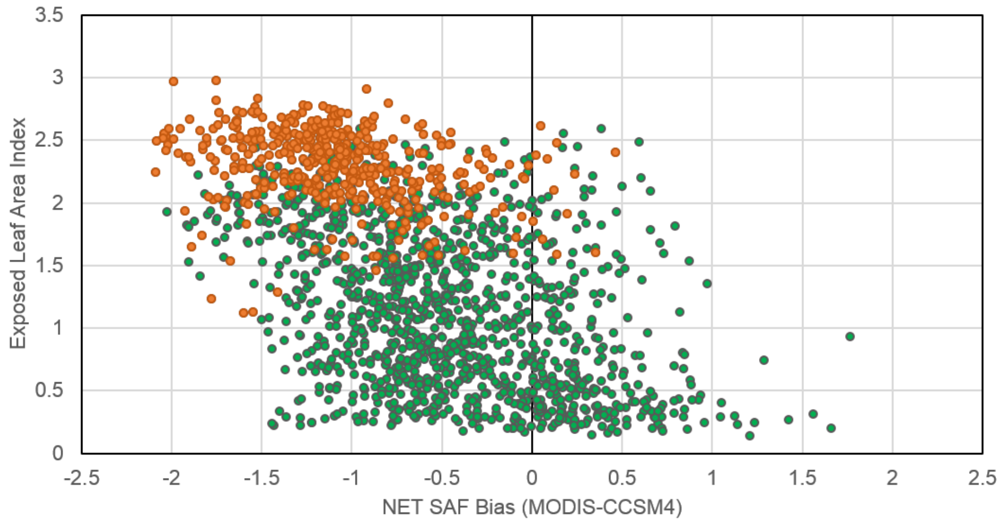

Figure 2-7: Relationship between April-May NET SAF bias (MODIS-CCSM4) and exposed leaf area index over all grid cells with at least 1% boreal evergreen forest (green circles), and grid cells with at least 75% boreal evergreen forest (orange circles).

2.5 Conclusions

SAF in climate model simulations over the boreal forest was shown to be weaker than in multiple observational datasets. This bias in CCSM4 was found to be largest in Apr-May, when simulated SAF is one-half the value from observations. Large differences in the monthly evolution of albedo between models and satellite observations have a significant impact on the total SAF strength. A larger-than-observed decrease in canopy snow during Jan-Mar was the main cause of these albedo differences, occurring as a result of mid-winter warming events that remove all canopy snow instantaneously. This situation, where the canopy is snow-free but the ground is snow-covered, reduces the model's grid cell albedo over boreal regions to ~0.17; i.e., in between fully snow-covered and snow-free albedo values. Much of the simulated decrease in albedo occurs before the melt period (MAMJ), and is therefore not captured in calculations of SAF. Thus leading to a weaker seasonal feedback strength than the observations.

In areas with higher wintertime LAI, where the boreal canopy plays a more important role, the SAF bias has a tendency to be large ($-1.25\%K^{-1}$). This is due to the way canopy snow is parameterized in CLM4: snow cannot be removed from the canopy while the ambient air temperature is below freezing. This creates a 'sticking' effect of snow in the canopy during the coldest months that is not found in field measurements. The measurements instead reveal that sublimation and wind-driven unloading events prevent albedo from persisting at its peak value for an extended period of time (Bartlett et al., 2006). This model overestimation of albedo has also been demonstrated using an idealized model by Essery (2013), but our study is the first to link this behavior to a hemispheric-scale climate feedback. The bias is not strictly limited to boreal forests (Fig. 2-2), but it is the boreal region in particular in which the bias is most pronounced and has consistent sign. These results offer the prospect for further development of a more realistic canopy-snow parameterization based on satellite and in situ measurements, which should reduce model biases in simulations of SAF.

It is clear that there are still some uncertainties in the observational data that limit our understanding of boreal forest snow processes. For example, rapid melt events during spring are often not captured using the 16-day mean albedo product from MODIS (Wang et al., 2014). A finer spatial resolution in observational data would also help improve knowledge of the relationship between snow and albedo, especially within densely or mixed forested landscapes. Future work should also address the positive SAF bias over the Canadian subarctic tundra, and also attempt to investigate whether the biases found here exist within other CMIP5 models. Importantly, reducing the spread in SAF among CMIP5 models could help narrow the spread in projections of future warming over the Northern Hemisphere.

2.6 Acknowledgements

Data used in this study can be found at the following locations - CCSM4 simulations (Gent et al., 2011) are from the CMIP5 archive. The MODIS albedo and snow cover data was obtained from the online Data Pool at the NASA Land Processes Distributed Active Archive Center (LP DAAC; https://lpdaac.usgs.gov/data_access). GlobSnow snow water equivalent data can be found at the GlobSnow archive (http://www.globsnow.info/swe/). APP-x data based on Wang and Key (2005) were obtained from the University of Wisconsin, Madison (ftp://stratus.ssec.wisc.edu/pub/appx). BERMS site measurements were acquired from the Fluxnet Canada Data Information System for which we would like to thank Alan Barr (berms.ccrp.ec.gc.ca). We acknowledge funding from the Natural Sciences and Engineering Research Council of Canada's Climate Change and Atmospheric Research Initiative via the Canadian Sea Ice and Snow Evolution (CanSISE) Network. We also thank Mark Flanner and David Lawrence for helpful suggestions regarding the canopy snow parameterizations.

Chapter 3

Conclusions

3.1 Summary

Variation in snow albedo feedback (SAF) among CMIP5 models has been shown to explain much of the variation in projected 21st Century warming over Northern Hemisphere land (Qu and Hall, 2013). Prior research has found a considerable spread in the observed and modeled snow albedo over forested regions. In Chapter 2, the parameterization of canopy snow processes has been demonstrated to have a drastic influence on snow albedo feedback over the boreal forest. SAF in model simulations was shown to be much weaker than observations, with the largest bias in Apr-May. Large differences in the seasonal evolution of albedo between models and satellite observations have been shown to be influencing the total feedback strength. This bias is due to two canopy snow parameterizations within CLM4. First, there is no mechanism for the dynamic removal of snow from the canopy when temperatures are below freezing, which causes albedo values in midwinter to exceed observations. Second, when temperatures do rise above freezing, even for one time step, all snow on the canopy is melted instantaneously. This results in an unrealistically early transition from a snow-covered to a snow-free canopy. In this situation, where the canopy is snow-free but the ground is snow-covered, the model's grid cell albedo over boreal regions is reduced to approximately 0.17. However, since much of the simulated decrease in albedo occurs before the melt period (MAMJ), it is not captured in our calculations of SAF. Thus leading to a much weaker seasonal feedback strength than the satellite observations.

In Section 1.3 there were a few research questions that throughout the course of this study have been answered. First, finding out the inner workings of CCSM4 (as laid out in section 2.3.1) was crucial to understanding how the model simulates albedo and snow properties. In answering the question of why the model's simulated SAF is biased low, the seasonal evolution of albedo and the differences between datasets needed to be evaluated. It was determined that the albedo discrepancies were not being influenced by either snow cover fraction or snow depth as both showed fairly good agreement with observations. The mid-winter decrease in simulated albedo was in fact being driven by the simplistic canopy snow parameterization and its vulnerability to winter warming events. It is in densely forested regions (ELAI > 2; Figure 2-7) where the SAF bias has a tendency to be large, thus showing where the model is least accurate at representing springtime climate.

3.2 Limitations

There remain some issues with the representation of these physical processes within models that limit modelling performance. Notably the issues discussed in Section 1.1.2, such as the representation of snowpack variability at different layers, or the way in which snow albedo is parameterized. Certain aspects of the snow regime need to be better understood before modelling improvements can be made, such as the impact of blowing snow on both snow buildup and sublimation (Essery et al., 2003). There are other limitations as a result of model output, whereby all observational data needed to be averaged up to the coarse climate model grid for comparisons to take place. The output from the model also only extended to the end of 2004, which severely limited the length of our study period as MODIS did not become available until February of 2000.

Observations are essential for the improvement of model simulations, but there is a measure of uncertainty with them. Long term records of snow for instance are hard to come by because of difficulties in making consistently accurate measurements (Vaughan et al., 2013). Monitoring of snowfall by weather stations is often sporadic or affected by measurement techniques changing over the observational period (Kunkel et al., 2007). Some sources of inhomogeneities in the data include measurement location changes, adherence of observers to instructions, the more recent implementation of snowboards, and the time of observations (immediately after a snowfall event or at a set interval; Kunkel et al., 2007). On the other hand, satellite-based records of snow depth and SWE tend to have accuracy issues in mountainous and densely forested regions (Vaughan et al., 2013; Hall et al., 2001; Takala et al., 2011), with a consistent low bias for estimations of SWE in dense forest regions (Takala et al., 2011).

The effectiveness of visible sensors to observe the surface is often limited by cloud cover (Foster et al., 2005). This is particularly troublesome during the boreal winter when cloud cover is often persistent, and can hinder the detection of snow on the ground (Hall et al., 2010). Also, because of the northern location of our study area there were some issues introduced as a result of large solar zenith angles during winter (which can impact albedo; Schaaf et al., 2002). Uncertainty in these satellite estimates of surface albedo also stem largely from the use of a 16-day mean value, in the case of MODIS. This duration can be too long to effectively capture rapid change events such as the spring melt (Wang et al., 2014). All of these issues can limit our confidence in the observations used to evaluate the model over long time periods. However, records of snow cover extent (SCE), an important indicator of climate change, have been extended back over 90 years by blending satellite and in situ measurements (Brown and Robinson, 2011). Assimilation approaches similar to this one are likely the most viable option to compiling long term observational datasets for model comparison while avoiding limitations from a single set of observations.

3.3 Future work

There are many interesting questions that arise following the findings of this work, providing great potential for future research. First there is a wide variety of novel experiments that could be run to investigate canopy/snow processes now that its importance to snow albedo feedback has been established. The primary avenue of research is to evaluate the seasonal albedo evolution and snow albedo feedback for all CMIP5 models over the boreal forest region. This would allow us to see where CCSM4 fits within the hierarchy of models at representing forest processes. The CMIP5 models could be characterized by their surface/canopy albedo schemes to see if the type of parameterization has a drastic impact on seasonal albedo evolution.

Another experiment that should follow this work is to eliminate the albedo bias between models and observations by prescribing observational albedo forcing over the boreal forest region. This approach would use either daily satellite-based observations or a weighted average of satellite and point observations through an assimilation technique to see what impact the current albedo biases (CLM4 peak albedo too high, but also decreases too early) have on climatic variables. Prior studies (including our own) have shown that CLM4 overestimates boreal peak snow albedo compared to observations (Essery, 2013; Loranty et al., 2014). Due to the expansive size of the boreal region, this albedo bias could have a significant impact on climate during winter-spring.

Following along the theme of improving model representations of snow and albedo over boreal forests, there is room for the creation of a revised canopy albedo parameterization. Since the current model is lacking an explicit snow removal mechanism, certain biases develop over the winter-spring transition period. Therefore a new parameterization should remove all snow after a certain number of days, with canopy storage affected by wind (increased sublimation in windy conditions). There should be a non-linear melt pattern that depends on how close the temperature is to the freezing point and how long the melt duration lasts. These experiments and others along these lines will go a long way to helping improve the current representation of boreal winter processes through diagnosing areas of uncertainty.

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