Sensitivity of snow processes to warming in the Canadian Rockies

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ABSTRACT

The Canadian Rockies region has experienced substantial climate warming, snowpack decline and glacier retreat, and is anticipated to undergo further warming due to anthropogenic climate change. To better understand the sensitivity of snow processes to warming, a spatially detailed physically based snow hydrology model was constructed for Marmot Creek Research Basin, Alberta, Canada using the Cold Regions Hydrological Modelling platform and used to assess the snow hydrology of a high elevation alpine environment and two medium elevation environments, one densely forested and the other a small forest clearing. Processes modelled include precipitation phase, snow redistribution by wind, snow interception and canopy unloading, sublimation from blowing, intercepted and surface snow, and energy budget snowmelt. The model was run with current and then with perturbed climates with increased temperature, holding relative humidity constant. Increasing temperatures increase the rainfall fraction of precipitation which reduces snow water equivalent (SWE), but also limits blowing snow erodibility. Warming also enhances unloading of snow from conifers which reduces sublimation loss and so increases peak SWE. Increasing temperatures accelerate the initiation of snow ablation by increasing incoming longwave, turbulent and advected heat fluxes, and reducing albedo, but advance ablation into periods of lower insolation that sustain slower snowmelt. Whilst snowfall in the alpine is proportionately reduced the most, peak SWE is most reduced in the forest and then the clearing. Blowing snow ablation processes (transport and sublimation) common in the alpine are reduced more than intercepted snow ablation (sublimation) processes found in the forest. Melt rate reductions are greatest under forest canopies followed by clearings, with much smaller reductions found for the alpine environment. The advance of the snow-free date was greatest in the forest, followed by the alpine and then the clearing environments. Impacts do not proceed linearly with rising temperatures; 2 °C of warming leads to a shift from snowfall to rainfall dominance, a substantial decline in snowpacks and shortening of snow seasons at all elevations. However, 5 °C of warming leads to ephemeral low elevation forest snowpacks and an order of magnitude reduction in high elevations snowpacks, with snow-free dates advanced by from four to six weeks.

Keywords: snow hydrology, snow redistribution, snowmelt, climate change, Canadian Rockies

INTRODUCTION

Climate change is expected to alter Canadian Rockies hydrology substantially because of the importance of snowmelt to streamflow generation (Pomeroy et al., 2012; Fang et al., 2013) and the impact of warming on temperature-sensitive snow processes (Woo and Pomeroy, 2012; Zhang et al., 2000). The average warming in western Canada has been 0.5°C to 1.5°C, over 1900 to 1998, with the greatest increases in winter daily minimum temperatures (Zhang et al., 2000). This

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warming has been exceeded in the Front Ranges of the Canadian Rockies at all elevations with winter minimum temperatures at mid-elevations increasing by 3.6 °C and mean temperatures increasing by 2.6 °C since the early 1960s (Harder et al., 2015). With rising air temperatures, the ratio of rainfall to snowfall is increasing in the region (Davis et al., 1999; Shook and Pomeroy, 2012; Zhang et al., 2000). Increases in the rainfall ratio (rainfall fraction of total precipitation) along with more frequent mid-winter snowmelt events have led to a decline in the seasonal snowpack of western North America (Mote et al., 2005) and this is manifest in the Front Ranges of the Canadian Rockies where peak snowpacks at low elevation have dropped to half of their 1960s values (Harder et al., 2015). Snowcovered area in the northern hemisphere has shrunk 5.4% over the period from 1972 to 2006 (Déry and Brown, 2007), especially in March (7%) and in April (11%) when loss of snow cover is associated mainly with spring warming (Brown and Robinson, 2011). Trends in precipitation volume, in contrast to precipitation phase, are contradictory in the Canadian Rockies as some studies show increases (Zhang et al., 2000) and others do not (Valeo et al., 2007; Harder et al., 2015). A recent study of Canadian Rockies Front Ranges precipitation showed no trends but greater clustering into spring and early summer as multiple day events at higher elevations (Harder et al., 2015). This clustering may explain why there is no temporal trend in high elevation peak snowpacks in the Front Ranges of the Canadian Rockies (Harder et al., 2015).

Nogués-Bravo et al. (2007) compiled projections of greenhouse gas induced climate warming in high mountains for the end of the 21st century and found that these regions are expected to warm substantially by up to 5.3°C. Climate model projections under different greenhouse gas emission scenarios are often used to investigate snowpack response to climate change. However, in mountainous regions, climate models outputs cannot be confidently applied to simulate future snow processes because of differences in process response with grid cells resolution (e.g., Eum et al., 2014), and the very high uncertainty in projected changes to local scale precipitation due to coarse simulation of synoptic dynamics, unrealistic model treatment of orographic effects and use of regional-scale climate models, which show large simulation biases for downscaling when compared to current control conditions (Fowler et al., 2007).

Temperate zone snow models that rely on empirical temperature-index or related techniques have great difficulty in cold mountain regions (Swanson, 1998) and in general do not perform well because of their lack of physical basis (Walter et al., 2005). The Cold Regions Hydrological Modelling platform (CRHM) offers a complete range of processes for the Canadian Rockies (i.e. blowing snow, interception and sublimation of snow, energy balance snowmelt, slope radiation, canopy influence on radiation, canopy gap effect on snow), and the process algorithms have been extensively field tested (Pomeroy et al., 2012; Fang et al., 2013; Rasouli et al., 2015). CRHM includes algorithms rarely found in hydrological models such as those for calculating shortwave radiation through forest canopies on slopes (Ellis and Pomeroy, 2007), longwave radiation from partly cloudy skies and mountain terrain (Sicart et al., 2006), enhanced longwave emissions from canopies (Pomeroy et al., 2009), snow surface temperature (Ellis et al., 2010), canopy gap radiative transfer, forest canopy snow interception, sublimation, drip and unloading (Ellis et al., 2010, 2011, 2013), alpine blowing snow transport and sublimation (MacDonald et al., 2010), and alpine snowmelt and snowmelt runoff (DeBeer and Pomeroy, 2010).

This paper focusses on the sensitivity of snow processes to temperature increases up to the maximum range of increase expected in in the Canadian Rockies in the 21st Century. Such an approach can provide a useful understanding of the impact of various degrees of warming on mountain snow dynamics at various elevations in an intensively-researched and well-documented catchment where hydrological models have been designed to simulate the major snow processes with physical fidelity. The objective is to examine how snow processes and dynamics operate and snow states change as air temperatures increase in a cold regions mountain catchment. The catchment chosen for analysis is Marmot Creek Research Basin, in the headwaters of the South

Saskatchewan River and upstream of several major cities (Calgary, Medicine Hat, Saskatoon), hydroelectricity generating plants and irrigation districts in western Canada.

STUDY SITE

The study was conducted in the Marmot Creek Research Basin (MCRB) (50˚57'N, 115˚09'W), Kananaskis Valley, Alberta, Canada, located in the Front Ranges of the Canadian Rockies (Figure 1). Marmot Creek is a tributary of the Kananaskis River and is a headwater basin of the Bow River basin. The MCRB totals 9.4 km^2 area and elevation ranges from 1600 m.a.s.l. to 2825 m.a.s.l. Much of MCRB is covered by needleleaf vegetation which is dominated by Engelmann spruce (*Picea engelmanni*) and subalpine fir (*Abies lasiocarpa*) in the higher elevations and lodgepole pine (*Pinus contorta* var. Latifolia) in the lower elevations (Kirby and Ogilvy, 1969). Forest management experiments conducted in the 1970s and 1980s left large and small clearings (Golding and Swanson, 1986). Alpine larch (*Larix lyallii*) and short shrub are present around the treeline at approximately 2180 to 2250 m.a.s.l., and exposed rock surface and talus are present in the high alpine part of the basin. Continental air masses control the weather in the region, which has long and cold winters and cool and wet springs – there is a precipitation maximum in spring and early summer which can occur as rainfall or snowfall. Westerly warm and dry Chinook (foehn) winds lead to brief periods with the air temperature above 0 ˚C during the winter months. Annual precipitation ranges from 600 mm at lower elevations to more than 1100 mm at the higher elevations, of which approximately 70 to 75% occurs as snowfall with the percentage increasing with elevation (Storr, 1967). Mean monthly air temperature ranges from 14 ˚C in July to -10 ˚C in January.

Field observations to force the model in this paper were collected from two elevations: Fisera Ridge (2325 m.a.sl.) and Upper Clearing (1845 m.a.s.l.) and include air temperature, relative humidity, wind speed, precipitation, soil temperature, and incoming shortwave radiation. The measurements are described in several publications (DeBeer and Pomeroy, 2010; Ellis et al., 2010; MacDonald et al., 2010; Pomeroy et al., 2012; Fang et al., 2013). Precipitation was measured with an Alter-shielded Geonor weighing precipitation gauge, corrected for wind-induced undercatch.

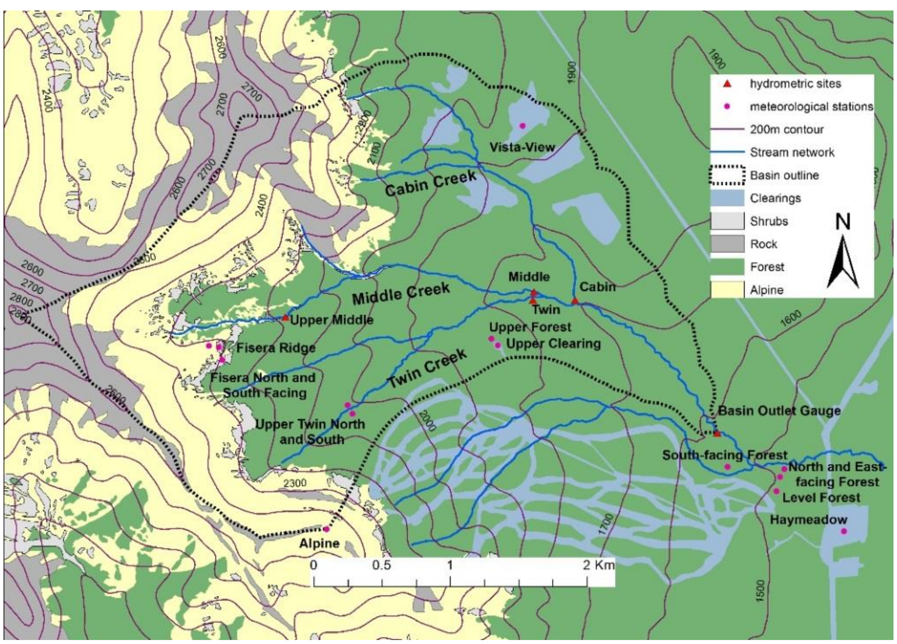

Figure 1. Marmot Creek Research Basin, Alberta, Canada showing land cover and instrumentation.

MODEL

The Cold Regions Hydrological Modelling platform (CRHM) was used to develop a snow dynamics model to simulate the primary snow processes in alpine and forested environments at the MCRB. CRHM is an object-oriented, modular and flexible platform for assembling physically based hydrological models (Pomeroy et al., 2007). Figure 2 shows the schematic setup of these modules, which include:

1) Observation module: reads the forcing meteorological data (temperature, wind speed, relative humidity, vapour pressure, precipitation, and radiation), adjusting temperature with environmental lapse rate and precipitation with elevation and wind-induced undercatch, estimating precipitation phase according to the psychrometric energy balance method of Harder and Pomeroy (2013) and providing these inputs to other modules.

2) Solar radiation module (Garnier and Ohmura, 1970): calculates the theoretical global radiation, direct and diffuse solar radiation, as well as maximum sunshine hours based on latitude, elevation, ground slope, and azimuth, providing radiation inputs to the slope radiation module, the longwave module, the canopy module and the energy-budget snowmelt module.

3) Slope radiation module: estimates incident short-wave to a slope using measurement of incoming short-wave radiation on a level surface. The measured incoming short-wave radiation from the observation module and the calculated direct and diffuse solar radiation from the radiation module are used to calculate the ratio for adjusting the short-wave radiation on the slope.

4) Long-wave radiation module (Sicart et al., 2006): estimates incoming long-wave radiation using measured short-wave radiation. This is inputted to the energy-balance snowmelt module.

5) Albedo module (Verseghy, 1991): estimates snow albedo throughout the winter and into the melt period and also indicates the beginning of melt for the energy-balance snowmelt module.

6) Canopy module (Ellis et al., 2010; Pomeroy et al., 1998): estimates the snowfall and rainfall intercepted by the forest canopy and updates the under-canopy snowfall and rainfall and calculates short-wave and long-wave sub-canopy radiation. This module has options for open environment (no canopy adjustment of snow mass and energy), small forest clearing environment (adjustment of snow mass and energy based on diameter of clearing and surrounding forest height), and forest environment (adjustment of snow mass and energy from forest canopy).

7) Blowing snow (PBSM) module (Pomeroy and Li, 2000): simulates the wind redistribution of snow transport and blowing snow sublimation losses throughout the winter period.

8) Energy-balance snowmelt module (Marks et al., 1998): this is a version of the Snobal model developed to simulate the mass and energy balance of deep mountain snowpacks. This module estimates snowmelt and flow through snow by calculating the energy balance of radiation, sensible heat, latent heat, ground heat, advection from rainfall, and the change in internal energy for snowpack layers consisting of a top active layer and basal layer.

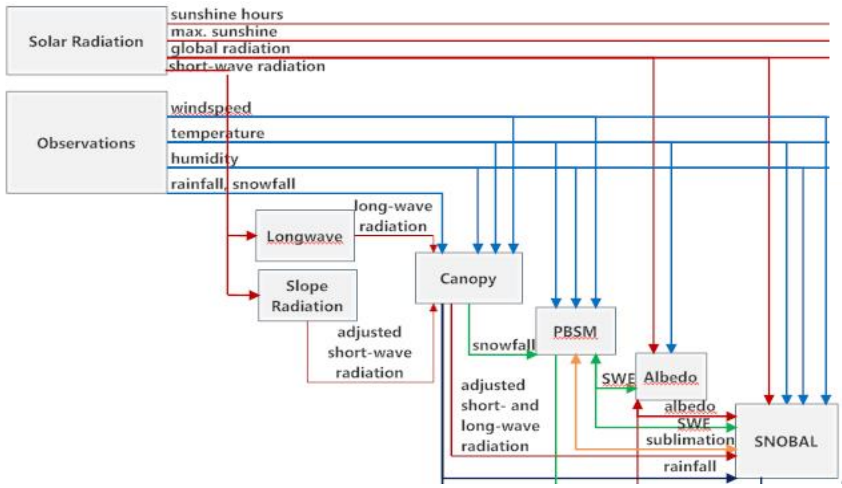

Figure 2. CRHM snowpack model, showing modules and flows of information. Red lines denote radiation term flows, blue lines denote meteorological variable flows, green lines denote snow flows, orange lines denote snow sublimation loss, and black lines denote liquid water flows.

METHODOLOGY

The model was parameterised from local measurements of topography and forest canopy, then set up and evaluated against detailed and spatially distributed snow survey measurements as described in Fang et al. (2013). Mean biases and root mean square errors from these evaluations against bi-weekly snow surveys were less than 10%. For the climate warming evaluation, the model was then run using high quality hourly measurements over nine years (2005-2014). Air temperatures were then perturbed by adding integer increases to the original hourly time series of air temperature measurements over nine years. These air temperature perturbations affected many calculations in the model, principally those for water vapour pressure (relative humidity was held constant), incoming longwave radiation from the atmosphere, terrain and forest canopy, snowpack emission of longwave radiation, precipitation phase, threshold wind speed for blowing snow, sublimation of blowing snow, albedo decay, canopy sublimation, drip and unloading of snow and turbulent fluxes of sensible and latent heat to snowpacks.

Snow fluxes and states were extracted for three characteristic environments in MCRB – the alpine environment in the upper elevations of the catchment, a mid-elevation spruce-fir forest and a small (100 m) clearing at the same elevation as the forest. The alpine environment is a spatially-weighted compilation of exposed north-facing and south-facing slopes and ridgetops near Fisera Ridge at 2325 m.a.s.l. This site is heavily wind-scoured and much snow is eroded and transported to treeline forests in the basin (MacDonald et al., 2010; Musselman et al., 2015). The forest and clearing are level sites near the Upper Clearing at 1845 m.a.s.l. Though there are important differences in clearing snow energetics on slopes compared to level sites (Ellis et al., 2011; 2013), the response of level clearings to warming will be intermediate to those on north and south facing slopes. There are minimal differences in forest snow energetics between north and south facing slopes in this environment (Ellis et al., 2013) and so forest responses from a level site should represent that on a range of slopes quite well.

RESULTS AND DISCUSSION

Results were obtained for hydrological years starting 1 October over nine years. The initial impact of warming on snow is on snowfall and rainfall and this is strongly impacted at all elevations as is shown in Figure 3. With a temperature increase of 5 °C, alpine snowfall decreases 350 mm or 42%, whereas the forest and clearing snowfall decrease by 250 mm or 55%. Similarly, rainfall increases were greatest in the alpine and smaller at the forest and clearing for 5 °C of warming, noting that changes in rainfall must match those in snowfall as annual precipitation does not change in these simulations. The larger proportional decreases in snowfall at the lower elevations of the forest and clearing are due to the warmer temperatures during snowfall, and the smaller decreases in snowfall are due to lower annual precipitation depths at these elevations. Initial rainfall ratios for the alpine and forest/clearing elevations were 0.29 and 0.47 respectively, rising to 0.59 and 0.76 with 5 °C of warming. As a result, a catchment whose precipitation is now dominated by snowfall becomes rainfall dominated with a warmer climate. The implications of this on the snow hydrology of MCRB are profound, even a very small warming causes the lower elevations to become rainfall dominated and warming of 3 °C causes a transition in the alpine precipitation from primarily snowfall to rainfall. This can increase the prevalence of rain-on-snow melting events and also promote higher threshold wind speeds for blowing snow initiation and more frequent unloading of intercepted snow from the forest canopy. As much of the catchment's precipitation occurs in spring as snowfall, the reduction in spring snowfall can promote a much earlier snow-free period and lower streamflow generation as was experienced in 2015 (Hume, 2015). How this transition and others associated with changing snow hydrological processes affect the peak snowpack and melt rates is explored next.

The impacts of warming on the snowpack regimes in the alpine, forest and clearing environments are shown in Figure 4. There is an important seasonality to these impacts that differs with elevation. In the alpine environment, peak SWE in snowpacks occurs in the current climate on 14 May with

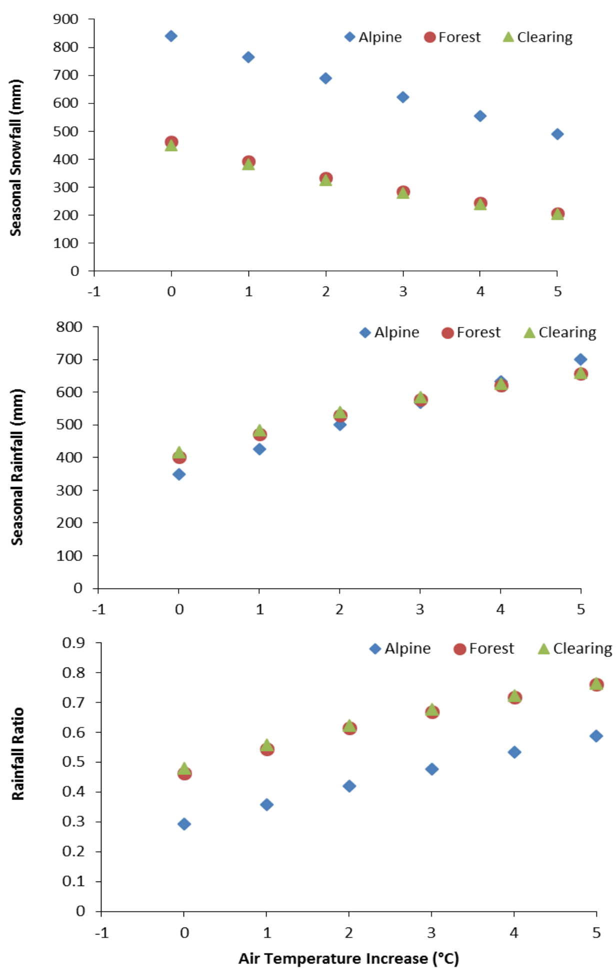

Figure 3. Seasonal snowfall and rainfall and the rainfall ratio at three ecozones in MCRB as calculated for the current climate (2005-2015) and with perturbations to air temperature.

85 mm. Snow accumulation in the cold mid-winter is not strongly affected by warming and by mid-March there are still small (<30 mm SWE) accumulations under both current and 5 °C of warming simulations. However, large differences due to warming develop after 1 April, the primary snow accumulation season in the alpine, where relatively warm, wet snow accumulates with limited redistribution. By 1 May under the current climate 78 mm SWE accumulates and depletion occurs almost two months later, but with 5 °C of warming this drops to 14 mm and snow depletion occurs by mid-May. In the alpine, the mid-winter snowpack is insensitive to warming but the spring snowpack is very sensitive, leading to an advance of snowpack depletion by approximately one week per degree of warming. In contrast, the clearing and forest SWE undergo substantial decreases

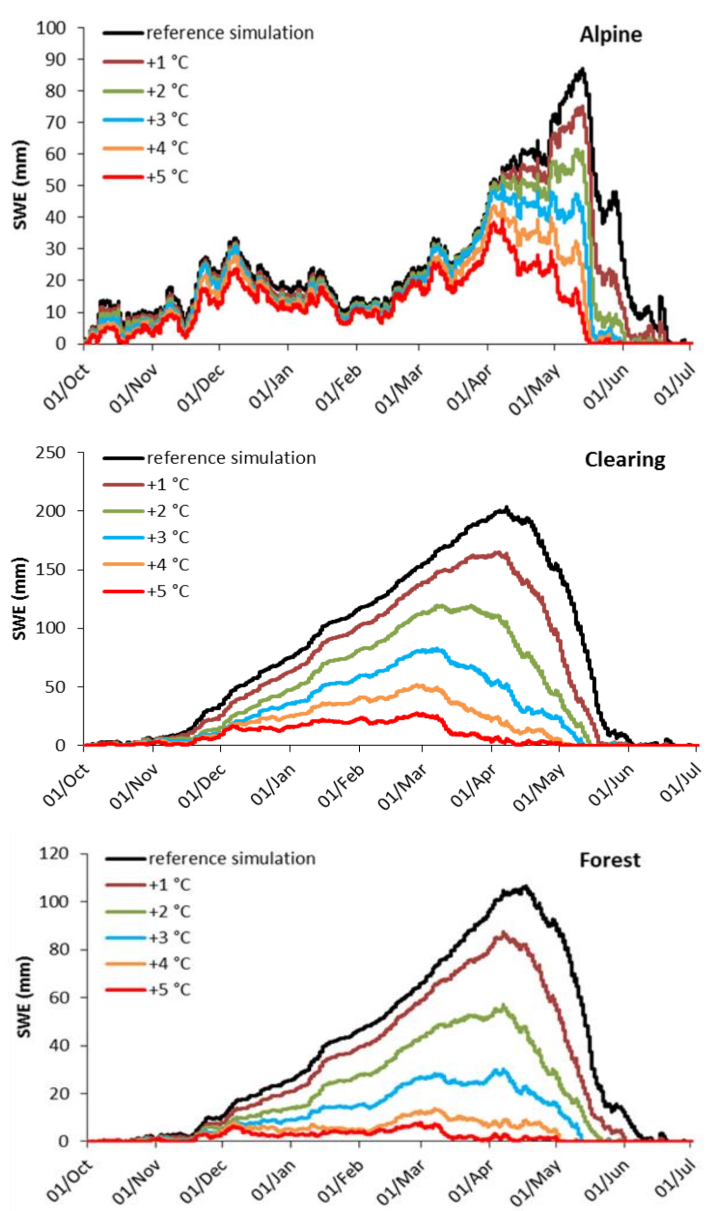

Figure 4. Snow accumulation and ablation regimes for three ecozones in MCRB as calculated for the current climate (2005-2015) and with perturbations to air temperature.

throughout the winter with 5 °C of warming, with almost complete disappearance of the initially 106 mm peak SWE forest snowpack and reduction of the clearing snowpack from a peak of 200 with the current climate to 25 mm SWE. With 5 °C of warming, the date of snowpack depletion advances from mid-June to early May in the forest and from early June to early May in the clearing. Snowpack depletion is more synchronous as temperature warms with a 20 day difference amongst sites narrowing to a 10 day difference with 5 °C of warming.

There is partial compensation for the impacts of warming on precipitation phase through its impact on snow redistribution and sublimation processes. For instance, the alpine receives very shallow mid-winter snow accumulation despite the large snowfall depth over the winter because wind transport redistributes snow from the exposed alpine ridges to the treeline and some blowing snow sublimates in transit (Musselman et al., 2015). Warming is expected to suppress blowing snow transport and sublimation which should partially compensate for reduced snowfall. Figure 5 shows that alpine ablation due to blowing snow transport (288 mm) and sublimation (289 mm) is reduced 47% and 35%, respectively due to warming of 5 °C. Sublimation losses from intercepted snow in the forest canopy (40 mm) are reduced by 71% by warming, most likely due to unloading of snow under warmer winter air temperatures.

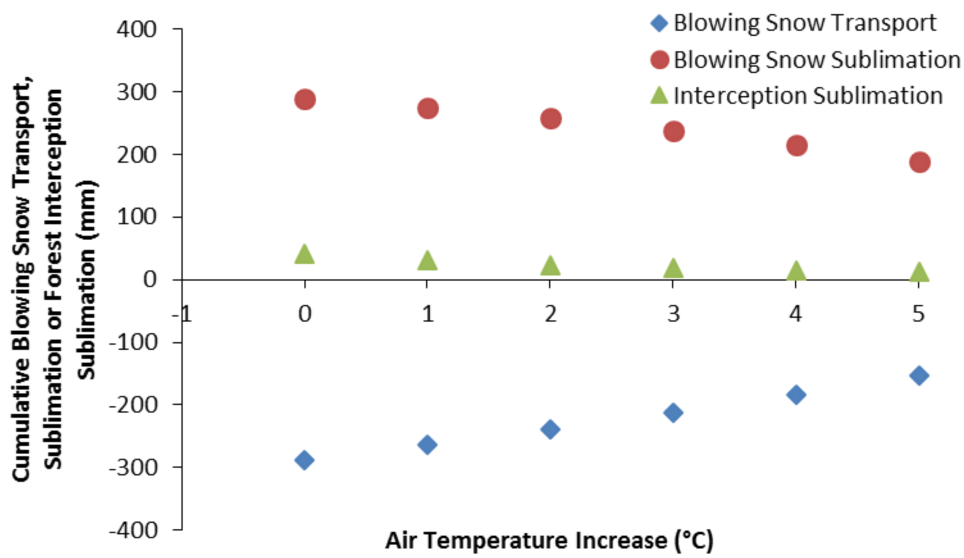

Figure 5. Snow redistribution and sublimation fluxes at MCRB with warming. Blowing snow transport and sublimation in the alpine are shown as negative and positive respectively when a mass loss to snowpack. Intercepted snow sublimation in the forest is shown as positive when a mass loss from the snowpack.

The influence of warming on melt rates is complex because air temperature increases albedo decay and hence net shortwave radiation, net longwave radiation and turbulent energy transfer to snow as well as reducing snowfall and snow redistribution as previously discussed. The melt rate was estimated by dividing the peak SWE by the number of days from peak SWE to snowpack ablation for each site and is shown in Figure 6 for the alpine, forest and clearing environments. At all sites the snowmelt rate tended to decrease with increasing air temperature, with the most dramatic decreases for 5 °C of warming found at the clearing, from 3.7 to 0.4 mm/day, and the smallest decrease found at the alpine, from 2.2 to 1.0 mm/day. Though a decrease in melt rate with increasing air temperature is counterintuitive and certainly would not be captured by a temperature index melt model, warming conditions reduce the snowpack available to melt and advance the melt timing into earlier periods when insolation is lower. For instance warming of 5 °C shifts alpine peak SWE forward by 36 days, forest by 47 days and clearing by 40 days.

Similarly, warming of 5 °C shifts alpine snowpack depletion forward by 37 days, forest by 43 days and clearing by 32 days.

The various impacts of warming on snowfall, redistribution and ablation are complex, and sometimes counterintuitive and compensating. To put them into context, the fractional decreases in snowfall, peak SWE, blowing snow transport, blowing snow sublimation, intercepted snow sublimation and melt rate were calculated per 1 °C of warming and are shown in Table 1. Snowfall and peak SWE decline the most rapidly with warming in the forest and then the clearing and least rapidly in the alpine. Blowing snow ablation processes (transport and sublimation) common in the alpine are more sensitive to warming than intercepted snow ablation (sublimation) processes found in the forest. Melt rate is reduced most rapidly with warming under forest canopies followed by clearings with much slower reductions found for the alpine environment. This is likely due to the strong control that air temperature exerts on incoming longwave radiation under forest canopies as opposed to environments with high sky view factors. The advance of the snow-free date with warming was fastest in the forest, followed by the alpine and then the clearing environments.

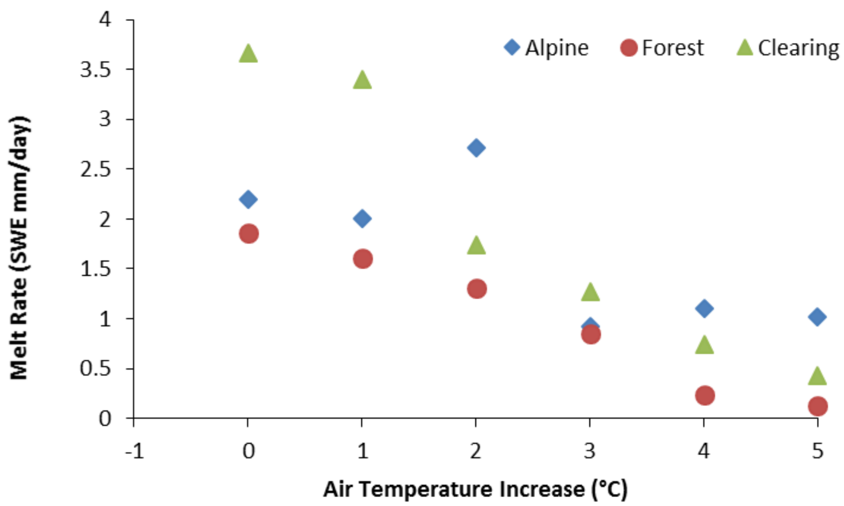

Figure 6. Melt rate (peak SWE divided by days to snowpack depletion) for the alpine, forest and clearing environments at MCRB under current climate and with warming.

Table 1. Fractional decrease in snowfall, peak SWE, blowing snow transport, intercepted snow sublimation, melt rate and snow free date per degree of warming for alpine, forest and clearing environments in MCRB compared to the current climate.

	Snowfall mm/°C	Peak SWE mm/°C	Blowing Snow Transport mm/°C	Blowing Snow Sublimation mm/°C	Intercepted Snow Sublimation mm/°C	Melt Rate mm / (°C day)	Snow Free Date day/°C
Alpine	8.3	11	9.4	7		10.8	6.4
Forest	11	18.6			14.3	18.7	7.4
Clearing	10.9	17.3				17.6	5.5

CONCLUSIONS

This study is a first attempt to quantify the impact of temperature increases on snow processes governing snowfall and snow accumulation and ablation in the Canadian Rockies. It was found that warming reduces low elevation snowpacks more than high elevation snowpacks and forest snow more than clearing snow. While the transition from snowfall to rainfall is dramatic and has important impacts on peak snow accumulation, snow redistribution and sublimation processes in alpine and forest environments are dampened by warming, and this counteracts the reduced snowfall impact on peak SWE to a small degree. Melt rates are reduced by warming, which is counter-indicated by simple temperature index melt calculations and reflects both shallower snowpacks and the substantial shift forward in the melt period into lower insolation periods with warming. A crucial finding in light of current discussions on "safe" limits to global warming is that 2 °C of warming will still lead to a shift from snowfall to rainfall dominance, a substantial decline in Canadian Rockies snowpacks and shortening of snow seasons at all elevations, but increases in temperature greater than this lead to dramatic declines in snowpacks and shortening of snow seasons. With 5 °C of warming, low elevation forest snowpacks are ephemeral and others reduced by an order of magnitude from current levels and snow free dates are advanced by from four to six weeks. Further work will determine the impact of this change on streamflow and soil moisture storage.

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REFERENCES

Brown RD, Robinson DA. 2011. Northern Hemisphere spring snow cover variability and change over 1922–2010 including an assessment of uncertainty. *The Cryosphere* **5**: 219–229.Davis RD, Lowit MB, Knappenberger PC, Legates DR. 1999. A climatology of snowfall-temperature relationships in Canada. *Journal of Geophysical Research* **104**(D): 11985-11994.

DeBeer CM, Pomeroy JW. 2010. Simulation of the snowmelt runoff contributing area in a small alpine basin. *Hydrology and Earth System Sciences* **14**: 1205-1219. doi:10.5194/hess-14-1205-2010.

Déry SJ, Brown RD. 2007. Recent Northern Hemisphere snow cover extent trends and implications for the snow-albedo feedback. *Geophysical Research Letters* **34**: L22504. doi:10.1029/2007GL031474.

Ellis CR, Pomeroy JW. 2007. Estimating sub-canopy shortwave irradiance to melting snow on mountain slopes. *Hydrological Processes* **21**: 2581–2597. DOI:10.1002/hyp.6794.

Ellis CR, Pomeroy JW, Brown T, MacDonald J. 2010. Simulation of snow accumulation and melt in needleleaf forest environments. *Hydrology and Earth System Sciences* **14**: 925–940.doi:10.5194/hess-14-925-2010.

Ellis CR, Pomeroy JW, Essery RLH, Link TE. 2011. Effects of needleleaf forest cover on radiation and snowmelt dynamics in the Canadian Rocky Mountains. *Canadian Journal of Forest Research* **41**: 608-620. doi:10.1139/X10-227.

Ellis CR, Pomeroy JW, Link TE. 2013. Modeling increases in snowmelt yield and desynchronization resulting from forest gap thining treatments in a northern mountain catchment. *Water Resources Research* **49**: 936-949. doi:10.1002/wrcr.20089.

Eum HI, Dibike Y, Prowse T, Bonsal B. 2014. Inter-comparison of high resolution gridded climate data sets and their implication on hydrological model simulation over the Athabasca Watershed, Canada. *Hydrological Processes* **28**: 4250–4271.

Fang X, Pomeroy JW, Ellis CR, MacDonald MK, DeBeer CM, Brown T. 2013. Multi-variable evaluation of hydrological model predictions for a headwater basin in the Canadian Rocky Mountains. *Hydrology and Earth System Sciences* **17**: 1635-1659. doi:10.5194/hess-17-1635-2013.

Fowler HJ, Blenkinsop S, Tebaldi C. 2007. Linking climate change modelling to impacts studies: recent advances in downscaling techniques for hydrological modelling. *International Journal of Climatology* **27**: 1547–1578.

Garnier BJ, Ohmura A. 1970. The evaluation of surface variations in solar radiation income. Solar Energy 13: 21–34.

Hume, M. 2015. Signs of drought appear to be in Western Canada for the long term. *Globe and Mail*, Monday June 15, 2015. Accessed 23 Sept 2015 at http://www.theglobeandmail.com/news/british-columbia/signs-of-drought-appear-to-be-in-western-canada-for-the-long-term/article24954511/?click=sf_globe

Golding DL, Swanson RH. 1986. Snow distribution patterns in clearings and adjacent forest. *Water Resources Research* **22**: 1931-1940.

Harder P, Pomeroy J. 2013. Estimating precipitation phase using a psychrometric energy balance method. *Hydrological Processes* **27**: 1901–1914. doi:10.1002/hyp.9799.

Harder P, Pomeroy JW, Westbrook CJ. 2015. Hydrological resilience of a Canadian Rockies headwaters basin subject to changing climate, extreme weather, and forest management. *Hydrological Processes* **29**: 3905-3924. Doi:10.1002/hyp.10596.

Kirby CL, Ogilvy RT. 1969. The forests of Marmot Creek watershed research basin. Canadian Forestry Service Publication No. 1259, Canadian Department of Fisheries and Forestry, Canadian Forestry Service, Ottawa, 37.

MacDonald MK, Pomeroy JW, Pietroniro A. 2010. On the importance of sublimation to an alpine snow mass balance in the Canadian Rocky Mountains. *Hydrology and Earth System Sciences* **14**: 1401-1415. doi:10.5194/hess-14-1401-2010.

Marks D, Kimball J, Tingey D, Link T. 1998. The sensitivity of snowmelt processes to climate conditions and forest cover during rain-on-snow: a case study of the 1996 Pacific Northwest flood. *Hydrological Processes* **12**: 1569–1587.

Mote PW, Hamlet AF, Clark MP, Lettenmaier DP. 2005. Declining mountain snowpack in western North America. *Bulletin of the American Meteorological Society* **86**: 39–49. DOI:10.1175/BAMS-86-1-39.

Musselman KN, Pomeroy JW, Essery RLH, Leroux N. 2015. Impact of windflow calculations on simulations of alpine snow accumulation, redistribution and ablation. *Hydrological Processes* **29**: 3983-3999. doi:10.1002/hyp.10595.

Nogués-Bravo D, Araújo MB, Errea MP, Martinez-Rica JP. 2007. Exposure of global mountain systems to climate warming during the 21st Century. *Global Environmental Change* **17**(3): 420–428.

Pomeroy JW, Gray DM, Shook KR, Toth B, Essery RLH, Pietroniro A, Hedstrom NR. 1998. An evaluation of snow accumulation and ablation processes for land surface modelling. *Hydrological Processes* **12**: 2339-2367.

Pomeroy JW, Li L. 2000. Prairie and Arctic areal snow cover mass balance using a blowing snow model. *Journal of Geophysical Research* **105**: 26619–26634.

Pomeroy JW, Gray DM, Brown T, Hedstrom NR, Quinton W, Granger RJ, Carey S. 2007. The Cold Regions Hydrological Model, a platform for basing process representation and model structure on physical evidence. *Hydrological Processes* **21**: 2650-2667. doi:10.1002/hyp.6787.

Pomeroy JW, Marks D, Link T, Ellis C, Hardy J, Rowlands A, Granger RJ. 2009. The impact of coniferous forest temperature on incoming longwave radiation to melting snow. *Hydrological Processes* **23**: 2513–2525.

Pomeroy J, Fang X, Ellis C. 2012. Sensitivity of snowmelt hydrology in Marmot Creek, Alberta, to forest cover disturbance. *Hydrological Processes* **26**: 1892-1905. doi:10.1002/hyp.9248.

Rasouli K, Pomeroy JW, Marks, DG. 2015. Snowpack sensitivity to perturbed climate in a cool mid-latitude mountain catchment. *Hydrological Processes* **29**: 3925–3940. doi: 10.1002/hyp.10587.

Shook K, Pomeroy J. 2012. Changes in the hydrological character of rainfall on the Canadian prairies. *Hydrological Processes* **26**: 1752-1766. DOI: 10.1002/hyp.9383.

Sicart JE, Pomeroy JW, Essery RLH, Bewley D. 2006. Incoming longwave radiation to melting snow: observations, sensitivity and estimation in northern environments. *Hydrological Processes* **20**: 3697–3708. doi:10.1002/hyp.63837.

Storr D. 1967. Precipitation variation in a small forested watershed. *Proceedings of the 35th Annual Western Snow Conference*, 18-20 April 1967, Boise, Idaho; 11-17.

Swanson RH. 1998. Forest hydrology issues for the 21st century: A consultant's viewpoint. *Journal of the American Water Resources Association* **34**(4): 755–763.

Valeo C, Xiang Z, Bouchart FJ, Yeung P, Ryan MC. 2007. Climate change impacts in the Elbow River watershed. *Canadian Water Resources Journal* **32**: 285–302. doi:10.4296/cwrj3204285.

Verseghy DL. 1991. CLASS-A Canadian land surface scheme for GCMs, I. soil model. *International Journal of Climatology* **11**: 111-133.

Walter MT, Brooks ES, McCool DK, King LG, Molnau M, Boll J. 2005. Process-based snowmelt modeling: does it require more input data than temperature-index modeling? *Journal of Hydrology* **300**: 65-75.

Woo M-K, Pomeroy J. 2012 Snow and Runoff: Processes, Sensitivity and Vulnerability. In *Changing Cold Environments: A Canadian Perspective*, French H, Slaymaker O (eds.) John Wiley & Sons, Ltd: Chichester, UK; 105-125. doi:10.1002/9781119950172.ch6

Zhang XB, Vincent LA, Hogg WD, Niitsoo A. 2000. Temperature and precipitation trends in Canada during the 20th Century. *Atmosphere-Ocean* **38**: 395-429.