Snowmelt Energy Flux Recovery During Rain-on-Snow in Regenerating Forests

by

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Abstract

Rain-on-snow (ROS) is a major contributor to flooding and landslides in many temperate coastal watersheds around the world. Research has shown that forest harvesting can increase melt rates during ROS at both the stand and watershed scale. Because of this, post disturbance hydrological recovery is of interest in watersheds where forest management is prevalent. Recent research that pairs events by frequency rather than chronologically has indicated that forest cover removal can have a significant effect on the magnitude and frequency of extreme events, which is counter to the dominant view in forest hydrology. Hydrological modelling provides a means to apply frequency based analysis in watersheds with short data records, but models must be tested and validated in coastal watersheds before they can be applied extensively. A key challenge to testing models is the inherent difficulty with collecting data in ROS environments. Therefore, the objectives of this research were to design a methodology that recorded previously unobserved processes, use these data to validate model simulations and assess stand level energy flux recovery during ROS. Data were collected at a range of elevations within recently harvested, regenerating and old growth forests. The Cold Regions Hydrological Model generally performed well at capturing the dynamics of snow accumulation and melt, however, snow water equivalent was generally over-predicted. Depths of transient snowpacks were generally under-predicted, however, once a snowpack was established model performance improved. Clear-cut forests had higher mean and greater variability of energy inputs resulting in large events occurring more frequently than in old or second growth forests. Energy flux recovery was evident within the regenerating forests; however, both the rates of recovery and differences among stands depended on the location and the variables compared. When either the mean or standard deviation of energy inputs differed from that of old growth forests, energy flux recovery was reduced as events became larger and less frequent. It is probable that results obtained from this study will translate to stream flow in watersheds with steep slopes, shallow soils and extensive preferential flow networks (i.e. high run-off coefficients), especially when run-off generating areas are synchronized.
Preface

A version of chapter 4 has been published: Floyd, W. and M. Weiler., 2008. Measuring snow accumulation and ablation dynamics during rain-on-snow events: innovative measurement techniques. Hydrol. Process. 22, 4805-4812. Dr. Markus Weiler helped with modifications to the camera and in overall study design and image analysis. This chapter has been modified from the original manuscript to reduce redundancies and to fit within the overall dissertation.

Chapter 5 will be submitted for publication with contributions from the following: Dr, Markus Weiler assisted in study design and interpretation of results, Dr. John Pomeroy assisted with model development; and both Dr. Younes Alila and Dr. Rob Hudson contributed to the study design and discussion.

Chapter 6 will be submitted for publication with contributions from the following: Dr. Younes Alila provided direction for data analysis and interpretation of results; Dr. Markus Weiler and Dr. Robert Hudson provided input to study design and interpretation of results.
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<td>$A$</td>
<td>advective heat from rainfall</td>
</tr>
<tr>
<td>$A_s$</td>
<td>slope</td>
</tr>
<tr>
<td>a.s.l.</td>
<td>above sea level</td>
</tr>
<tr>
<td>$ac$</td>
<td>radiation absorbed by ozone</td>
</tr>
<tr>
<td>$aw$</td>
<td>radiation absorbed by water</td>
</tr>
<tr>
<td>CC</td>
<td>clear-cut</td>
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<tr>
<td>CDF</td>
<td>cumulative distribution function</td>
</tr>
<tr>
<td>$C_e$</td>
<td>intercepted snow exposure coefficient</td>
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<td>CRHM</td>
<td>Cold Regions Hydrological Model</td>
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<tr>
<td>$C_i$</td>
<td>heat capacity of ice (2.102e-3 MJ/kg/K)</td>
</tr>
<tr>
<td>$C_l$</td>
<td>canopy leaf contact area per unit ground</td>
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<td>CP</td>
<td>chronological pairing</td>
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<tr>
<td>$e_a$</td>
<td>vapour pressure</td>
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<td>ECA</td>
<td>equivalent clear-cut area</td>
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<td>FP</td>
<td>frequency pairing</td>
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<td>ground heat</td>
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<td>sensible heat</td>
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<td>$H_a$</td>
<td>hour angle from solar noon</td>
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<td>$h_m$</td>
<td>snow depth</td>
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<td>$H+Lv$</td>
<td>turbulent energy, combination of latent and sensible heat</td>
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<td>$H_s$</td>
<td>heat of sublimation (2.838e-6 MJ/kg)</td>
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<td>$I$</td>
<td>intensity of radiation from space</td>
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<td>rainfall interception rate</td>
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<td>$I_s^*$</td>
<td>maximum intercepted snow load</td>
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<tr>
<td>$K_{inF}$</td>
<td>net shortwave radiation reaching the forest floor</td>
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<td>leaf area index</td>
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<tr>
<td>$L_{inF}$</td>
<td>longwave radiation from the forest canopy</td>
</tr>
<tr>
<td>$L_o$</td>
<td>longwave irradiance in open environments</td>
</tr>
</tbody>
</table>
LR  Lower Russell (520-560m a.s.l.)

$M_o$  optical air mass

$m$  rank of observation

$M$  melt loss rate

MB  model bias

$M_{max}$  maximum potential melt assuming all $Qn$ is directed towards a ripe snowpack

$N$  number of observations

$n$  length of record in years

NS  Nash-Sutcliffe model efficiency index

OG  old growth

$P$  precipitation [mm]

$p$  atmosphere’s mean zenith path transmissivity

PDF  probability density function

PMA  percent model agreement

PNW  Pacific Northwest

$P_r$  rainfall

$P_s$  snowfall rate

$Q_{cc}$  Cold content of snow pack

$Q_{dif}$  diffuse clear sky radiation

$Q_{dir}$  solar radiation on a sloping surface

$q_e$  canopy sublimation flux

$Q_{ext}$  extraterrestrial radiation on a horizontal surface at the top of the earth’s atmosphere

$Q_M$  total energy to melt snow

$Qn$  energy available for snowmelt (does not include ground heat or advective heat)

$r$  correlation co-efficient

$R_d$  rate of canopy drip

$RH$  relative humidity

$R_l$  longwave radiation

RMSE  root mean square error

$Rn$  net radiation

ROS  rain-on-snow

$R_s$  shortwave radiation
<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$S$</td>
<td>maximum snow load per unit branch</td>
</tr>
<tr>
<td>SWE</td>
<td>snow water equivalent</td>
</tr>
<tr>
<td>SWE&lt;sub&gt;o&lt;/sub&gt;</td>
<td>antecedent snow water equivalent</td>
</tr>
<tr>
<td>$t$</td>
<td>time step for calculation</td>
</tr>
<tr>
<td>$T_a$</td>
<td>air temperature</td>
</tr>
<tr>
<td>$T_f$</td>
<td>forest temperature</td>
</tr>
<tr>
<td>$T_s$</td>
<td>soil temperature</td>
</tr>
<tr>
<td>$T_i$</td>
<td>ice bulb temperature</td>
</tr>
<tr>
<td>$T_m$</td>
<td>melting point of snow</td>
</tr>
<tr>
<td>$T_r$</td>
<td>return interval for an event of given magnitude</td>
</tr>
<tr>
<td>$T_s$</td>
<td>average temperature of snowpack</td>
</tr>
<tr>
<td>$T_{us}$</td>
<td>user defined ice bulb temperature threshold that unloads solid snow</td>
</tr>
<tr>
<td>$T_{uw}$</td>
<td>user defined ice bulb temperature that unloads snow as water</td>
</tr>
<tr>
<td>$u$</td>
<td>wind speed [m/s]</td>
</tr>
<tr>
<td>$U_i$</td>
<td>rate of canopy unloading</td>
</tr>
<tr>
<td>US</td>
<td>Upper Stephanie (740-770m a.s.l.)</td>
</tr>
<tr>
<td>$v$</td>
<td>sky view factor</td>
</tr>
<tr>
<td>VIC</td>
<td>Variable Infiltration Capacity model</td>
</tr>
<tr>
<td>$V_s$</td>
<td>sublimation flux for a 500 $\mu$m radius ice sphere</td>
</tr>
<tr>
<td>WAR</td>
<td>water available for runoff</td>
</tr>
<tr>
<td>$x_{avg}$</td>
<td>mean value of $n$ number of $x_{obs}$</td>
</tr>
<tr>
<td>$x_{obs}$</td>
<td>observed value</td>
</tr>
<tr>
<td>$x_{sim}$</td>
<td>simulated value</td>
</tr>
<tr>
<td>$Z$</td>
<td>aspect</td>
</tr>
<tr>
<td>$\delta$</td>
<td>declination of the sun</td>
</tr>
<tr>
<td>$\theta$</td>
<td>latitude</td>
</tr>
<tr>
<td>$\tau_{atm}$</td>
<td>atmospheric transmissivity to solar radiation on a cloudy day</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>Stephan-Boltzmann constant ($5.67 \times 10^{-8}$W m$^{-2}$K$^{-4}$)</td>
</tr>
<tr>
<td>$\rho_s$</td>
<td>density of falling snow</td>
</tr>
<tr>
<td>$\rho_w$</td>
<td>density of water (1000kg m$^{-3}$)</td>
</tr>
<tr>
<td>$u_z$</td>
<td>estimated canopy wind speed</td>
</tr>
</tbody>
</table>
\( \psi \)  
\( \tau_{fs} \)  
\( \theta_s \)  
\( \varepsilon_f \)  
\( \alpha \)  
\( \alpha_{\text{max}} \)  
4m  
11m  
13m  

- canopy wind speed extinction coefficient
- forest shortwave transmittance
- solar angle above the horizon
- forest thermal emissivity
- snow albedo
- albedo reset value when minimum snowfall is met
- 4m tall forest
- 11m tall forest
- 13m tall forest
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Chapter 1: Introduction

Watersheds are often characterized by the dominant hydro-climatic regime generating runoff, mainly rain, rain-on-snow (ROS) or snowmelt. On a global scale, rain is the principal form of precipitation, however, in almost all regions of the world, snow plays a role in storage and release of water to streams and rivers. Of particular interest is when rain falls on snow during active snowmelt and enhances runoff processes at the hillslope and watershed scale, contributing to flooding and landslides in both maritime (Jones and Perkins, 2010; Marks et al., 1998; Mazurkiewicz et al., 2008; McCabe et al., 2007; Sui and Koehler, 2001) and continental environments (e.g. Marks et al., 2001; Marks et al., 2002; Merz and Blöschl, 2003). The effects of ROS are generally greatest when accompanied by high amounts of energy when snowpacks are either shallow and/or have a small amount of cold content (Maclean et al., 1995; Marks et al., 1998; Marks et al., 2001). It is well documented that coastal watersheds in temperate climates with seasonal snowpacks are particularly sensitive to climate change, and as temperatures warm, ROS events will not only become more common, but they may become more extensive in regions where snowmelt was historically the dominant producer of peak streamflows (Adam et al., 2009; Brown and Mote, 2009; Herbst and Cooper, 2010; Jefferson, 2011; Leung et al., 2004; Nolin and Daly, 2006). It is also probable that watersheds with limited relief which routinely receive ROS only at higher elevations, will transition to completely pluvial environments as temperatures warm. Improving our understanding of ROS processes will be integral to climate change adaption strategies aimed at mitigating potential impacts to peak flows, water supply and natural hazard.

Of particular importance to ROS processes is the influence of forest cover removal on snow accumulation and melt and the subsequent effects on streamflow. In general, it has been shown that removal of forest cover results in deeper snowpacks and higher melt rates at the stand level (Berris and Harr, 1987; Christner and Harr, 1982; Harr, 1981; Harr and Coffin, 1992; Hudson, 2000a; Marks et al., 1998; Storck, 2000; Storck et al., 2002), however, these increases do not always translate to significant increases in streamflow at the watershed scale. This can be partially explained by the complex nature of watersheds and the requirement for synchronization of runoff generating areas to produce high or
extreme flows (e.g. Jones and Perkins, 2010). Because we cannot create effective controls on both climate inputs and the underlying watershed characteristics that affect streamflow (i.e. soils, bedrock geology and topography), it has been challenging to directly link land use changes to increases in discharge, especially for extreme events. With that said, in coastal watersheds with steep mountainous terrain, shallow soils, and high runoff coefficients, it seems logical that increases in snowmelt during ROS due to forest harvesting could translate to higher streamflows, especially when source areas are synchronized. Watersheds with these characteristics are common in coastal regions with a history of glaciation, such as in Washington, British Columbia and Alaska. Forest harvesting is prevalent within watersheds in these areas and they routinely experience ROS events associated with flooding and landslides (e.g. Marks et al., 1998; Guthrie et al., 2008).

One of the fundamental questions asked by land managers and scientists is: At what threshold does forest cover removal lead to increases in streamflow and negative impacts on water quality, stream channel morphology, fish habitat and engineered structures such as bridges and culverts? Answering this question has proved difficult and forms the basis of one of the biggest debates in forest hydrology (Jones and Grant, 1996; Jones and Grant, 2001; Thomas and Megahan, 1998, Beschta et al., 2000). More recently the debate has resurfaced with the ability to use flood frequency based analysis on long term streamflow records in place of traditional Chronological Pairing (CP) and ANOVA/ANCOVA data analysis (Alila et al., 2010; Alila et al., 2009; Lewis et al., 2010). Results using frequency analysis have shown that not only does forest harvesting impact the magnitude of events, but also their return intervals, with the largest or most extreme flows being very sensitive to even slight increases in discharge (Alila et al., 2009; Kuraś et al., 2012, Green and Alila, 2012). This is a fundamental shift from the dominant view in forest hydrology that forest harvesting only affects small and moderate flows (Bathurst et al., 2011; Beschta et al., 2000; Grant et al., 2008; Jones and Perkins, 2010; Thomas and Megahan, 1998). Further research is required to determine if this is generally true and if so, management strategies must be developed to determine what limits should be placed on forest harvesting. It also enhances the need to assess rates of watershed recovery following disturbance to provide guidance for future management.
Physically-based hydrological models supplemented with experimental results have proved useful to determine the potential cumulative hydrological effects of forest management (e.g. Alila and Beckers, 2001; Beckers and Alila, 2004; Kuraś et al., 2012; Schnorbus and Alila, 2004; Storck, 1998; Whitaker et al., 2003). Hydrological models enable researchers to focus on specific processes, such as interception, flow routing and groundwater dynamics to assess the impact of forest management on streamflow. Within British Columbia this approach has been successfully applied to interior watersheds where spring snowmelt is the primary generator of peak flows (e.g. Alila and Beckers, 2001; Kuraś et al., 2012; Schnorbus and Alila, 2004; Whitaker et al., 2003); however, its application in coastal watersheds dominated by ROS is limited to Oregon and Washington (e.g. Storck et al., 1998; Storck, 2000; Waichler et al., 2005).

The availability of meteorological and hydrometric data imposes strict limitations on the applicability of hydrological models to assess the effects of forest management and the subsequent hydrological recovery on streamflows. The spatial density of representative sites is very sparse in most areas and the length of data records tends to be short. There is also a high degree of uncertainty surrounding model inputs, general model structure and observations (Essery et al., 2009; Rutter et al., 2009), especially in watersheds with transient snowpacks. For example, there may be instances when the uncertainties in streamflow measurements are greater than the effects trying to be detected, such as peak flows which can be up to 50% higher or lower than the rating curves suggest. This makes it very difficult to formally assess the potential impacts of forest harvesting. Because of uncertainty, a great deal of work must still be done before physically-based hydrological models can be applied with confidence at the landscape scale (Essery et al., 2009; Rutter et al., 2009). In ROS environments, this uncertainty is especially high due to the difficulty collecting information related to snow accumulation and melt (e.g. Coffin and Harr, 1992; Storck, 2000).

ROS events often occur over short time periods and create dense, shallow snowpacks making it very difficult to access remote locations for direct observation. These conditions require monitoring methods that acquire high resolution temporal data which can be applied over large areas to capture spatial variability in mountainous watersheds. Traditionally, continuous monitoring of meteorological conditions has been
limited to meteorological stations which can measure precipitation, air temperature, relative humidity, wind speed, radiation (longwave and shortwave), snow depth with ultrasonic sensors and snow water equivalent (SWE) with snow pillows. These installations can be capital intensive, thus networks are limited, even in research watersheds. Some studies in ROS environments have installed snowmelt lysimeters to supplement existing networks (e.g. Beaudry and Golding, 1983; Coffin and Harr, 1992; Storck 2000; Storck et al., 2002), however, drains and tipping buckets often freeze, reducing confidence in measurements (e.g. Coffin and Harr, 1992). Manual snow courses can be used to monitor depth and SWE, but, they are labour intensive and usually temporally coarse (e.g. Hudson, 2000a). Time-lapse photography has also been used in ROS environments (Berris and Harr, 1987), however, at the time analysis techniques were limited making it difficult to derive quantitative data from the images. There have, however, been more recent technological advances in both the collection of time-lapse photography and photogrammetric analysis, allowing this type of approach to become more prevalent in fields such as geomorphology (e.g. Dexter and Cluer, 1999; Hancock and Willgoose, 2001; Lawler, 2005; Sheets et al., 2002). Combining traditional tools of measurement with innovative technologies may present opportunities to collect new types of data to improve our understanding of ROS processes and to test models using previously unavailable observations.

Despite limited data and model uncertainty, an operational requirement exists to develop tools to mitigate the potential negative effects of forest harvesting on streamflow. In British Columbia this is accomplished primarily by assessing the stand-level hydrological status within watersheds and using that information to impose limits on forest cover removal. The simplest and most widely used approach in British Columbia to assess the cumulative impact of canopy removal on peak flows is to determine the equivalent clear-cut area (ECA) within the watershed of concern (MoF, 2001). The ECA is an index based on comparisons of snow accumulation and ablation rates in forested and clear-cut stands using tree height and canopy closure as the metric for recovery. Within British Columbia, research on stand level hydrological recovery used to derive the ECA has focused exclusively on the role of interception and differences in accumulation and snowmelt (e.g. Winkler et al., 2005; Hudson and Horel, 2007). The ECA is used in
combination with other factors, such as dominant climate regime, topography, road density, pedology, disturbance of riparian areas and active fluvial units, and geology to determine cumulative impacts of forest harvesting on streamflow in the watershed. Because of its simplicity and limited data requirements it has been used extensively throughout the province. Despite the limitations and assumptions associated with ECA, its use will continue until better models are developed and thus it requires continual improvement based on the best available science.

Research to refine ECA methods for ROS environments has been primarily limited to coastal British Columbia, with indices based on elevation and tree height (Hudson, 2000a; Hudson, 2000b; Hudson, 2001; Hudson, 2003; Hudson and Horel, 2007); however, the methods used to measure accumulation and ablation during ROS were temporally coarse and did not use data to describe the physical processes governing interception and accumulation/ablation of snow. Furthermore, ECA is only an indicator and it does not explicitly link forest cover removal to increases in streamflow. If additional meteorological data were available in watersheds where ROS events occur, a physically-based interception, accumulation and ablation model could be tested and validated for ROS events in regenerating forests and provide valuable information to assess stand level hydrological recovery. This information could then be used to improve physically-based hydrological models, thereby allowing incorporation into methodologies to assess recovery at the watershed scale.

There is a clear need to improve our understanding of ROS processes due to the potential increase in its spatial and temporal extent as a result of a changing climate. This includes the increased use of hydrological models to identify gaps in our understanding and to provide direction for adaptation strategies and landuse planning. A knowledge gap exists in the effect of forest cover removal on streamflow, especially for extreme events. New analysis methods have begun to shed light on the relationship between forest harvesting and the impact on return intervals, highlighting the need to compare the mean, variability and frequency of snowmelt and streamflows among treatments. Limited information is available on how snowmelt during ROS in regenerating forests compare to old growth or mature stands. To fill these information gaps, existing models will need to be tested in new areas using new types of data to identify their strengths and weaknesses.
If uncertainty can be reduced to acceptable levels, then modelling results can be used to guide future research and land management strategies. It is with these needs in mind that I propose a project to answer the following primary research question:

*What are the similarities and differences in energy available for snowmelt during ROS in recently harvested, regenerating and old growth forests?*

In order to answer the primary research question, this dissertation has the following objectives:

1. Design an innovative method to monitor ROS events that can be applied across a wide range of locations;

2. Test and validate a physically-based interception and snowmelt model using data collected from the innovative method developed above;

3. Use results from the validated model to describe similarities and differences in the frequency of energy sources available for melt during ROS among different forest types.

This dissertation is composed of 5 additional chapters. Chapter 2 provides background information on forest canopy interception of rain and snow, methods to measure interception, models used for interception, snow accumulation and melt, current stand level models used to assess hydrological recovery, and methods used to assess differences in hydrological variables. Chapter 3 describes the study area and design common to all chapters. Innovative methods designed to monitor ROS as part of this project and a discussion of strengths and weaknesses of the design are found in Chapter 4. Chapter 5 describes the application of a physically-based interception and snowmelt model using observations from harvested, regenerating and old growth forests collected via the innovative design. The model was applied within two different elevation ranges where ROS is common to assess performance at locations with differing antecedent snow conditions. Model performance was assessed by comparing simulations to observations and a sensitivity analysis was undertaken to identify potential areas for model improvement. In chapter 6 selected model outputs were used to assess stand level energy
flux recovery during ROS events. Analysis focuses on recovery of mean, variability and frequency of snow melt energy flux during ROS among regenerating forests. A discussion on stand level hydrological recovery and the potential to scale results to the watershed will follow. Chapters 5 and 6 have a detailed introduction to provide context of the problem being addressed. Chapter 7 summarizes the major findings of the thesis, addresses study limitations and provides direction for future research.
Chapter 2: Background

This section provides an overview of interception processes for rain and snow. Following this, methods to monitor interception, accumulation and snowmelt are presented. A description of typical methods to model snowmelt and examples of application in rain-on-snow (ROS) environments will follow. An overview of commonly used methods to assess stand level hydrological recovery in forest management will also be reviewed. The final section discusses chronological and frequency pairing to detect change in forest hydrology.

2.1 Interception

Interception plays an important role in the water cycle of forested watersheds. Depending on the state of precipitation (rain vs. snow) and prevailing meteorological conditions, such as wind speed, relative humidity and net radiation, the fate of intercepted precipitation can vary markedly. The following sections describe how forest canopies influence rain and snow.

2.1.1 Rain

This section focuses on coniferous forests and temperate climates, and in particular on research completed in the Pacific Northwest (PNW) of North America. Forest canopies affect rainfall in two ways: reducing the intensity of precipitation reaching the soil profile and storing water in the canopy where it can be lost to evaporation. A review of interception loss in coniferous forests shows values ranging from 9 and 48% of total annual and seasonal precipitation, with most studies reporting at least 20%, varying with tree species, climate and stand structure (Ford and DeWalle, 1978; Hormann et al., 1996; Horton, 1919; Hudson, 2003; Pypker et al., 2005; Pypker et al., 2006b; Rutter et al., 1975; Spittlehouse, 1998). Individual events in an immature Douglas fir forest in western Washington had interception rates from 10.8 to 59%, with a mean of 31% (Pypker et al., 2005). Interception losses for individual events ranged from 6.1 to 100%, with a mean of 51% in an old growth Douglas fir forest in the western Cascades of Oregon (Pypker et al., 2006b). In another study, interception losses from rain events in an old growth stand in the western Cascades of Oregon ranged from 7.5 to
97%, with a mean of 45% per event (Link et al., 2004). In southwestern British Columbia, average interception rates of an old growth and a mature forest were 30%, while average interception by a series of immature stands ranged from 9 to 29% of total precipitation (Hudson, 2003). On an event basis, interception losses ranged from 12 to 99%, with a mean interception loss of 38% per event, while in the immature forests interception ranged between 3.6 to 71%, with a mean loss of 25% per event (Hudson, 2003). More recent research in an 120 year old redwood and Douglas fir forest in northwest California had interception rates by event between 2 and 100%, with interception decreasing as event size increased to a steady state of approximately 21% after 70mm of rainfall (Reid and Lewis, 2009). From the research reviewed, as rain events became larger, the proportion lost to interception decreased.

Rain follows several pathways through a forest stand, including direct throughfall, canopy drip, stem flow and rainfall splash. Direct throughfall is rain reaching the ground through openings within the stand. Canopy drip tends to occur once the canopy becomes saturated, although it can also initiate via preferential pathways prior to saturation (Pypker et al., 2006a). Stem flow can make up a significant component of throughfall in young stands, accounting for up to 27% of total precipitation reaching the ground (Ford and DeWalle, 1978). Stem flow depends on many factors, such as bark roughness, species, branch angle, stem density and tree height. Old growth and mature forests in the PNW tend to have low rates of stem flow (<3%) due to very rough bark and tall trees that provide a longer pathway to reach the ground and greater exposure to evaporation energy (Reid and Lewis, 2009; Rothacher, 1963; Spittlehouse, 1998). The final mechanism for rainfall to reach the ground is through rainfall splash when it first comes in contact with the canopy.

Precipitation stored in the forest canopy depends on age of the tree, structure, species and rainfall intensity. Pypker et al. (2005) found that a young forest canopy in south western Washington had less than half the canopy storage capacity of an old growth stand. Canopy storage averages between 3.0 to 5.0 mm in old growth coastal Douglas fir forests in Washington and Oregon (Link et al., 2004; Pypker et al., 2005; Pypker et al., 2006a). Much of this extra storage capacity in old growth forests is attributed to the epiphytic lichens found within the canopy (Pypker et al., 2006a; Pypker
et al., 2006b). In two separate laboratory experiments, stored water on needle-leaved branches reached a maximum at 1.1 mm (Keim et al., 2006), while epiphytes were found to increase storage capacity by approximately 1.3 mm (Pypker et al., 2006b). Furthermore, it has been shown that the storage capacity of forests reduces the rainfall intensity during precipitation events (Keim and Skaugset, 2003; Keim et al., 2006), which may reduce the landslide hazard (Keim et al., 2006; Pypker et al., 2006b). Keim et al. (2003) and Pypker et al. (2006b) both found the storage capacity of forests increased with increasing rainfall intensity, with Pypker et al. (2006b) attributing this to roughness of the lichen and bryophyte mats limiting splash and drainage. Keim et al. (2003) attributed reduction in intensity and increased storage to a loss of precipitation momentum when the rain came in contact with the canopy, suggesting that storage is dynamic rather than static as proposed in previously developed “bucket” models. As a result, there is general consensus that old growth forest canopies have a high capacity to store precipitation, much of which is attributed to arboreal epiphytic lichens. This may play an important role in hydrological recovery in harvested stands because young forests have significantly lower amounts of canopy lichen (McCune, 1993; Price and Hochachka, 2001). In fact, forests less than 70 years of age on the west coast of Vancouver Island had no lichen litterfall, compared to up to 680 kg/ha in an old growth stand (Price and Hochachka, 2001), although this does not necessarily mean there wasn’t lichen present in the canopy.

2.1.2. Snow and ROS

Compared to rain, the influence of the forest canopy on snow is generally more significant. The canopy not only interacts with snow when first deposited on the landscape, but also has a strong influence on the energy balance components that drive accumulation and melt processes (Berris and Harr, 1987; Ellis et al., 2010; Essery et al., 2003; Harr and Berris, 1983; Marks et al., 1998; Marks et al., 2001; Marks et al., 2002; Pomeroy and Dion, 1996; Pomeroy et al., 2009; Pomeroy et al., 1998b; Pomeroy and Schmidt, 1993). This section will first focus on how individual trees influence the amount of snow reaching the ground. A review will then be provided comparing snow interception in different forest types and their influence on energy balance components.
Canopy snow storage is approximately one order of magnitude larger than rain and can remain for several weeks (Lundberg and Halldin, 2001). Depending on species, maximum storage of snow per canopy unit area ranges from 3.0 and 5.5 mm snow water equivalent (SWE) (Schmidt and Gluns, 1991). Other studies use interception per ground area, with values in Japan coniferous forests being almost 50 mm (Lundberg and Halldin, 2001). Maximum storage in a Western Canadian boreal Jack pine and black spruce forests was 3.5 and 7 mm per ground area, respectively (Hedstrom and Pomeroy, 1998). In the PNW, depending on event size, snow storage ranged between 2.5 and 40 mm (Storck, 2000). Storage of snow within the canopy is highly variable and may depend on location, species and physical structure of the trees.

Research into the role of the forest canopy on snow accumulation and melt falls into two broad categories. The first includes potential increases in water yield, generally where increased yield is beneficial to human consumption (Golding and Swanson, 1978; Kittredge, 1953; Troendle and Reuss, 1997). The other category is potential impacts of forest cover removal on snow accumulation and melt and its relation to peak flow increases (Harr, 1981; Harr and Berris, 1983; Storck et al., 2002; Winkler et al., 2005). In general, peak snow accumulation in openings is 30-50% greater than mature forests (Kittredge 1953; Troendle and King, 1987; Troendle and Ruess, 1997; Storck et al., 2002; Winkler et al.; 2005). In continental climates, differences in snow accumulation between forests and openings is a result of losses due to sublimation and evaporation (Lundberg and Koivusalo, 2003; Schmidt et al., 1988; Stegman, 1996 ), provided factors such as aspect and slope are similar. In regions with maritime climates or transitional snowpacks, loss of intercepted snow due to evaporation or sublimation is relatively low because conditions in these environments encourage rapid release of snow from the canopy through meltwater drip, mass release or rain wash (Satterlund and Haup, 1970; Storck et al., 2002). Under certain conditions snow melts faster from the canopy than on the ground (Beaudry and Golding, 1983; Berris and Harr, 1987), resulting in higher water input to the soils below a forest and less storage as snow. A study in the coast mountains of British Columbia observed events where old growth forests had greater inputs of water to the soil than a clear-cut and it was hypothesized that this occurred when snow was in the canopy at the onset of an event (Beaudry and Golding, 1983). While this may not
result in lower water yields over the course of a season, it can reduce the amount of water stored in the snowpack available for melt during a ROS event (Coffin and Harr, 1992). This implies that while differences between snow accumulation in forested areas and clear-cut areas in maritime and continental climates may be similar, the physical processes resulting in those differences are not the same. This may have implications when assessing the potential impact of forest harvesting on water yield and peak flow.

During the spring freshet snowmelt is the primary source of water and is driven primarily by net radiation (U.S.A.C.E, 1956). The spring freshet is the primary producer of peak flows in continental climates dominated by snowmelt, however, in watersheds located in maritime climates where ROS events drive peak flow generation, turbulent energy transfer of latent and sensible heat are the primary energy sources (Beaudry and Golding, 1983; Berris and Harr, 1987; Marks et al., 1998; Mazurkiewicz, 2006; van Heeswijk et al., 1996). Because of this, clear-cuts have significantly more snow accumulation than forests and snowmelt rates can be 30 to 150% greater in openings (Berris and Harr, 1987; Storck et al., 2002; Winkler et al., 2005). This increase in melt occurs due to increased energy flux in clear-cut areas from solar radiation during spring snowmelt (e.g. Winkler et al., 2005) and enhanced wind speeds resulting in high rates of turbulent exchange during ROS (Marks et al., 1998; Marks et al., 2001). While solar radiation may play a reduced role during ROS events, it can be very important in determining the antecedent snow conditions leading to the event (Mazurkiewicz et al., 2008). For example, a south facing slope in the ROS zone of a watershed will generally have less of a snowpack than a north facing slope due to higher solar inputs. Thus placement of cutblocks within a watershed can be very important in determining the antecedent snow conditions prior to either ROS or spring snowmelt events and affect the synchronization of processes governing peak flows (Moore and Wondzell, 2005).

In summary, the forest canopy plays a strong role in both the accumulation and melt of snow. In maritime watersheds, melt water drip can be a major factor in reduced snow accumulation under forests. Forest openings tend to have snowpacks with higher SWE and are exposed to greater energy fluxes than forests, resulting in significantly higher rates of water output during both radiation and ROS melt events. Because of this, extensive forest cover removal may result in significant increases to streamflow.
However, watershed processes are extremely complicated and as discussed in Chapter 1, linking forest practices directly to increased peak flows is a contentious issue. Hydrological modelling provides a means to link forest practices to changes in flow, yet, uncertainty remains in data used to run models, model structure and observations to validate/calibrate outputs.

2.1.3. Measuring snow interception

A number of factors make it difficult to measure canopy interception and/or snowmelt directly during ROS. The events tend to be relatively rapid, with trees often losing intercepted snow within hours of the initial snowfall and shallow snowpacks melting within days. As a result, traditional methods of measuring snowpack development, such as weekly, or bi-weekly manual snow surveys (e.g. Hudson, 2000a; Lundberg and Koivusalo, 2003; Stottlemeyer and Troendle, 2001; Talbot and Plamondon, 2002; Winkler et al., 2005) can miss such events. Climate station networks and snowmelt lysimeters can be used to describe how forests influence snow interception and melt processes, but they provide limited information as to state of precipitation in the canopy (rain vs. snow), how long it stays in the canopy, how much of the precipitation reaches the ground as snow vs. rain, and what processes cause snow to be lost from the canopy (unloading, meltwater drip and/or sublimation). Some of these data could be collected using cut tree methods (Hedstrom and Pomeroy, 1998; Lundberg et al., 1998; Storck, 2000; Storck et al., 2002) or by measuring branch deflection through strain gauges (Huang et al., 2005) or video cameras (Brundle et al., 1999). Limitations to these approaches include high costs both in time and resources, while focusing on individual trees and in some cases individual branches making it very difficult to scale results to the stand or watershed level. Additionally, all cut tree experiments are limited to those that can be operationally secured to a weighing device, eliminating most trees found in PNW old growth forests. There is potential to directly measure interception using a non-destructive technique where mechanical displacement sensors are placed along a trunk to measure compression as a function of water storage, however it has not been applied extensively (Friesen et al., 2008).
The most comprehensive research to date that directly measured canopy snow interception in a watershed frequented by ROS was done in the Oregon Cascades (Storck, 2000; Storck et al., 2002). Individual cut trees from regenerating forests were weighed using load cells to determine maximum snow loads and monitor general interception dynamics. Direct measurement of interception was limited to trees that could be secured to the load cells, so there were no measurements and observations of mature trees. These data were supplemented using snow courses, weighing snowmelt lysimeters and meteorological stations to describe the fate of intercepted snow within coastal Douglas fir forests. Within British Columbia there has been limited research completed on ROS processes and the role of interception, with Beaudry and Golding (1983) being the only researchers who directly compared lysimeter and snowmelt energy during ROS in a clear-cut and old growth stand. However, they did not attempt to monitor nor model canopy interception directly. Furthermore, they only focused on a clear-cut and old growth stand, providing no information pertaining to regenerating forests.

As described in the introduction there is potential to use automated time-lapse photography to monitor snow accumulation and ablation (Lundberg and Halldin, 2001). Automated time-lapse photography has been used in the physical sciences to monitor changes in channel characteristics (e.g. Dexter and Cluer, 1999; Hancock and Willgoose, 2001; Lawler, 2005; Sheets et al., 2002), cloud cover, (Dexter and Cluer, 1999; Holle et al., 1979) and snow cover on glaciers (Haefner and Laager, 1988), in arctic foothills (Evans et al., 1989) and in urban catchments (Matheussen and Thorolfsson, 2003). Within the PNW, time-lapse photography has been used to monitor snow depth and state of precipitation (Berris and Harr, 1987), however, the images were not used for quantitative analysis. Increasingly, digital and video cameras are being used in snow hydrology to measure canopy interception (Brundle et al., 1999; Pomeroy et al., 1998b; Pomeroy and Schmidt, 1993) and even changes in snow albedo (Melloh et al., 2001). These cameras allow for hundreds of pictures to be acquired and stored electronically, and as a result digital photogrammetric data collection is becoming more frequent. Remote sensing now offers many tools previously unavailable to utilize images as data sources and there is potential to use this technology in ROS environments to monitor rapidly changing snow conditions in the canopy and on the ground.
2.1.4 Snow interception models

Snow initially interacts with the forest canopy prior to deposition on the ground, which is an inherently complex process to describe, monitor and model. Many different snow interception models are available covering a range of scale, from individual trees or branches (e.g. Hedstrom and Pomeroy, 1998; Satterlund and Haupt, 1967), to the stand level (e.g. Essery et al., 2003; Hellström, 2000). Some models focus specifically on the amount of snow trees or branches can physically hold (e.g. Satterlund and Haupt, 1970) while others attempt to measure sublimation/evaporation directly and/or through energy balance equations (Andreadis et al., 2009; Ellis et al., 2010; Hedstrom and Pomeroy, 1998; Lundberg et al., 1998). Most research to date has focused on continental forests in watersheds where the bulk of water lost from these systems during winter months is through sublimation and/or evaporation of intercepted snow. Limited research on interception processes in maritime regions has occurred, due in part to the difficulty associated with collecting data in such climates for developing and testing models. Furthermore, while the absolute amount of evaporation/sublimation in the maritime watersheds may be large compared to continental climates (100 mm/year), the relative amount is only around 10% (Storck et al., 2002), thus it is of less concern in terms of annual water yield in these water “rich” areas. In maritime watersheds, differences in snow depth between forests and openings are not due to sublimation, but rather because the majority of snow in the canopy melts before it is released to the ground (Berris and Harr, 1987; Storck et al., 2002). Even though water is not lost from the system, melt from the canopy results in significantly lower snowpacks under forests, decreasing potential water available during floods and thus it is important to have the ability to both monitor and model snow interception within coastal watersheds.

The forest canopy not only influences the amount of snow that initially reaches the ground, it also affects the transfer of energy to the snowpack. Therefore, interception models must not only include routines to store, melt, sublimate, evaporate and release snow from the canopy, they must also simulate the different energy sources available to melt the snowpack under the canopy. Recently interception, models have been developed
for incorporation into existing snowmelt models: the Cold Regions Hydrological Model (CRHM) (Pomeroy et al., 1998) and the Variable Infiltration Capacity (VIC) model (Liang et al., 1994). The Canopy interception module (Ellis et al., 2009) for CRHM is quite robust and will be described in detail in Chapter 5. It is based primarily on work undertaken in the Alberta Rockies and estimates the energy fluxes in forested environments (Ellis et al., 2009). While it has been tested in a number of environments (Ellis et al., 2009; Essery et al., 2009; Rutter et al., 2009), it has yet to be used within either regenerating forests or in coastal watersheds frequented by ROS. Within the CRHM framework, the Canopy model can be combined with SNOBAL, a surface energy model that has been applied extensively in ROS environments (e.g. Marks et al., 1998; Mazurkiewicz et al., 2008). The interception sub-routine within the VIC model (Andreadis et al., 2009) was developed using research from Oregon (Storck, 2000; Storck et al., 2002), however, the VIC model is structured for meso-scale watersheds making it less appropriate for third order watersheds used in forest management. Both interception models are comparable in how they function and produced similar results when they were validated against depth and SWE data (Andreadis et al., 2009; Ellis et al., 2010).

2.2 Snowmelt models

Snow that reaches the ground is subject to a number of energy sources that contribute to melt, including longwave and shortwave radiation, sensible and latent heat, advective heat from rainfall and ground heat flux (Male and Granger, 1981). Snowmelt models simulate energy inputs through climate variables such as air temperature, solar radiation, wind speed, soil temperature and relative humidity. Snowmelt models fall within two broad categories: empirical and physical. For reasons of simplicity and because this research project will focus on plot and stand level measurements, only point scale models are reviewed.

2.2.1 Empirical snowmelt models

Operationally, empirical snowmelt models tend to be more common, with the temperature index model widely used for both point and watershed modelling (e.g. Hock, 2003; Micovic and Quick, 1999; Ohmura, 2001; U.S.A.C.E, 1956). The temperature index model has minimal data requirements, primarily precipitation and temperature.
Temperature has a strong relationship to energy balance components, especially longwave radiation, which is often the largest component of melt models under forest canopies and cloudy skies (Ohmura, 2001). In some instances, particularly at the catchment scale, temperature index models can outperform full energy balance models, although accuracy tends to decrease with increasing temporal resolution and when accounting for topographic controls (Hock, 2003). These limitations have resulted in development of “hybrid” temperature index models incorporating solar radiation or other available meteorological measurements (e.g. Brubaker et al., 1996; Hock, 2003; Kustas et al., 1994). A temperature index model in isolation may be unsuitable for maritime climates during ROS events due to high amounts of turbulent transfer and high humidity (Hock, 2003). Early work completed by the U.S.A.C.E. (1956) in maritime watersheds recommended using temperature and rainfall in forested stands, while adding a wind speed coefficient in clear-cut areas when using a temperature index approach. If ROS events are of primary concern in maritime climates, it may prove useful to develop hybrid temperature index models for various forest types for use in full hydrological models. The low cost of instruments that can measure temperature, RH and even wind speed make the use of hybrid approaches increasingly feasible. However, in well instrumented watersheds, energy balance models can provide much more detailed information related to forests and their effects on both snow accumulation and melt.

2.2.2 Energy balance models

Most of the early work on snow energy balance was completed by the U.S.A.C.E. (1956), and the same basic equations are used today. In theory, the major advantage of energy balance models is they can be applied to any watershed or location, without having to create local parameterization such as degree day factors when using the temperature index approach. In reality, most physically-based models are not readily transferable between watersheds due to a poor understanding of spatial variability for model inputs, use of empirical functions for certain components and a lack of suitable datasets to run the models (Essery et al., 2009; Rutter et al., 2009). To overcome this uncertainty, models are often calibrated by altering the input parameters to produce results that best match observed values (Kirchner, 2006), often stream discharge, snow
water equivalent from snow course measurements or snow pillows, or snow depth measurements from sonic snow depth sensors. For this reason, most energy balance models are used in heavily instrumented watersheds to provide insight into the physical processes that drive watershed processes (e.g. Andreadis et al., 2009; Boon, 2007; Ellis et al., 2010; Kuraś et al., 2011; MacDonald et al., 2010; Marks and Dozier, 1992; Marks et al., 1998; Marks et al., 2001; Marks et al., 2002; Mazurkiewicz et al., 2008; Pomeroy et al., 2007; Pomeroy et al., 2009; Pomeroy et al., 1998b; Rutter et al., 2009; Storck, 1998; Thyer et al., 2004; van Heesjwick et al., 1996; Whitaker et al., 2003). The majority of models applied in maritime climates that attempt to differentiate between melt rates in forests and openings usually do not have observations such as SWE or snow depth below a forest to validate results (e.g. Marks et al., 1998; Mazurkiewicz, 2006), making it tenuous to use model output from forested stands. Even in areas with detailed observations below forested stands, model performance varies depending on location and dominant processes that drive snow accumulation and melt (Essery et al., 2009; Rutter et al., 2009), which can partly be attributed to the use of empirical equations developed elsewhere to describe processes such as albedo decay (e.g. Essery and Etchevers, 2004; U.S.A.C.E, 1956).

A number of factors affect energy exchange processes including latitude, season, slope, aspect and forest cover, along with conditions at the snow surface including albedo and roughness (Male and Granger, 1981). The U.S.A.C.E have published a series of reports on snowmelt algorithms and energy balance equations that form the basis of many point snowmelt models (e.g. Melloh, 1999; U.S.A.C.E, 1956). In addition, Dingman, (2002) and Male and Granger (1981) present excellent reviews of the snow accumulation and ablation energy balance. The application of energy balance models to snowmelt is extensive, however, most work done in North America has focused on continental environments (e.g. DeBeer and Pomeroy, 2009; Ellis et al., 2010; Essery et al., 2009; Kuraś et al., 2012; Marks et al., 2001; Marks et al., 2002; Parviainen and Pomeroy, 2000; Pomeroy and Dion, 1996; Pomeroy et al., 1998a; Pomeroy et al., 2009; Pomeroy et al., 1998b; Pomeroy et al., 2003; Rutter et al., 2009; Whitaker et al., 2003) with limited work completed in forested maritime watersheds dominated by ROS processes (e.g. Andreadis
et al., 2009; Marks et al., 1998; Storck, 1998; Storck, 2000; van Heesjwick et al., 1996). Within the literature reviewed for ROS environments, there were very few examples where energy balance models were applied to regenerating stands (e.g. Storck, 2000), rather most models have been applied exclusively in clear-cut and/or mature/old growth forests. As a result, there is little information related to both model performance and general snowmelt energy dynamics in regenerating forests. This type of information will be critical to assessing both stand level and watershed scale hydrological recovery following disturbance.

2.3 Stand level hydrological recovery models

Within British Columbia one of the primary indicators used for watershed management is Equivalent Clearcut Area (ECA), as described in Chapter 1. Its application is widespread due to its limited data requirements and lack of other suitable tools. Research has been completed in the interior of the province to improve the ECA methodology and test its appropriateness (Winkler et al., 2005). The only other research found on interception recovery using different aged forests was completed in continental forests in eastern Canada and Montana (Buttle et al., 2005; Hardy and Hansen-Bristow, 1990; Talbot and Plamondon, 2002; Talbot et al., 2006) and thus has limited application in maritime watersheds. In the interior of British Columbia peak SWE was 32% and 14% lower in a mature and regenerating stand compared to a clear-cut (Winkler et al., 2005), however solar radiation was the dominant form of melt energy, thus results may not be applicable to watersheds driven by ROS processes.

A case study in coastal BC (Hudson 2000a) found that snowmelt in an 8 meter tall coniferous forest was 75% recovered to pre-harvest conditions (old growth forest), with trees 20 meters in height and 95% canopy closure approaching full recovery. Hudson and Horel (2007) combined rainfall data and snowmelt data from two different regions to develop ROS recovery curves to calculate ECA. Hudson (2000a) and Hudson and Horel (2007) used a minimum canopy closure along with tree height to model recovery, however, others have found measurements of crown closure and crown length are better predictors (Buttle et al., 2005; Winkler et al., 2001). There has been no research on stand level snowmelt recovery using data collected specific to ROS events in British Columbia,
and as a result existing recovery models were developed using data from weekly or bi-weekly snow courses (e.g. Hudson and Horel, 2007). Because of this, it is uncertain if the current ECA methods are appropriate to use in watersheds where peak flows are generated by high intensity ROS events. In addition, all ECA methods developed within British Columbia are based on research that used chronological pairing (see section 2.4) to compare the mean response of melt rate among different forests. As a result, they do not account for differences in variability or frequency of melt events. With the limited research available to develop methods to calculate ECA and its wide application, there is a need to collect more data specific to melt rates during ROS across a range of forest types and elevations in coastal watersheds to test and improve existing stand level recovery models.

2.4 Chronological pairing vs. frequency pairing

Most paired watershed research compare streamflows by chronologically pairing (CP) events that occur within the same time frame between a control and treatment watershed (e.g. Beschta et al., 2000; Christner and Harr, 1982; Harr, 1986; Harr et al., 1982; Harr and McCorison, 1979; Harr et al., 1975; Jones and Grant, 1996; Jones and Perkins, 2010; MacDonald and Stednick, 2003; Moore and Wondzell, 2005; Thomas and Megahan, 1998; Troendle and Stednick, 1999; Troendle et al., 2001). Following this either ANOVA or ANCOVA analysis is used to determine if significant differences exist in the mean response of streamflow in the treatment vs. control watershed. One of the difficulties in deciding which events to pair arises from the variability in timing of flows due to watershed characteristics and the processes that generate them (rain, ROS or snowmelt) (e.g. Jones and Grant, 1996). Because of this, it can sometimes be difficult to separate whether flow differences are due to land management or because of differing meteorological inputs. Despite these difficulties, CP remains the dominant method to compare streamflows and water delivery to the soil profile in disturbed vs. harvested areas.

A key assumption of ANOVA or ANCOVA analysis is that both populations have equal variability, and if they do not, the data are transformed. While this may be valid if only interested in comparing the differences in mean response, it is often the events
smaller or larger than the mean that are of greatest concern, such as flooding or drought. In a probability density function (PDF) or cumulative distribution function (CDF), these are the values found at the tails. A change in the mean and variability (e.g., variance, standard deviation, etc) of a population will result in a change to the frequency of occurrence (Hosking and Wallis, 1997), which in forest hydrology would result in changes to the return interval of certain events. If either mean or variability increases, the CDF will shift, and events will become more frequent (Figure 2.1.). Because of this relationship, using ANOVA or ANCOVA to compare watersheds with different variability in streamflows can actually suppress the attribute that is of greatest importance (Alila et al., 2009). In fact, by transforming the data to reduce variability, it is possible that ANOVA or ANCOVA may indicate that the means are equal, when in fact the frequency distribution has changed. The importance of changes to variability has long been recognized in climate change research and methods have been designed that focus on detecting changes to frequency distributions rather than solely the mean (e.g., Katz, 1993; Katz and Brown, 1992).

To account for changes in the frequency of streamflows between treatment and control watersheds, recent research has used frequency pairing (FP) (Alila et al., 2009; Green and Alila, 2012; Kuraś et al., 2012). Rather than comparing streamflows chronologically, this method pairs events by frequency and examines the entire flood frequency distribution. Alila et al. (2009) argued that CP analysis does not account for changes to frequency and that it is a fundamentally wrong approach to detect changes in streamflow because it de-couples frequency from magnitude and prevents any assessment of the return interval of streamflow events. This assertion resulted in a rebuttal defending the use of CP and questioning some of the results obtained in Alila et al. (2009), focusing mainly on the reported changes to extreme flows which were much greater than previously published (Lewis et al., 2010). The reply (Alila et al. 2010) acknowledged the possible uncertainties surrounding their estimates in changes to magnitude and frequency of larger flows, but they maintained their claim that CP methods are inappropriate, and mask the true effects of forest harvesting on frequency of streamflows, especially for large or extreme events. Since then a number of studies have been published using CP (e.g., Bathurst, 2011; Jones and Perkins, 2010) suggesting that forest cover removal does
not impact large or extreme flows. There has been one additional published study that uses FP, which concluded that as events become larger in magnitude, the effects on return interval increase, with extreme events being very sensitive to slight increases in magnitude (Kuraś et al., 2012). Another paper that is currently under review using FP has come to similar conclusions (Green and Alila, 2012). All of the studies reviewed in the previous sections that compared differences in snowmelt/rainfall between harvested and mature forests or assessed stand level hydrological recovery, used CP to compare events and only focused on differences in mean. This suggests that there is a need to apply FP analysis to new and existing forest hydrology datasets.

![Figure 2.1](image)

Figure 2.1. The effect of changing the mean or variability on the cumulative frequency of a normal distribution. The red, blue and green lines have equal means and different variability, while the blue and black lines have equal variability and different means.

### 2.5 Needs and opportunities

The majority of stand level research has focused on differences in rainfall and snowmelt between recently harvested and old growth or mature stands, with limited focus on regenerating forests. Numerous snow interception and snowmelt models are available to assess differences in energy available for melt during ROS, however, they have not been extensively applied in maritime watersheds. Testing the models will require data to
run and validate simulations. To accomplish this, existing climate station networks in research watersheds can be used, though they must be supplemented with data collected at the stand level, especially in regenerating forests. Key processes to monitor will include snow accumulation and melt, and interaction of snow with the forest canopy to validate simulations of these processes. This will require the development of methods which can be applied extensively and will collect high temporal resolution data to capture interception and snowmelt dynamics during ROS. Modelled results can then be used to generate datasets to compare differences in snowmelt during ROS among different phases of forest development with a focus on mean, variability and frequency to improve methods used to assess stand level hydrological recovery in coastal watersheds.
Chapter 3: Common study design and methods

3.1 Site description

Russell Creek Experimental Watershed is located in southwestern British Columbia on Vancouver Island at 50° 19’ N and 126° 21’ W (Figure 3.1). It is approximately 31 km² with elevations ranging between 300 and 1680 m (a.s.l.). The transitional snow zone receives multiple rain-on-snow (ROS) events between November and April and occurs primarily between 300 and 900 m a.s.l., with persistent snowpacks generally occurring above 900 meters a.s.l. It is not uncommon for ROS events to occur over the entire elevation range, usually in late fall and early winter. Average winter (December through February) temperatures are -0.1°C at the highest elevation (900m a.s.l.) in the transitional snow zone. Temperatures generally do not fall below -10°C and when they do it is only for days or a few weeks. Snowpacks within the transitional snow zone completely melt and reform many times throughout the fall and winter season. The watershed receives mean annual precipitation of approximately 2900 mm, of which 70% typically falls between October and March. Major tree species include old growth and regenerating stands of western hemlock (*Tsuga heterophylla*), western red cedar (*Thuja plicata*), amabilis fir (*Abies amabilis*) and regenerating stands of Douglas fir (*Pseudotsuga menziesii ssp. menziesii*).

3.1.1 Study design

Two long-term climate stations (1995-present) are located in Russell Creek Experimental Watershed at 490m and 840m a.s.l. These climate stations measure air temperature (+/- 0.5°C) and relative humidity (RH) (+/-5%) (Unidata Corporation Model 6501), total precipitation (liquid and solid) (400mm opening PVC, salt solution displacement gauge), wind speed (RM Young Model 6533) and incoming solar radiation (Apogee SP-110, accuracy +/- 5%) at 3m above ground. Precipitation under-catch was corrected based on wind speed and precipitation type (rain, mixed, snow) (Yang et al, 1998). As part of this project, two subzones within the transient snow zone were delineated based on elevation to gain an understanding of snow accumulation and melt among forest types with different antecedent conditions. A total of 8 plots were installed within two general elevation bands of the transient snow zone, Lower Russell (LR)
between 520 and 560 m a.s.l. and Upper Stephanie (US) between 740 and 770 m a.s.l. (Table 3.1). The LR plots included regenerating stands with mean heights of 4.5 m (LR4m) and 13 m (LR13m), along with a clear-cut (LRCC and old growth (LROG) stand. The US plots included a regenerating stand 4.1 m (US4m) and 11 m (US11m) tall, along with a clear-cut (USCC) and old growth (USOG). The US11m plot was located outside of the watershed because no forests of that height were in Russell Creek at that elevation. There were no regenerating stands with mean heights taller than 11 or 13 m within the elevation ranges at Russell Creek that we were interested in. Figure 3.2 provides examples of forests compared in the study.

Air temperature was measured 3 m above ground at each plot (N=8) using Hobo Pendant Data Loggers (Model # UA-003-064; Onset Computer Corporation, Bourne, MA, USA) with an accuracy of 0.47°C and resolution of 0.1°C, with relative humidity measured at each location using iButton Hygrochoms with a resolution of 0.6% (model # DS1923-F5; Maxim Integrated Products, Inc, Sunnyvale, California 2006). Soil temperature was measured 10 cm below ground using iButton (model # DS2422) temperature sensors with an accuracy of 0.5°C and resolution of 0.0625°C (Maxim Integrated Products, Inc, Sunnyvale, California 2006). Snow temperature data were collected at fixed depths at the plot locations using iButtons as described above. Snow temperature measurements were only available in 2006-2007. All data were recorded on an hourly basis.

Snow courses were completed every 2 to 4 weeks at each plot. Snow courses were laid out based on a sampling design similar to that of Jost et al. (2007) in cardinal directions on 60 m transects with 1 measurement every 10 m for depth and density and 1 depth measurement every 2 m. Snow water equivalent (SWE) measurements were completed using a standard federal snow sampler (diameter 4.13 cm). Snow depth data from the snow courses were used to compare the variability of snow depth within the research plots with the depths measured using the time-lapse cameras described in the next chapter.

At each research plot center, a 12-meter fixed radius plot was used to determine stem density, species composition and average stand height of dominant and co-dominant
trees. At the plot center and at each snow density measurement point, hemispherical photography using an 180° fisheye lens (CoolPix 5400-FC E9) on a self leveling upward facing mono-pod was used to capture sky-view images through the forest canopy. Images were processed using HemiView Canopy Analysis Software (Delta T Devices, Version 2.1 1999, Burwell, Cambridge, UK) to calculate Leaf Area Index (LAI) and Visible Sky, a value inversely proportional to Canopy Fraction. Table 3.1 provides details for each plot location. Figure 3.3 provides examples of sky view images from the forested locations.

Non-weighing snowmelt lysimeters were installed at each plot. The snowmelt lysimeters were 7.5 m x 0.1 m and drained into a 350 ml tipping bucket attached to a Hobo Pendant time stamped event recorder (Model # UA-003-064; Onset Computer Corporation, Bourne, MA, USA). This length was chosen based on research completed (Keim et al., 2005) in western Oregon indicating that throughfall in forests was best captured using instruments 3 to 10 meters in length. The lysimeter troughs were buried to ground level and stuffed with western hemlock (*Tsuga heterophylla*) branches collected locally to prevent ice from building up in the trough. The tipping buckets were enclosed within a plywood box and buried in the ground to prevent freezing. To prevent ponding of water that can lead to leaks and blockages, a standard 110 mm PVC 90° elbow was connected directly to the trough. The elbow drained directly into a large screened funnel that emptied into the tipping bucket. The lysimeters were designed for relatively shallow snowpacks.

The remaining chapters will present additional methods specific to their objectives. Some information may be repeated from this chapter purely for convenience of the reader and to provide a more thorough understanding of how the overall study design works with the individual chapter.
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<th>Plot Name</th>
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<th>Average Tree Height (m)</th>
<th>Leading Species*</th>
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<th>LAI</th>
<th>Canopy Fraction</th>
<th>Latitude</th>
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*Fd – *Pseudotsuga menziesii* (Douglas fir); Hw – *Tsuga heterophylla* (Western hemlock); Cw – *Thuja plicata* (Western red cedar); Ba – *Abies amabilis* (Amabilis fir)
Figure 3.1. Location of Russell Creek Experimental Watershed (Data SIO, NOAA, U.S. Navy, NGA, GEBCO; © 2012 Cnes Spot Image; Image © 2012 TerraMetrics; ©Google Earth, 2012).
Figure 3.2. Examples of clear-cut (CC), 4m, 13m and old growth (OG) forests in Russell Creek Experimental Watershed where plots were located. The 11m forest had similar structure to the 13m forests.
Figure 3.3. Upward looking view from the forests in Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740-770m a.s.l.) taken with a fish eye lens camera.
Chapter 4: Innovative methods to monitor snow accumulation and melt during rain-on-snow

4.1 Introduction

A number of factors make it difficult to measure canopy interception directly during ROS. The events tend to be relatively rapid, with trees losing intercepted snow often within hours of the initial snowfall and shallow terrestrial snowpacks melting within hours to days. Due to this, traditional manual methods of measuring snowpack development, such as weekly or bi-weekly snow surveys (e.g. Hudson, 2000b; Lundberg and Koivusalo, 2003; Stottlemyer and Troendle, 2001; Talbot and Plamondon, 2002; Winkler et al., 2005) can completely miss such events. While rapid changes in snow depth can be measured using a sonic snow depth sensor, the measurement footprint is relatively small and cannot capture the variability of depth underneath forest canopies. Through the use of a climate station network and snow lysimeters, precipitation inputs to the watershed can be measured, however, minimal information is collected as to the state of precipitation in the canopy (rain vs. snow), length of time it stays in the canopy, whether precipitation falls as snow vs. rain, and the processes causing snow to be lost from the canopy (i.e. mass release, meltwater drip and/or sublimation). Some of these data could be collected by using cut tree methods (Hedstrom and Pomeroy, 1998; Lundberg et al., 1998; Storck et al., 2002) or by measuring branch deflection through strain gauges (Huang et al., 2005) or video cameras (Brundle et al., 1999). However, these approaches are costly both in time and resources, and focus on individual trees and in some cases individual branches, making it difficult to scale results to the stand or watershed level. Additionally, all cut tree experiments are limited to those that can be secured to a weighing device, eliminating most trees found in old growth forests. Due to these limitations additional methods are required to monitor state of precipitation, both terrestrially and arboreally, relative differences in snowfall interception between forest types, duration snow stays in the canopy and processes that cause canopy snow removal.

An innovative system was designed to monitor canopy interception and throughfall during ROS events using a combination of climate stations, non-weighing snowmelt lysimeters, suspended spring scales and a remote camera network. It is the
objective of this chapter to present this innovative experimental design. We will show how the use of digital time-lapse photography in conjunction with a meteorological network can be used to generate both qualitative and quantitative data that cannot otherwise be measured without direct personal observations in areas with rapidly changing climatic conditions.

4.2 Methods

4.2.1 Study design

At each study plot described in Chapter 3, a time-lapse camera was installed to monitor snow in the canopy and on the ground. At sites where a single camera could not capture both the snow below the canopy and within the canopy, a camera was placed under the stand and one was placed at a location outside the stand. A series of snow stakes were installed to gauge snow depth at all camera locations (Figure 4.1b). In addition, each camera location contained a scale elevated above the ground to measure mass of snow throughfall using a 50 cm x 100 cm platform connected to two springs (Figure 4.1c). Known weights were used to calculate the deflection of the springs. Springs were manufactured specifically for this project using stainless steel 302. Each spring was 30 cm long prior to load, extending to 80 cm under a full load of 18 kg per spring. With two springs attached per scale, a 1 cm change in SWE resulted in a deflection of 7.9 cm. A cable was secured to each spring to ensure it did not extend past its maximum design deflection. The weight and length of deflection were selected based on knowledge of the typical range in snowpack conditions experienced at the site. Depending on application and required precision, scale design will vary. Because temperature fluctuations were minimal at the sites compared to the range of temperatures stainless 302 is designed to experience, changes in the modulus of elasticity were assumed to be negligible.
Figure 4.1. Masks used to analyze time-lapse images, (a). Polygon masks (red) used to isolate the canopy to calculate snow fraction as an index for interception. (b). Linear masks used to calculate changes in snow depth and (c) throughfall SWE on spring scale. Note that smaller vertical lines between the stakes (b) were not used in our analysis because the individual pixels were not narrow enough to isolate the line from the snow.
4.2.2 Camera design

Canon PowerShot A430® cameras were chosen for use in the remote camera design. They were selected due to low cost, a time stamp feature and a lens which retracted and extended during power up. Images were taken on an hourly basis and stored on a 512 MB SD card allowing for a deployment of 42 days. The camera was turned on and off using a 12 volt repeat cycle time delay relay (Macromatic TR-66126), with a timing range between 0.6 seconds and 24 hours for both the off time and the on time (Figure 4.2a and 4.2b.). Wires were soldered to the power switch on the camera and a time delay relay was installed to turn the camera on and off (Figure 4.2c). To avoid having to use another time delay relay to take the photo, a magnetic contact switch was soldered to the shutter release mechanism. A magnet was then attached to the end of the camera lens (Figure 4.2d). When the camera was powered up, the lens retracted, then extended. Upon extension the magnet at the end of the lens closed the magnetic contact switch attached to the shutter and a photo was taken. The wiring of the camera allows for a data logger to be used for event based photos if required. This design used a minimal amount of power because the camera was only turned on for 10 seconds for every photo taken. Typically photos could be taken for periods of six to eight weeks on an hourly basis using a 12 volt 12 Ah battery. Battery life depended on temperature and the number of times the flash is used. A voltage regulator with an input range of 4.3 to 16 volts (NTE 1904) was installed to provide the 3.3V for the camera. Cameras were housed within a commercially available waterproof Lexan® case (GSI 73510). Desiccant was placed within the case to prevent condensation. Previous studies using cameras to measure interception were generally limited to one or two cameras due to high costs and could take relatively few photos due to memory limitations (Brundle et al., 1999; Pomeroy et al., 1998b; Pomeroy and Schmidt, 1993). The major advantage of this camera design is the ability to take high resolution images at a high frequency over a relatively large area for minimal cost, typically at ¼ the cost of available commercial versions. Furthermore, by using available digital image processing software, analysis of the photos was done automatically and provided previously unavailable quantitative and qualitative information.
Figure 4.2. Time-lapse camera design information: (a) rear view of camera showing time delay relay mechanism and voltage regulator; (b) top view of time delay relay; (c) shutter and power wiring on internal circuit board of camera; (d) front view of camera showing contact switch and mount with the magnet on the end of the retractable lens.

4.2.3 Image analysis

Photogrammetric image analysis was used to calculate presence of snow in the canopy, accumulation/ablation on the ground and throughfall SWE. We used ENVI and IDL image processing software for analysis (ITT Visual Information Solutions, Boulder, Co, 2007). A selection of photos was used from each station to identify the threshold value of snow vs. non-snow in the RGB (red-green-blue) color space. We arbitrarily selected the green color space to determine snow vs. not snow due to the similarity between all three color spaces when comparing snow to no snow pixels. Thresholds were determined using the “Arbitrary Profile Transect” tool within ENVI to collect digital number values between 1 and 256 for snow covered and non-snow covered areas within the image. A selection of 20 transects were created for each light condition to determine the most appropriate threshold value. The “Arbitrary Profile Transect” tool allows the
user to select areas of the image to create a profile of colour distribution. Figure 4.3 provides an example of the output generated by the tool when a transect is created over a snow covered area and a snow free area. Gray scales were painted on 30cm by 45cm plywood boards and attached to a tree or stake so they could be used to adjust for changes in illumination during photogrammetric analysis. As images became darker, in the early morning and early evening, thresholds were more difficult to identify. Once a threshold value was established, polygon masks were created in the images to isolate trees for analysis of canopy interception by calculating percent white (Figure 4.1a). The number of pixels within each analysis mask that were snow and not snow was calculated by creating a batch analysis routine with IDL to load and analyze each photo and output results in a text file. From this, canopy snow cover fraction was calculated ranging between zero and one, with zero indicating no snow in the canopy and one indicating 100% snow cover. It is important to note that snow cover fraction is only an indicator of total snow interception and is based on camera position and angle. It is unlikely any image would have a canopy snow cover fraction of one as this would require every pixel in the analysis mask to be white. Images taken at night could not be used for analysis, thus gaps in data were typical when it was dark. Figure 4.4 provides an example of a photo sequence used for analysis.
Figure 4.3. Example of output from ENVI’s “Arbitrary Profile Transect” tool for a transect over snow and non-snow. The difference between snow and non-snow changes depending on illumination of the image.

Because the length of the stakes was known, a linear mask was created over the stake (Figure 4.1b) and the length of each pixel was calculated within the mask. Threshold values on the stake for snow vs. not snow were determined. Using these threshold values, the number of snow pixels was calculated and then multiplied by the pixel length to calculate snow depth. A similar approach was used to calculate deflection of the spring scales (Figure 4.1c). Each spring was covered with a linear mask. Two distinct colors (red) were placed on each spring and were tracked on the linear mask. As the colors moved apart, the distance was measured, and mass of the snow throughfall was calculated. Changes to the spring scale at night were also measured because the camera flash interacted with reflectors on the springs and could be identified clearly in the image. IDL was used to create batch analysis routines for snow stake and spring scale analysis.

The photogrammetric analysis data were plotted against climate data to identify anomalies associated with camera movement, snow rime build up on snow stakes or
condensation/frost on the camera housing. For example, at some sites the camera shifted due to snow accumulation on top of the housing. This changed the location of the depth stakes in relation to the linear masks. When this happened a substantial jump in snow depth occurred that did not match the amount of total precipitation or melt recorded by the total precipitation gauges or the lysimeters. Another common problem was frost build up on the camera housing during cold clear periods or condensation during warm, humid periods. In both instances the images had more “white” pixels (i.e. snow) than there actually was. These anomalies were easily identified because large increases in snow canopy cover or snow depth were occurring when the meteorological data indicated either warm, humid rainy conditions or cold clear conditions. Anomalies were investigated by visually inspecting the photo sequences and making the required adjustment to the data.

4.3 Results and discussion

A dataset collected from February 20 to March 12, 2007 at LR13m was selected to show the dynamic climate conditions which occur in the ROS zone of Russell Creek Experimental Watershed. Figure 4.4 provides a photo sequence as an example of loading and unloading of snow from the canopy on February 23 and 24, 2007. Loading and unloading of snow from the canopy between February 20th and 27th, 2007 produced very little output from the lysimeter and a steady increase in SWE on the scale (Figure 4.5). Shortly after throughfall SWE peaked on March 3rd, temperatures rose and a rainfall event caused rapid depletion of snow below the canopy and substantial output from the lysimeter, with SWE on the scale approaching zero on March 06, 2007. As temperatures dropped on March 07, 2007, precipitation fell as snow or rain and snow interception began again, with rapid loading and unloading for the remainder of the analysis period, combined with high lysimeter output. This dataset shows the importance of daily temperature fluctuations in determining rain vs. snow along with storage of snow in the canopy. While these data cannot distinguish between mass release, melt water drip, sublimation and evaporation, it suggests that melt water drip and mass release were the primary forms of snow removal from the canopy during this period. This observation is supported by work done in the Oregon Cascades with a climate similar to that of Russell
Creek indicating that mass release and meltwater were the dominant mechanism for release of snow from the canopy (Storck et al., 2002).
Figure 4.4. Example of time-lapse photo sequence between February 23, 2007 at 17:15 and February 24, 2007 at 15:43. Note that images taken between February 23, 2007, 17:15 and February 24, 2007, 07:53 were not used for analysis because it was dark.
Figure 4.5. Example of climate and snowfall data collected between February 20, 2007 and March 12, 2007 at Lower Russell (LR) 13m. (a) Total precipitation in an opening and temperature below the forest canopy. Note that a snow course was done on February 20th, 2007, but one was not done on March 12 because there was no snow. (b) Lysimeter output in mm/hr in the regenerating forest. (c). Canopy snow cover fraction, throughfall SWE and depth. A value of 1 indicates the entire canopy is white, 0 indicates no snow in the canopy.

It is interesting to note that mass release events from the canopy usually led to increases in throughfall SWE on the scale, however, snow depth was often unchanged. When the images were examined more closely it was noticed that snow released during the unloading events occurred over a broad, seemingly random area. Where clumps hit the ground, they compacted the snow without altering the depth, presumably by increasing the density. Because a limited number of snow depth stakes were used, some of the small changes in depth associated with the falling clump distribution may have
been missed. Throughfall from large, sustained snowfall events tended to produce less meltwater drip and more sloughing of snow from branches as mass release, especially if the capacity of the tree to hold the snow was exceeded (data not shown). This led to larger increases in snow depth below the canopy over a greater area, which the design captured more effectively. When large snowfall events occur, we would expect to see a stronger relationship between increases in throughfall SWE and increases in depth. It is clear that the use of a camera to capture changes in depth and throughfall SWE provided much more data than using only a sonic snow depth sensor, especially for determining antecedent snow conditions prior to a ROS event under forest canopies. It is important to note that because the scale was elevated above the ground, melt rates differed compared to snow on the ground.

Prior to installation of the cameras none of the data described above could be collected. The established weather stations in Russell Creek measured temperature, precipitation, incoming solar radiation, relative humidity and wind speed/direction. Snow depth could have been measured using ultra-sonic snow depth sensors at the weather stations, but they were more expensive than the digital cameras and would provide less information. In addition, they would not provide data about conditions of snow within the canopy. Using only data from the meteorological network, lysimeter output could be modelled using a full energy balance equation, assuming our design captured the variability of snowmelt processes. By adjusting parameters the model could be calibrated to match snow lysimeter output. However, it is likely that this calibration would suffer from equifinality, in part due to the uncertainty as to whether precipitation is falling as rain or snow and how long precipitation remains in the canopy.

This effect of uncertainty surrounding snow accumulation and ablation becomes extremely important in the ROS zone because temperatures are usually near freezing for many parts of the winter season and routinely fluctuate above and below 0°C on a daily basis. Interception depends on the state of precipitation, with trees generally having a higher capacity to store snow than rainfall (Lundberg and Halldin, 2001). In addition, precipitation stored as snow is not readily available for output as water to the soil profile, requiring additional energy for melt. As a general rule, snow within the canopy will melt
faster than snow on the ground below the canopy, due to a higher surface area and more exposure to turbulence and warmer air. Storck et al. (2002) found that approximately 70% of snow released from the forest canopy is through meltwater drip. This can lead to significantly more SWE in an opening than a forest, and result in higher water delivery potential to the soil profile (Harr 1981; Coffin and Harr, 1992; Marks et al., 1998). In order to design or construct hydrological models to assess the impact of land use or climate change on water quantity, we will need to use methods similar to those presented in this chapter.

A number of limitations are associated with the experimental design. Light conditions affected image analyses, with images taken during the night being unusable in many instances. Differing illumination in the images was also a challenge that was not fully overcome using a single colour spectrum to develop thresholds for snow. There was a distinct diurnal signal in the canopy analysis associated with changes to illumination that we were unable to remove from the dataset. In fact we were not able to assess photos efficiently using automated techniques over the three years of the study period, forcing much of the image analysis to be done by hand. Despite these limitations, an incredible amount of data were collected for snow depth at all locations. We were also able gain a qualitative understanding of snow dynamics within different forest types through the process of examining the images, which proved useful in the interpretation of model results in the next chapter. Data from both the canopy and snow depth measurements will be used extensively in the next chapter for validation of model outputs. Future work in photogrammetric analysis of these types of images would benefit from the development of an index similar to the RGB Normalized Difference Snow Index (Hinkler et al., 2002) which takes into account differing illumination.

During large snow storm events the scales would reach maximum capacity. This occurred on only a few occasions, but field visits were required to remove the snow and reset the scales. It is challenging to design a scale sensitive enough to detect small changes in SWE and have the capacity to hold large amounts of snow. As well, the scales were too small to capture the variability of throughfall, and if used again a larger single or multiple small scales should be used. Because of the above problems we were
unable to collect continuous datasets. We were also unable to capture these data efficiently using automated procedures and so it will not be used further in the thesis.

The study design initially placed a great deal of focus on the lysimeter data and it was assumed the design would overcome difficulties encountered in past research, such as freezing tipping buckets or drains (e.g. Coffin and Harr, 1992). Unfortunately, these problems could not be surmounted and the lysimeter data were deemed largely unreliable and were used sparingly in the analysis. In many instances we could not verify whether differences in lysimeter outputs were due to frozen buckets, preferential flow past the lysimeters or storage of rain within the snowpack. Even though snow temperature data were available and sites were visited routinely, observations of the snowpack led us to conclude that for many ROS snow events, the lysimeter data were not reliable. An observation which called into question the lysimeter data were the large number of ice lenses in the snowpack, increasing the potential for diversion of water away from the lysimeter. The lysimeters were also designed for shallow snowpacks, with the sheer mass of snow at some locations damaging the installation. Future work will focus on creating larger footprint lysimeters that will better capture the variability of melt through the snowpack, although in certain conditions it is obvious that snowmelt lysimeters are wholly unreliable.

4.4 Conclusion

Even though the experimental design had certain limitations, it was a valuable addition to meteorological installations providing frequent observations during ROS. One of the major advantages was the low cost of purchase and installation of a time-lapse camera, snowmelt lysimeter, depth stakes and a spring scale, allowing for an extensive network to better capture spatial and temporal variability. Data were collected that traditional monitoring networks would miss. In addition, photogrammetric analysis software provided effective image processing, although development of a snow index is required to improve efficiency. Data of this quality can be used to develop new models or calibrate and test assumptions of existing hydrological models. It also provides insight into snow accumulation and melt dynamics in watersheds with rapidly changing meteorological conditions. The cameras used in this project could be used to monitor
meteorological stations for quality assurance and quality control to identify snow covered pyranometers, snow bridged precipitation gauges and ice rimed anemometers. Data collected using the methods described above will be used to test and validate a physically based snow interception and melt model in the next chapter.
Chapter 5: Application of the Cold Regions Hydrological Model within forested stands frequented by rain-on-snow

5.1 Introduction

Hydrological modelling combined with multiple scenario analysis has proven popular in the past two decades to assess the potential effects of forest management and large scale disturbance on streamflows. In British Columbia, this type of analysis has been primarily confined to the interior of the province in snow-dominated environments (Alila and Beckers, 2001; Kuraś et al., 2012; Whitaker et al., 2003) and in one case to a rain-dominated watershed (Beckers and Alila, 2004). Within the Pacific Northwest (PNW), hydrological models have been used to assess the potential effects of forest harvesting on snow accumulation and melt at the stand level (Andreadis et al., 2009; Marks et al., 1998; Storck, 2000; Storck et al., 2002; van Heesjwick et al., 1996) with a primary focus on the differences between clear-cut and old growth or mature forests. More recently, models have been used with mixed success to compare the differences in snow accumulation and melt among different forest types, however, most of the focus has been in continental environments. The majority of model development and testing has occurred in forests where sublimation of snow from the canopy is a primary mechanism of snow loss (e.g. Ellis et al., 2010; Essery et al., 2009; Rutter et al., 2009), which is generally less prevalent in the maritime or warmer portions of watersheds where the majority of snow is removed from the canopy through meltwater drip and mass release (Storck et al., 2002). In the SnowMIP2 comparisons of multiple model types among different climate regimes, model performance generally was poorer the warmer the environment or the warmer the winter, due in part to the difficulty in differentiating rain from snow when temperatures routinely fluctuate above and below freezing (Essery at al., 2009; Rutter et al., 2009). Models also differed in how meltwater and mass release were portioned from the canopy, resulting in differing accumulation patterns under forested stands (Essery et al., 2009). Both of these processes are of particular importance in watersheds that routinely receive ROS events.

Despite models being applied extensively at the plot scale, there has been minimal model application to regenerating forests. Measuring hydrological recovery for both
annual water yield and peak flow generation is a key component of forest management, with level of recovery often dictating future operational planning. It seems logical that hydrological recovery within a forest is directly related to the height and crown closure or leaf area index (LAI). The presence of a forest canopy can have major effects on interception, snow distribution and energy available for melt (e.g. Ellis et al., 2010; Pomeroy et al., 1998; Storck, 2000). Exposure of snow to the varying components of energy while stored in the canopy depends on species and the physical characteristics of the tree and the stand. Older trees will generally have thicker branches and deeper canopies which can hold more snow, thus intercepted snow will be exposed to energy fluxes for longer periods than in younger forests with smaller branches and shallower canopies, increasing potential for evaporation, sublimation and melt. In addition, the density of a forest canopy is the primary control on the attenuation and reflectance of solar energy and release of longwave energy, with a direct impact on net radiation to drive snowmelt. Forests play a key role in reducing turbulent exchanges required for sensible and latent heat fluxes, with wind speeds being greatly reduced under forest canopies compared to openings (Berris and Harr 1992; Marks et al., 1998). Wind speeds under forest canopies depend on canopy closure and tree height, so forests of different ages and structure will have varying degrees of turbulent exchange.

In watersheds frequented by rain-on-snow, high intensity energy inputs associated with warm, wet and windy frontal systems often generate peak flows of concern (e.g. Marks et al., 1998). Forests tend to moderate ROS energy inputs with much lower melt rates than in openings (Berris and Harr, 1987; Coffin and Harr, 1992; Marks et al., 1998). It is these differences in energy inputs during ROS that must be quantified in regenerating stands to assess stand level hydrological recovery. Snowmelt modelling using an energy balance approach provides the theoretical means to accomplish this.

The snow cover energy and mass balance model (SNOBAL) (Marks et al., 1999) has been successfully used in coastal watersheds in the Pacific Northwest (Marks et al., 1998; Mazurkiewicz et al., 2008; van Heeswijck et al., 1996). Recently, SNOBAL has been incorporated in the Cold Regions Hydrological Model (CRHM) which is a suite of empirical and physically based models (MacDonald et al., 2010; Pomeroy et al., 2007)
providing an opportunity to use SNOBAL with the interception models (e.g. Ellis et al., 2010) found within CRHM. Of the 33 models compared in SnowMIP2, CRHM had among the lowest normalized Root Mean Square Error (RMSE) for open and forested plots. To our knowledge CRHM has never been tested in coastal watersheds where ROS events occur frequently.

Model uncertainty is still a major concern when attempting to apply results to operational and problems. To determine the effects of forest management on hydrological processes using models, it will be important to test individual models in different environments using multiple types of observations to assess performance. Because of this, a novel approach to collecting snow depth data and interception observations combined with snow course measurements within regenerating and old growth forests in a watershed frequented by ROS will be utilized to evaluate an interception and snowmelt model within the CRHM framework. High temporal resolution snow depth data collected via a network of digital time-lapse cameras at varying elevations and in different forest types will comprise the primary observations to test model performance. The time-lapse images of canopy interception will be used to qualitatively assess simulations of snow interception, primarily loading and unloading of snow from the canopy. A combination of traditional model indices using snow depth and snow water equivalent (SWE) observations and “soft” qualitative data (Seibert and McDonnell, 2002) from time-lapse images will be used to assess model performance. Combining “hard” data and model indices with “soft” data and qualitative assessments of model performance will improve validation of model structure and outputs. Attention will not only focus on model performance over the entire season, but also on distinct snowpack development phases. Limitations and suggestions for model improvement will be discussed for future research and operational applications. A discussion will follow on the similarities and differences in energy available for melt among forest stands across different elevations over the entire season.

5.2 Model description

The Cold Regions Hydrological Model (CRHM) is a modelling platform that allows users to create a framework based on the amount of data required to drive the
model and meet the objectives of the project (Pomeroy et al., 1998). The model will be applied to the forested locations described in Chapter 3. Table 5.1 provides a summary of stand characteristics of the different forest types. One of the distinguishing characteristics of CRHM is a lack of calibration routines, thus users are required to address deficiencies either through improved data or by changing the algorithms that drive the model. As described in Pomeroy et al. (2007) CRHM is comprised of 4 major components: (1) Observations; (2) Parameters; (3) Module; and (4) Variables and States. The following is a brief description of each component:

1. Observations: These are the primary inputs to CRHM and are either collected directly through a meteorological station at the research location in question or derived from existing data through known relationships. In this application, observation data were provided at an hourly time step. The following meteorological forcing data were required for simulation using CRHM:
   a. Air temperature, \( T_a (^\circ C) \);
   b. Humidity as relative humidity, \( RH (%) \) or vapour pressure, \( e_a (kPa) \);
   c. Precipitation, \( P (mm) \);
   d. Wind speed, \( u (ms^{-1}) \);
   e. Shortwave radiation, \( R_s (W m^{-2}) \);
   f. Longwave radiation, \( R_l (W m^{-2}) \) (was estimated using \( T_a \) and \( e_a \) (eq.4))
   g. Soil temperature, \( T_s (^\circ C) \).

2. Parameters: Includes information about the physical characteristics of the site such as latitude, elevation, slope, aspect, tree height, LAI, heights of \( u, T_a \) and depth of \( T_s \) in relation to the soil surface. Within CRHM, Hydrological Response Units (HRU) are assigned a unique set of parameters describing their physical characteristics. Each plot as described in chapter 3 were defined as individual HRU’s. Details for specific parameters used at each plot are found in Table 5.1.
3. Modules: A selection of algorithms which are best able to meet the available data and the objectives of the user.

4. Initial States and variable: this provides the initial conditions at the start of each model run.

5.2.1 Modules

The energy balance is comprised of the major heat fluxes available for snowmelt ($Q_M$), as follows:

$$Q_M = R_s + R_l + H + L + G + A$$  

where $R_s$ is the net shortwave radiation (total solar energy absorbed by the snowpack, primarily controlled by surface albedo), $R_l$ is the net longwave radiation, $H$ is the sensible heat exchange (the energy associated with a change in temperature, or the convective transfer of heat at the snow-air interface, relying on turbulent exchange), $L$ is the latent heat exchange (the energy associated with evaporation, condensation and sublimation, relying on turbulent exchange), $G$ is the heat flux generated from the ground (coastal soils rarely freeze in the winter, thus it is important to account for this flux) and $A$ is the advected heat from precipitation (heat is transferred from the rain to the snowpack due to temperature differences). If $Q_M$ is positive the snowpack absorbs energy reducing its overall cold content. Once the cold content is depleted and the snowpack is brought to an isothermal state, net positive energy inputs will result in snowmelt.

The following modules were selected to account for the energy balance components described above in open and forested environments:

1. Global and Slope_Qsi: Utilizes $R_s$ adjusted for slope, aspect and elevation (Garnier and Ohmura, 1970);

2. Long: calculates incoming long wave radiation based on $T_a$ and $e_a$ (Sicart et al., 2006);

3. Canopy: accounts for energy exchanges within the canopy for sublimation, evaporation, meltwater drip and mass release to the canopy floor, as well as
shortwave and longwave radiation adjustments below the forest canopy (Ellis et al., 2010);

4. albedo_Richard: calculates snow albedo based on decay since last snowfall (Essery and Etchevers, 2004)

5. SNOBAL: the primary snowmelt model (Marks et al., 1998);

5.2.1.1 Global and Slope_QSI

Solar radiation’s theoretical direct beam component on a sloping surface \( Q_{dir} \) is determined using the Global module in CRHM, based on the following equation (Garnier and Ohmura, 1970):

\[
Q_{dir} = I \cdot p^n \left[ \left( \sin \theta \cos H_a \right) \left( - \cos A_s \sin Z \right) - \sin H_a \left( \sin A_s \cos Z \right) + \left( \cos \theta \cos H_a \cos Z \right) \cos \delta + \left[ \cos \theta (\cos A_s \sin Z) + (\sin \theta \cos Z) \right] \sin \delta \right]
\]  

(2)

where \( I \) is the intensity of radiation from space, \( p \) is the atmosphere’s mean zenith path transmissivity, \( m_o \) is the optical air mass, \( \delta \) is the declination of the sun, \( \theta \) is the latitude, \( H_a \) is hour angle from solar noon, and \( A_s \) and \( Z \) are the slope aspect and angle, respectively. To calculate the diffuse clear sky radiation, \( Q_{dif} \), the following equation was used (List, 1968):

\[
Q_{dif} = 0.5 \left( (1 - aw - ac)/(Q_{ext} - Q_{dir}) \right)
\]  

(3)

where radiation absorbed by water is represented by \( aw \), \( ac \) is ozone absorbed radiation, and \( Q_{ext} \) is the extraterrestrial radiation on a horizontal surface at the top of the earth’s atmosphere.

The Slope_QSI module estimates shortwave radiation on a slope from the measured shortwave radiation at the site from a level surface in combination with the direct and diffuse radiation calculated using the Global module.

5.2.1.2 Long

The Long module calculates longwave irradiance in open environments \( L_o \) based on the following equation (Sicart et al., 2006):

\[
L_o = 1.24 \left( \frac{e_a}{T_a} \right)^{1/7} (1 + 0.44RH - 0.18\tau_{atm})\sigma T_a^4
\]  

(4)
where \( \sigma \) is the Stephan-Boltzmann constant \( (5.67 \times 10^{-8} \text{W m}^{-2} \text{K}^{-4}) \). Atmospheric transmissivity to solar radiation on a cloudy day is represented by the index \( \tau_{atm} \):

\[
\tau_{atm} = \frac{K_{in}}{Q_{ext}}
\]  

(5)

5.2.1.3 Canopy

The Canopy module calculates energy and water/snow input to the ground beneath the forest canopy (Ellis et al., 2010). The module is constructed of multiple components that account for storage of snow in the canopy where it is exposed to the various energy components that drive melt, evaporation and sublimation. In addition, the module simulates reduced transmittance of shortwave radiation through the canopy and enhanced longwave irradiance from the forest canopy to the snow surface. This module simulates unloading from the canopy due to physical exceedence of the branches and the breakdown of the physical structure of the snow due to increasing air temperature.

Within the Canopy module a forested environment snow water equivalent (SWE) is expressed in the following mass balance equation:

\[
\text{SWE} = \text{SWE}_o + (P_s - (I_s - U_i) + P_r - (I_r - R_d) - M)t
\]  

(6)

where \( \text{SWE}_o \) is the antecedent conditions, \( I_s \) is the canopy snow interception rate, \( U_i \) is the rate of canopy unloading, \( P_s \) and \( P_r \) are the snowfall and rainfall rates, respectively, \( I_r \) is the rainfall interception rate of the canopy, \( R_d \) is the rate of canopy drip, \( M \) is the melt loss rate and \( t \) is the time step for the model calculation.

Snowfall interception is governed by the physical factors of the individual tree structure, density of forest and the antecedent intercepted snow load \( (I_{s,o}) \); the Canopy module deals with this dynamically through the following expression:

\[
I_s = (I_s^* - I_{s,o})(1 - e^{-C_l P_s I_s^{*} / I_s^*})
\]  

(7)

Where \( C_l \) is the canopy leaf contact area per unit ground and \( I_s^* \) is the maximum intercepted snow load which is determined using the following:

\[
I_s^* = S(0.27 + 46/\rho_s)\text{LAI}
\]  

(8)
where $S$ is mean maximum snow load per unit branch, $\rho_s$ is the density of falling snow and LAI is the leaf area index. The $S$ is usually input as a single value derived from literature, however, through the study design described in chapter 3 a selection of events during heavy snow fall was closely examined to determine thresholds for mass release due to exceeding capacity. This was accomplished by monitoring the time-lapse photos of canopy interception to identify when mass release events occurred due to physical exceedence and comparing to total precipitation during this period. Results from this analysis will be presented later in this chapter.

Sublimation of intercepted snow from the canopy is estimated by (Pomeroy et al., 1998b) via:

$$q_e = V_i I_s$$

(9)

where $q_e$ is the canopy sublimation flux, and $V_i$ is determined by:

$$V_i = V_s C_e$$

(10)

where $V_s$ is the sublimation flux for a 500 $\mu$m radius ice sphere and $C_e$ is the intercepted snow exposure coefficient. $C_e$ is calculated as follows (Pomeroy and Schmidt, 1993):

$$C_e = k \left( \frac{I_s}{I_s^*} \right)^{-F}$$

(11)

where $k$ is a dimensionless coefficient indexing the structure and age of the snow and $F$ is derived empirically as 0.4. To determine wind speed within the canopy, the following expression was used:

$$u_\xi = u \varepsilon^{-\psi \xi}$$

(12)

where $u_\xi$ is the estimated canopy wind speed, $\psi$ is the canopy wind speed extinction coefficient which is a linear function of LAI (Eagleson, 2002). Mass release of snow from the canopy as a result of increasing temperature was originally based on specifying a threshold ice bulb temperature in which all remaining intercepted snow was released
when exceeded for 3 hours. This threshold was too simplistic and was adjusted to account for melt water drip as described below.

It is extremely difficult to differentiate proportions of snow released from the canopy as mass release and melt water drip, even with detailed measurements of snow load within the canopy (e.g., Storck et al., 2002). When CRHM and the Canopy module was initially tested at Russell Creek, snow depths under the forested stands were greatly overestimated, with mass release of solid snow being the dominant component. As a result, snow levels under the canopy increased throughout the entire season with minimal melt. Snow melting from the canopy was routinely observed both in the time-lapse photography and while in the field, thus a new parameter was incorporated into the model based on ice bulb temperature thresholds for melt water drip and mass release. Canopy drip \((R_d)\) was calculated using the following:

\[
R_d = I_s \times \frac{(T_i - T_{us})}{(T_{uw} - T_{us})}
\]  

(13)

where \(T_i\) is the ice bulb temperature, \(T_{us}\) is a user defined ice bulb temperature threshold that unloads solid snow and \(T_{uw}\) is a user defined ice bulb temperature that unloads snow as melt water. \(T_i\) is calculated with the following equation:

\[
T_i = T_a \left( \frac{H_s}{C_i} \right)
\]  

(14)

where \(H_s\) is the heat of sublimation \((2.838e-6 \text{ MJ/kg})\) and \(C_i\) is the heat capacity of ice \((2.102e-3 \text{ MJ/kg/K})\).

Shortwave radiation transmission to the snow surface through the forest canopy is expressed as:

\[
K_{inF} = K_{in} \tau_{fin}(1 - \alpha_s)
\]  

(15)
where $K_{inF}$ is the net shortwave radiation reaching the forest floor, $\tau_{fst}$ is the forest shortwave transmittance estimated by the following equation (Pomeroy and Dion, 1996; Pomeroy et al., 2009):

$$\tau_{fst} = e^{\left( \frac{1.08 L_{\rho} \cos(\theta_s) \text{LAI}}{\sin(\theta_s)} \right)}$$

(16)

where $\theta_s$ is the solar angle above the horizon.

Longwave radiation from the forest canopy to the snow surface, $L_{inF}$ is given by:

$$L_{inF} = v L_o + (1-v) \varepsilon_f \sigma T_f^4$$

(17)

where $v$ is the sky view factor, $\varepsilon_f$ is the forest thermal emissivity and $T_f$ is the forest temperature.

5.2.1.4 Albedo_Richard

Snow albedo, $\alpha$, is calculated based on an exponential decay function to an asymptotic minimum of 0.5 (or lower depending on user inputs) for each time step with snowmelt (Essery and Etchevers, 2004) as below:

$$\alpha \rightarrow (\alpha - 0.5)e^{-\Delta t/\phi} + 0.5$$

(18)

where $\Delta t$ is the time step length and $\phi$ is the time constant for melting snow. When snow is falling, albedo is increased using the following function:

$$\alpha \rightarrow \alpha + (\alpha_{\text{max}} - \alpha) \frac{P_s \Delta t}{10}$$

(19)

where $\alpha_{\text{max}}$ is the albedo reset value once it snows; 0.85 is used by the model for this experiment and 10mm of new snow is required to fully refresh albedo.

5.2.1.5 SNOBAL

SNOBAL (Marks et al., 1999; Marks and Dozier, 1992; Marks et al., 1998) has been applied across western North America to model snowmelt within a variety of regimes (Garen and Marks, 2005; MacDonald et al., 2010; Marks et al., 1999; Marks and
Dozier, 1992; Marks et al., 1998; Marks et al., 2002; Mazurkiewicz et al., 2008; van Heesjwick et al., 1996). SNOBAL is a two layer snow model, with upper and lower snow layers as described in Figure 5.1. The upper layer is a fixed depth where energy is received, while the lower layer represents the remaining snowpack. A detailed description of the model is given by Marks et al., (1998 and 1999).

The outputs from the models described above were used to drive SNOBAL and include the following:

- albedo from the Albedo Richard module
- net radiation (longwave and shortwave)
- modified wind speeds from canopy to derive the turbulent energy exchanges that act on the snowpack.
- roughness length for deriving turbulent exchanges was set at the default value of 1 cm.

The soil component within SNOBAL is comprised of a single layer, with depth of the soil temperature measurement defining soil thickness. Heat transferred to the active layer from the soil is based on a relationship between soil temperature and conductivity. Advection from rainfall during ROS events is driven by rain temperature which is assumed to be equal to measured air temperature.
Figure 5.1. Schematic diagram of the SNOBAL model (after Marks et al., 1998).
Table 5.1. Site description of research plots at Russell Creek Experimental Watershed and parameters used in the Cold Regions Hydrological Model (CRHM). Leading species, stems/ha and canopy fraction are not used in CRHM.

<table>
<thead>
<tr>
<th>Plot Name</th>
<th>Elevation (m)</th>
<th>Average Tree Height (m)</th>
<th>Leading Species</th>
<th>Stems /ha</th>
<th>LAI $^2$</th>
<th>Canopy Fraction</th>
<th>Slope $^o$</th>
<th>$T_{us}/T_{uw}$ $^3$ (°C)</th>
<th>Max/Min Albedo</th>
<th>$I_s^{*}$ $^4$ (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LRCC</td>
<td>580</td>
<td>0</td>
<td>-</td>
<td>0</td>
<td>0.0</td>
<td>0</td>
<td>25</td>
<td>84/50</td>
<td>-</td>
<td>84/50</td>
</tr>
<tr>
<td>LR4m</td>
<td>530</td>
<td>4.5</td>
<td>Fd, Hw</td>
<td>3617</td>
<td>0.62</td>
<td>0.51</td>
<td>15</td>
<td>84/40</td>
<td>7</td>
<td>84/40</td>
</tr>
<tr>
<td>LR13m</td>
<td>500</td>
<td>14.2</td>
<td>Fd, Hw</td>
<td>1625</td>
<td>2.02</td>
<td>0.88</td>
<td>17</td>
<td>84/30</td>
<td>12</td>
<td>84/30</td>
</tr>
<tr>
<td>LROG</td>
<td>580</td>
<td>37.6</td>
<td>Ba, Hw</td>
<td>621</td>
<td>1.44</td>
<td>0.77</td>
<td>23</td>
<td>84/20</td>
<td>40</td>
<td>84/20</td>
</tr>
<tr>
<td>USCC</td>
<td>750</td>
<td>0</td>
<td>-</td>
<td>0</td>
<td>0.0</td>
<td>0</td>
<td>20</td>
<td>84/50</td>
<td>-</td>
<td>84/50</td>
</tr>
<tr>
<td>US4m</td>
<td>740</td>
<td>4.1</td>
<td>Hw, Cw</td>
<td>3066</td>
<td>1.28</td>
<td>0.70</td>
<td>16</td>
<td>84/40</td>
<td>9</td>
<td>84/40</td>
</tr>
<tr>
<td>US11m</td>
<td>780</td>
<td>10.9</td>
<td>Hw</td>
<td>973</td>
<td>2.08</td>
<td>0.87</td>
<td>18</td>
<td>84/30</td>
<td>14</td>
<td>84/30</td>
</tr>
<tr>
<td>USOG</td>
<td>790</td>
<td>44.9</td>
<td>Ba, Hw</td>
<td>1172</td>
<td>1.93</td>
<td>0.86</td>
<td>21</td>
<td>84/20</td>
<td>40</td>
<td>84/20</td>
</tr>
</tbody>
</table>

$^1$ Fd – Pseudotsuga menziesii (Douglas fir); Hw – Tsuga heterophylla (Western hemlock); Cw – Thuja plicata (Western red cedar); Ba – Abies amabilis (Amabilis fir)

$^2$ LAI was calculated based on hemispherical photography

$^3$ Ice bulb temperature thresholds that when exceeded cause snow to be released as snow vs. melt water drip; used in Canopy module

$^4$ Maximum snow load before mass release; used in Canopy module.
5.2.2 Model evaluation

The model outputs, primarily SWE and snow depth were evaluated against observed snow depth collected using the time-lapse photography network and snow course measurements of depth and SWE. Because of limited resources and difficulties associated with the automated processing of the time-lapse images as described in Chapter 4, formal evaluation focused primarily on snow depth measurements. In addition, snow depth provided the highest temporal resolution observations and thus were the primary observations used to determine if the model simulated the primary energy inputs during ROS, even with the relatively simple snow densification routine in SNOWBAL (Marks and Dozier, 1992), snow depth was still the preferred observation to assess model performance. Two general time periods of 0800 and 1600 were selected for the photo analysis for depth to maintain the most consistent light possible; however, much of the analysis was still done manually. Model outputs of simulated snow temperature, cold content and snow density were also compared to observed values. Observed cold content $Q_{cc}$ was calculated using the following equation:

$$Q_{cc} = -C_l \cdot \rho_w \cdot h_m \cdot (T_s - T_m)$$  \hspace{1cm} (20)

where $\rho_w$ is the density of water (1000 kg m$^{-3}$), $h_m$ is the depth of snow (m) derived from the time lapse images, $T_s$ is the average snow temperature based on a linear interpolation between snow temperature measurements and $T_m$ is the melting point temperature ($0^\circ$C).

Simulations for SWE and snow depth were evaluated using the formal indices described below over the entire season and for a selection of ROS events. Use of “hard” data to assess model performance can sometimes mask the deficiencies/strengths of the model by focusing on performance over the entire season. The dataset included quantitative data from snow course measurements and the snow depths derived from the time-lapse photography network. In addition, to this quantitative data, there was an extensive qualitative dataset from the time-lapse images of the forest canopy. As discussed in the previous chapter, we were unable to derive indices of snow canopy fraction systematically for the entire data set; however, the data will be used to assess model performance for interception processes over short periods.
Use of so called “soft data” has been promoted recently for both model development and validation, and it has been shown to improve understanding of model strengths and weakness and for increasing overall model performance (Bachmair et al., 2010; Kirchner, 2006; Seibert and McDonnell, 2002). Assessing model performance can take multiple forms, often based on user defined quality classes such as “Good”, “Acceptable” and “Failed” (e.g. Bachmair et al., 2010). This type of classification provides direction for model improvement and a rationale for using model outputs for further research or operational applications.

In this chapter two general approaches were used to assess model performance of SWE and snow depth simulations. The first used traditional model assessment indices over the entire snow season, namely root mean square error (RMSE), model bias (MB) and the Nash Sutcliffe (NS) model efficiency index (Nash and Sutcliffe, 1970). The RMSE is an indication of the absolute error between the simulations and observations; the MB provides a comparison of the total model output to the total of observation and the NS provides an indication of how well the model outputs compare to simply using the average.

The RMSE is calculated as follows:

$$\text{RMSE} = \sqrt{\frac{1}{n} \sum_{i=1}^{n} (x_{sim} - x_{obs})^2}$$

where $x_{sim}$ and $x_{obs}$ are the simulated and observed values at a given time step for $n$ number of paired simulated and observed values.

Model bias (MB) is determined by:

$$\text{MB} = \frac{\sum_{i=1}^{n} x_{sim}}{\sum_{i=1}^{n} x_{obs}}$$

with MB values greater than 1 indicating an over-prediction by the model and values less than 1 signifying an overall under-prediction of the model.
The NS is calculated by:

\[
NS = 1 - \frac{\sum_{i=1}^{n} (x_{sim} - x_{obs})^2}{\sum_{i=1}^{n} (x_{sim} - x_{avg})^2}
\]

where \(x_{avg}\) is the mean value of \(n\) number of \(x_{obs}\) values. As model efficiency increases, the NS value approaches 1, indicating a perfect match between the model and the observed values. As NS approaches 0, both model outputs and average of the observations are equally valid representations of the data. When NS becomes negative it indicates an increasingly poor representation of the model. Correlation coefficients were calculated for simulated and observed snow depths over the entire year to assess how well the model captured patterns of snow accumulation and melt.

Because model indices applied to long-term data can sometimes mask the periods during the year when the model performs well or poorly, a second assessment was used based on a semi–quantitative assessment of simulated vs. observed depth for periods during the snow season. These periods were defined as “Early” when snow is transient in nature, the “Accumulation” phase corresponding with sustained temperatures below freezing and a general increase in the snowpack, the “Melt” phase associated with the early to late spring characterized by sustained decreases in snow depth, and a “Late” season phase where snowpacks once again become transient. To provide a semi-quantitative index that is intuitive in nature and corresponds to how we visually assess simulation vs. observation plots, a simple approach called “Percent Model Agreement” (PMA) was used to assess average percent of simulated snow depth above or below observed for each period with the following equation:

\[
PMA = \frac{1}{n} \sum_{i=1}^{n} \left( \frac{x_{sim}}{x_{obs}} \right) \cdot 100
\]

This approach can introduce bias into the assessment, especially when observations or simulations approach zero with percentages above or below having infinite scales. To suppress this bias, we assigned a maximum or minimum under or over prediction of
Furthermore, problems arose when observations or simulations were zero returning error values. This was addressed by having a minimum depth of 0.5mm. PMA was not calculated for SWE due to a limited number of observations. To assess general model behavior using this index, we created categories of model performance, with *Excellent* corresponding to a PMA of +10%, *Good* of +10-20%, *Adequate* +20-40%, *Poor* +40 to 50% and *Unacceptable* >50%. To account for the variability of the simulations above or below the observation during each period, the standard deviation (SD) of PMA input values was also calculated.

Correlation coefficients ($r$) of observed and simulated snow depth were calculated for Early, Accumulation, Melt and Late season periods. The higher the correlation, the better the model captured the variability associated with accumulation and ablation dynamics, the processes of greatest interest in high energy rain-on-snow environments where events generally span 1 to 3 days. To assess the strength of the model using $r$, values > 0.90 were assigned a rating of *Excellent* (E), 0.89-0.80 was *Good* (G), 0.79-60 was *Acceptable* (A), 0.60-0.50 was *Poor* (P) and <0.50 was considered *Unacceptable* (U). Using a combination of PMA and correlation quality ratings will aid in determining model performance for simulated absolute snow depth and general melt dynamics.

The final approach to model assessment utilized the time-lapse photography dataset to determine if the canopy interception module captured snow canopy interactions. Events were selected to cover a range of meteorological conditions leading to canopy loading and unloading. Assessment focused on relatively short periods of 1 to 2 weeks due to limitations in analyzing a large number of images. This assessment was limited to a general description of the events to highlight where the model performed well and where it was deficient. No formal indices were calculated because the assessment periods were short in duration and could be adequately described using the figures.

5.3 Results

5.3.1 Maximum canopy snow load

Maximum snow load capacities within the forests at Russell Creek were determined by assessing the canopy snow load images and precipitation rates. To limit
other factors which could result in mass release from the canopy, only periods with moderate to heavy snow and calm winds where temperatures remained below freezing were examined. There were 3 periods that met these criteria and had images suitable for analysis. Storage was calculated by examining the total precipitation that occurred during the period when the images could be assessed. Maximum recorded storage capacity ranged from 20 to 27mm SWE for all forest types. No instances of maximum recorded storage followed by complete mass release from the canopy were observed. During some periods following maximum storage, portions of snow fell from the canopy, but not enough to release all snow. Capacity to hold snow was similar among the forest types (Figure 5.2).

![Images showing maximum snow loading capacity from January 9, 2007 at the Lower Russell (LR) (520-560m a.s.l) and Upper Stephanie (US) (740-770m a.s.l.) 4m, 11m, 13m, and old growth (OG) forests. Snow received up to this point was 20mm at the LR weather station and 24mm at the US station. Snow began to release from the canopy shortly after due to an increase in temperature and winds at all locations.](image)

Initially these maximum snow load values were used within the interception module, however, the storage capacities were too high at the 4m, 11m and 13m forests with snow depth being substantially underestimated. Because of this, a sensitivity analysis was performed using a range of values by forest type, to select a maximum load capacity which produced the best snow depth simulations. Figure 5.3 provides an example of how different maximum interception loads influence snow canopy dynamics and snow depth: the lower the maximum capacity, the higher the resulting snowpack.
This sensitivity analysis resulted in the selection of maximum snow capacity thresholds for the interception module far below the maximum observed values at the 4m, 11m and 13m stands, ranging from 5 to 25 mm. Maximum inputs to the interception module for storage for the OG stands ranged from 20 to 40 mm.

![Figure 5.3. Canopy snow load dynamics and resulting simulated snow depth using maximum snow canopy thresholds at the Lower Russell (LR) (540-570m a.s.l.) (a.) old growth (OG) (10, 25 and 50mm) and (b) 13m (10, 20 and 25mm) forests. The dark grey line in the snow depth figures represent observations.](image)

### 5.3.2 Model verification

Model performance in simulating SWE varied depending on season, year and location. In 2006 and 2007 at LR, SWE was simulated reasonably well at the CC, 13m and OG plots, and substantially over estimated for the entire season at the 4m plot where observations were available (Figure 5.4). Even though it appeared the model simulated
SWE well at three of the four LR plots, density was generally underestimated (Figure 5.4). Over the season, MB indicated there was a general over estimation for LR CC (1.12), 4m (1.86) and OG (1.13), and a slight under prediction at 13m (0.97) (Table 5.2). Model efficiency as described by NS was higher at LR 13m (0.99) and OG (0.97) than CC (0.80) with 4m showing poor model efficiency (-20.60). The RMSE values mirrored the results of MB and NS. Snow levels at LR were largely transient in 2006-2007, thus there was limited opportunity to measure SWE during the scheduled snow courses and no early season data were collected.

Comparing observed to simulated snow depth LR in 2006-2007 provided additional insight into model performance (Figure 5.5). Assessing snow depth simulations showed a general decrease in NS compared to SWE at LR CC (0.20), 13m (0.65) and OG (0.52) over the entire season (Table 5.3). The indices for SWE showed poor model performance at LR4m in 2006-2007; however, simulated depth indices were stronger with a decrease in MB (1.22), a substantial increase in NS from -20.60 to 0.15, and a 6 fold decrease in RMSE (Table 5.3). Correlation coefficients ranged from 0.75 to 0.87 at LR plots indicating the model generally captured the patterns of accumulation and melt throughout the year.

In 2006 and 2007 the US SWE simulations over the entire season had NS values that were relatively strong CC (0.80), 4m (0.57) and 11m (0.63) (Table 5.2). The MB values for US CC (0.98) and 4m (0.84) indicate a slight under prediction of SWE, with a moderate over prediction at 11m (1.23). The model did poorly at simulating SWE at USOG over the entire season (NS = -23.45), generally over estimating SWE (MB 3.06), despite having similar SWE simulations for the first 2 observations of the season (Figure 5.4). During the late season melt, the model under-predicted SWE values at US 11m and OG, with a slight under prediction at the 4m (Figure 5.4). Error bars from snow course measurements suggest the model was able to capture the variability of the US SWE measurements for all plots except OG. The simulated values tracked the general accumulation and melt patterns collected by the snow course measurements at all sites. Simulated snow density was much closer to observed values than SWE, however, there was a general overestimation of density throughout the season at all plots (Figure 5.4).
Nash-Sutcliffe values for snow depth were similar to SWE for US CC (0.81) and OG (0.62) in 2006-2007, and moderately lower at 4m (0.35) over the entire season (Table 5.3). The US OG plot still had a negative NS value, however, it was only -0.76 for depth compared to -23.45 for SWE. Over the entire season, there was positive MB at US CC (1.04) and OG (1.34) and a negative bias at 4m (0.97) and 11m (0.90). The model was able to simulate the general snow depth dynamics at all US plots in 2006-2007 however, the correlation coefficients were weaker as a group compared to the LR plots with values ranging between 0.62 and 0.84.

The 2007-2008 season provided a strong contrast to 2006-2007 with generally cooler temperatures and considerably deeper snowpacks at both LR and US. When comparing simulated vs. observed SWE values at LR, the model over-predicted SWE over the entire season with MB greater than 1 at CC (1.73), 4m (1.71) and 13m (1.54), with the exception of OG (0.92). The model performed poorly at predicting SWE in 2007-2008 with NS values between -0.71 and -3.35 at the LR plots, suggesting the average SWE was a better predictor (Table 5.2). As in the previous year, snow density was under-predicted through the year at all LR and US plots (Figure 5.6). The observed density at all locations remained relatively consistent, with a general decrease throughout the measurement period while simulated density tended to increase, especially through the melt period at the LR and US plots. The model over-predicted SWE at LR CC, 4m and 13m throughout the 2007-2008 winter season where observations were available, with modeled SWE values converging at the start and the end of the melt seasons (Figure 5.6). The LROG simulations were quite close to the measured SWE values for 4 of the 6 observations, with the model under predicting SWE at the end of the winter.

As in the previous year, snow depth observations provided additional insight into model performance. The model performed far better at simulating snow depth compared to SWE at the LR locations in 2007-2008, with NS values ranging between 0.58 and 0.81 (Table 5.3). In addition, all correlation coefficients were greater than 0.90, indicating the model did well at capturing the overall accumulation and ablation patterns over the entire year. There was a general increase in model performance for snow depth from 2006-
2007 to 2007-2008 at the LR locations. The MB for the entire season showed a general over prediction at the LR plots ranging from 1.16 to 1.22 for depth.

SWE observations were only available after peak accumulation at US in 2007-2008 with MB values of 1.23 (CC), 1.19 (4m) and 1.20 (OG) showing an over prediction at all locations (US11m had no snow course measurements in 2007-08) (Table 5.2). Similar to LR, the US model simulations for SWE were poor, with negative NS values (-2.52 to -9.02) over the entire season. The first two SWE observations in 2007-2008 at the US locations were much lower than the simulated values (Figure 5.6). There was general agreement with the simulated and modeled values for the third SWE measurement in mid April. Following this measurement, the model under-predicted SWE the last two measurements at all US plots. Simulated snow density values were over estimated for the entire season at the US plots, even for the April SWE measurement where observations were in close agreements the simulated values.

Snow depth was under-predicted at all US plots in 2007-2008 over the entire season with CC and 11m having MB values of 0.99, and US 4m and OG with MB values of 0.87 and 84, respectively (Table 5.3). As a group, the US plots had the strongest model performance of all years and locations when simulating depth, with NS values between 0.73 and 0.84. The greatest improvement was seen at USOG, where the NS from 2006-07 went from -0.76 to 0.73 in 2007-08. The strength of these simulations corresponded to the coldest location (US) and year for the study.
Figure 5.4. Simulated (solid line) and observed (solid circles) snow water equivalent (SWE) and density in 2006-2007 at the Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests. Error bars are the standard deviation from the snow course measurements.
Figure 5.5. Simulated (solid black line), observed snow courses depth (solid circles) and observed snow depth (solid grey line) at the Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests in 2006-2007. Error bars are the standard deviation from the snow course measurements.
Figure 5.6. Simulated (solid line) and observed (solid circles) snow water equivalent (SWE) and density in 2007-2008 at the Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests. Error bars are the standard deviation from the snow course measurements.
Figure 5.7. Simulated (solid black line), observed snow courses depth (solid circles) and observed snow depth (solid grey line) at the Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests in 2007-2008. Error bars are the standard deviation from the snow course plots.
Table 5.2 Model bias (MB), Nash-Sutcliffe (NS) and root mean square error (RMSE) of snow water equivalent (SWE) simulations at the Lower Russell (LR) (520-560 m a.s.l.) and Upper Stephanie (LR) (740-770 m a.s.l.) clear-cut (CC), 4m, 11m, 13m, and old growth (OG) forests.

<table>
<thead>
<tr>
<th>Plot</th>
<th>MB</th>
<th>NS</th>
<th>RMSE [mm]</th>
</tr>
</thead>
<tbody>
<tr>
<td>LR-CC</td>
<td>1.12</td>
<td>1.73</td>
<td>0.80</td>
</tr>
<tr>
<td>LR-4m</td>
<td>1.86</td>
<td>1.71</td>
<td>-20.60</td>
</tr>
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<td>LR-13m</td>
<td>0.97</td>
<td>1.54</td>
<td>0.99</td>
</tr>
<tr>
<td>LR-OG</td>
<td>1.13</td>
<td>0.92</td>
<td>0.97</td>
</tr>
<tr>
<td>US-CC</td>
<td>0.98</td>
<td>1.23</td>
<td>0.80</td>
</tr>
<tr>
<td>US-4m</td>
<td>0.84</td>
<td>1.19</td>
<td>0.57</td>
</tr>
<tr>
<td>US-11m</td>
<td>1.23</td>
<td>NA</td>
<td>0.63</td>
</tr>
<tr>
<td>US-OG</td>
<td>3.06</td>
<td>1.20</td>
<td>-23.45</td>
</tr>
</tbody>
</table>

Table 5.3 Model bias (MB), Nash-Sutcliffe (NS) and root mean square error (RMSE) of snow depth simulations at the Lower Russell (LR) (520-560 m a.s.l.) and Upper Stephanie (LR) (740-770 m a.s.l.) clear-cut (CC), 4m, 11m, 13m, and old growth (OG) forests.

<table>
<thead>
<tr>
<th>Plot</th>
<th>MB</th>
<th>NS</th>
<th>RMSE [mm]</th>
<th>r</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>06-07</td>
<td>07-08</td>
<td>06-07</td>
<td>07-08</td>
</tr>
<tr>
<td>LR-CC</td>
<td>0.58</td>
<td>1.16</td>
<td>0.20</td>
<td>0.81</td>
</tr>
<tr>
<td>LR-4m</td>
<td>1.26</td>
<td>1.22</td>
<td>0.15</td>
<td>0.58</td>
</tr>
<tr>
<td>LR-13m</td>
<td>0.82</td>
<td>1.20</td>
<td>0.65</td>
<td>0.81</td>
</tr>
<tr>
<td>LR-OG</td>
<td>1.7</td>
<td>1.18</td>
<td>0.52</td>
<td>0.65</td>
</tr>
<tr>
<td>US-CC</td>
<td>1.04</td>
<td>0.99</td>
<td>0.81</td>
<td>0.84</td>
</tr>
<tr>
<td>US-4m</td>
<td>0.97</td>
<td>0.87</td>
<td>0.35</td>
<td>0.80</td>
</tr>
<tr>
<td>US-11m</td>
<td>0.90</td>
<td>0.99</td>
<td>0.62</td>
<td>0.82</td>
</tr>
<tr>
<td>US-OG</td>
<td>1.34</td>
<td>0.84</td>
<td>-0.76</td>
<td>0.73</td>
</tr>
</tbody>
</table>

5.3.2.1 Model performance during ROS

Because of the importance of snowmelt on streamflow and landslides during ROS, model indices were also calculated for a selection of melt events. Periods of snowmelt during ROS were the focus, however, due to the timing of ROS and the monitoring of snow depth, ROS events could not be fully isolated, thus there were periods of ablation where rain was not occurring. In addition, because of problems associated with the cameras, an equal number of ROS events could not always be compared. Despite this, simulations and observations of snow depth for 25 ablation events in 2006-2008 where ROS occurred for the majority of the ablation period were compared. For both the LR and US plots, NS ranged from 0.69 to 0.86, with the LR and US 4m plots having the highest model efficiency and the LR 11m and US 13m plots having the lowest (Table 5.4). Model bias ranged from 0.73 to 0.89 for all plots except...
LR13m which was 1.05, suggesting the model generally under predicts ablation rates during events where ROS is the dominant condition. Correlation coefficients were all greater than 0.90, indicating the model performed well at capturing the snow depth ablation dynamics at all plots during ROS.

Table 5.4. Model bias (MB), Nash-Sutcliffe (NS), root mean square error (RMSE) and correlation coefficients of snow depth ablation for 25 ROS events at the Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (LR) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m, and old growth (OG) forests.

<table>
<thead>
<tr>
<th>Plot</th>
<th>MB</th>
<th>NS</th>
<th>RMSE [mm]</th>
<th>r</th>
</tr>
</thead>
<tbody>
<tr>
<td>LR-CC</td>
<td>0.72</td>
<td>0.80</td>
<td>0.048</td>
<td>0.93</td>
</tr>
<tr>
<td>LR-4m</td>
<td>0.84</td>
<td>0.86</td>
<td>0.021</td>
<td>0.92</td>
</tr>
<tr>
<td>LR-13m</td>
<td>1.05</td>
<td>0.73</td>
<td>0.032</td>
<td>0.93</td>
</tr>
<tr>
<td>LR-OG</td>
<td>0.88</td>
<td>0.76</td>
<td>0.018</td>
<td>0.90</td>
</tr>
<tr>
<td>US-CC</td>
<td>0.89</td>
<td>0.74</td>
<td>0.063</td>
<td>0.95</td>
</tr>
<tr>
<td>US-4m</td>
<td>0.82</td>
<td>0.85</td>
<td>0.069</td>
<td>0.92</td>
</tr>
<tr>
<td>US-11m</td>
<td>0.75</td>
<td>0.69</td>
<td>0.055</td>
<td>0.91</td>
</tr>
<tr>
<td>US-OG</td>
<td>0.73</td>
<td>0.76</td>
<td>0.038</td>
<td>0.92</td>
</tr>
</tbody>
</table>

5.3.2.2 Model assessment by snowpack development phase

Model performance was strongest during the accumulation period when comparing absolute differences using the PMA in both 2006-07 and 2007-08, with only LR13m and US11m having Poor and Unacceptable model performance in 2007-2008 with over-simulated depths during this period (Table 5.5). All plots that received Good or Acceptable ratings during the accumulation period had standard deviations that were greater than the PMA, indicating a high degree of variability. The Early and Late season periods generally had the worst model performance (Table 5.5). In 2006 and 2007 during the Early season, the model under-predicted snow depth at all locations and had Poor or Unacceptable model performance, except for USCC where it generally over-predicted depth and had Good model performance.

The model performed better in 2007-2008 during the Early season, with 4 of the 8 locations having Acceptable or Excellent ratings, however, the standard deviations were all larger than the PMA indicating there was a high degree of variation between observed and simulated snow depth (Table 5.5). Late season model performance was Poor to
Unacceptable at all locations, except LR4m and LR13m which received Acceptable ratings in 2007-2008, with LR4m having a standard deviation less than the mean.

Depth during the snowmelt phase was over-predicted at all locations except at LR13 and LROG in 2006-2007 and USCC and US4m plots in 2007-2008 (Table 5.5). In general, simulations during the Melt phase had stronger PMA values at US than LR. Simulations for all periods were strongest at USCC in 2007-2008 with all PMA values within +21% of the observed.

The correlation coefficients ranged from 0.62 to 0.96 on an annual basis, with the majority having r values greater than 0.8, indicating the model performed well at tracking patterns through the year. However, when examining each period in more detail, it became apparent the model performed generally worst in the early season (Table 5.6). Strength of correlations during the Accumulation phase varied by year, with r values ranging from 0.52 to 0.89 in 2006-2007. In 2007-2008 all but LR4m (r=0.78) had ratings of Good or Excellent during the accumulation phase. Correlation coefficients were highest during the Melt phase with r values between 0.84 and 0.99 and all simulations receiving Good or Excellent ratings. This is in contrast to comparatively weaker PMA values during the melt phase which suggests the model generally over-predicted depth for the duration of melt, but performed well at capturing the general dynamics. Simulations of late season snowpacks were mostly Good or Excellent, with the notable exceptions of Poor and Unacceptable ratings at LRCC in 2006-2007 and LROG in 2007-2008. Poor and Unacceptable ratings generally occurred when simulated snowmelt finished either too early or too late. Overall, the model performed well at capturing the patterns of snow ablation and melt among the different forest types.
Table 5.5. Percent Model Agreement (PMA), standard deviation (SD) and model performance rating (R) by snowpack development phase at the Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (LR) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m, and old growth (OG) forests.

| Site | Year | PMA | SD | R' | PMA | SD | R | PMA | SD | R' | PMA | SD | R' |
|------|------|-----|----|----|-----|----|---|-----|----|---|-----|----|---|----|
| LRCC | 06-07 | -82 | 21 | P | -19 | 26 | G | 28 | 20 | A | -47 | 55 | P |
|      | 07-08 | -21 | 67 | A | -5  | 13 | E | 63 | 67 | U | 310 | 700 | U |
| LR4m | 06-07 | -73 | 28 | P | 20  | 26 | A | 22 | 21 | A | NA  | NA  | NA |
|      | 07-08 | 46  | 104| P | -5  | 41 | E | 38 | 12 | P | 24  | 9   | A |
| LR13m| 06-07 | -54 | 38 | P | -22 | 29 | A | -26 | 25 | A | 36  | 77  | P |
|      | 07-08 | 569 | 1200| U | 45  | 38 | P | 78 | 37 | U | 29  | 42  | A |
| LROG | 06-07 | -58 | 54 | P | NA  | NA | NA | -54 | 45 | U | -51 | 53  | U |
|      | 07-08 | -7  | 43 | E | -5  | 34 | E | 67 | 15 | U | 200 | 241 | U |
| USCC | 06-07 | 19  | 118| G | 25  | 47 | A | 72 | 21 | U | 225 | 116 | U |
|      | 07-08 | 21  | 134| A | 16  | 17 | G | -5 | 28 | E | NA  | NA  | NA |
| US4m | 06-07 | -53 | 21 | U | -8  | 31 | E | 39 | 16 | A | NA  | NA  | NA |
|      | 07-08 | -3  | 81 | E | NA  | NA | NA | -34 | 37 | A | NA  | NA  | NA |
| US11m| 06-07 | -44 | 33 | P | -22 | 25 | A | 32 | 19 | A | NA  | NA  | NA |
|      | 07-08 | 72  | 68 | U | 73  | 78 | U | 4  | 4  | E | NA  | NA  | NA |
| USOG | 06-07 | -45 | 33 | P | 1   | 45 | E | 104| 31 | U | 368 | 128 | U |
|      | 07-08 | 44  | 58 | P | -3  | 16 | E | 9  | -22| G | NA  | NA  | NA |

Ratings (R) are based PMA ranges where PMA +10% is Excellent (E); +10 to 20% is Good (G); + 20-40% is Acceptable (A); +40-50% is Poor (P), and; >+50% is Unacceptable (U). NA signifies that observations were unavailable or that period did not occur.
Table 5.6. Correlation coefficients ($r$) and performance ratings ($R$) by Early (E), Accumulation (A), Melt (M) and Late (L) periods of snowpack development at the Lower Russell (LR) (520-560 m a.s.l.) and Upper Stephanie (LR) (740-770 m a.s.l.) clear-cut (CC), 4m, 11m, 13m, and old growth (OG) forests.

<table>
<thead>
<tr>
<th>Plot</th>
<th>Year</th>
<th>Early</th>
<th>Accumulation</th>
<th>Melt</th>
<th>Late</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>$r$</td>
<td>$R^*$</td>
<td>$r$</td>
<td>$R^*$</td>
</tr>
<tr>
<td>LRCC</td>
<td>06-07</td>
<td>0.64</td>
<td>A</td>
<td>0.89</td>
<td>$G$</td>
</tr>
<tr>
<td></td>
<td>07-08</td>
<td>0.53</td>
<td>P</td>
<td>0.94</td>
<td>$E$</td>
</tr>
<tr>
<td>LR4m</td>
<td>06-07</td>
<td>0.68</td>
<td>A</td>
<td>0.62</td>
<td>$A$</td>
</tr>
<tr>
<td></td>
<td>07-08</td>
<td>0.73</td>
<td>A</td>
<td>0.78</td>
<td>$A$</td>
</tr>
<tr>
<td>LR13m</td>
<td>06-07</td>
<td>0.76</td>
<td>A</td>
<td>0.89</td>
<td>$G$</td>
</tr>
<tr>
<td></td>
<td>07-08</td>
<td>0.24</td>
<td>$U$</td>
<td>0.90</td>
<td>$E$</td>
</tr>
<tr>
<td>LROG</td>
<td>06-07</td>
<td>0.74</td>
<td>A</td>
<td>NA</td>
<td>NA</td>
</tr>
<tr>
<td></td>
<td>07-08</td>
<td>0.82</td>
<td>$G$</td>
<td>0.88</td>
<td>$G$</td>
</tr>
<tr>
<td>USCC</td>
<td>06-07</td>
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<td>$G$</td>
<td>0.56</td>
<td>$P$</td>
</tr>
<tr>
<td></td>
<td>07-08</td>
<td>0.61</td>
<td>$A$</td>
<td>0.95</td>
<td>$E$</td>
</tr>
<tr>
<td>US4m</td>
<td>06-07</td>
<td>0.83</td>
<td>$G$</td>
<td>0.70</td>
<td>$A$</td>
</tr>
<tr>
<td></td>
<td>07-08</td>
<td>0.54</td>
<td>$P$</td>
<td>0.97</td>
<td>$E$</td>
</tr>
<tr>
<td>US11m</td>
<td>06-07</td>
<td>0.94</td>
<td>$E$</td>
<td>0.73</td>
<td>$A$</td>
</tr>
<tr>
<td></td>
<td>07-08</td>
<td>0.51</td>
<td>$P$</td>
<td>0.95</td>
<td>$E$</td>
</tr>
<tr>
<td>USOG</td>
<td>06-07</td>
<td>0.84</td>
<td>$G$</td>
<td>0.52</td>
<td>$P$</td>
</tr>
<tr>
<td></td>
<td>07-08</td>
<td>0.62</td>
<td>$M$</td>
<td>0.96</td>
<td>$E$</td>
</tr>
</tbody>
</table>

* Ratings ($R$) are based correlation coefficient ($r$) ranges where $r$ $\geq$ 0.90 is Excellent (E); $0.89 < r > 0.80$ is Good (G); $0.79 < r > 0.60$ is Acceptable (A); $0.59 < r > 0.50$ is Poor (P), and; $r < 0.50$ is Unacceptable (U). NA signifies that observations were unavailable or that period did not occur.
5.3.2.3  **Interception dynamics**

The event described in chapter 4 was used to assess interception simulations at LR13m from February 20th to March 12th, 2007. The period was characterized by multiple days when temperatures fluctuated above and below freezing, resulting in precipitation alternating between rain and snowfall, a period of 2 days where temperatures were largely below freezing and a period of warm temperatures with substantial rainfall. The model simulated the loading and unloading dynamics of snow from the canopy well, with timing of peak snow load and mass unloading often within hours of the observed values (Figure 5.8). There were situations where the model unloaded all the snow from the canopy, when the images indicated otherwise such as on February 20-21st. Timing of observed peak canopy snow fraction was markedly different than simulated peak storage during the cold period from February 27th to March 1st. The observed snow canopy fraction peaked on February 27th and began to decline. The simulated snow canopy load continued to increase until March 1st; the model continued to load the canopy with snow because the precipitation observation indicated it was snowing, however, when the images were examined it was clear that no snow had fallen during this period. Even though the simulated and observed values diverged during this period, the timing of mass release for the simulated and observed were within 1 hour of each other for this event and all others where the model and observations indicated a mass release. The snow depth values tracked closely at the end of the event, compared to the early portion when depth was substantially under estimated by the model. In general, the model performed well at timing the release of snow and the dynamics of the snow in the canopy at LR13m.

An early season period was selected to assess the interception module’s ability to capture loading and unloading dynamics at LROG from November 12 to December 05, 2006 (Figure 5.9). This period was characterized by heavy snowfall, a large ROS event and a period with temperatures largely below freezing. In this case, the model did not perform as well at tracking the loading and unloading dynamics compared to LR13m. From November 11 to 16th the observations indicated a number of loading and unloading events with air temperature around the freezing mark. These conditions provided a good
test of the simulations due to uncertainty in differentiating rain from snow. When temperatures dropped below freezing on November 17\textsuperscript{th} the simulated initiation of snow loading was delayed compared to the observed, however, mass release of snow from the canopy on the afternoon on the 17\textsuperscript{th} was within 2 hrs of the observation. Beginning on November 21\textsuperscript{st}, there was a cooling trend and by the 25\textsuperscript{th}, temperatures remained below freezing until November 30\textsuperscript{th}. The simulations tracked observed canopy loading during this period and matched the timing of canopy release on the morning of November 30\textsuperscript{th} within one hour; however, the observed release of snow from the canopy suggested that the simulations underestimated the amount of snow released, with canopy snow storage remaining steady until the end of the observation period. These differences between observed and simulated canopy snow dynamics would have impacted the snow depth simulations, however, they did not explain all the differences that were observed during December 1\textsuperscript{st} to 5\textsuperscript{th} when the simulations show a much higher rate of ablation than observed. The fact there was enough energy to melt snow on the ground, but not from the canopy, suggests that the model over estimated the ground heat flux (resulting in too-high melt rates under the canopy) and/or underestimated snow density (resulting in an excessive rate of settlement).
Figure 5.8. Temperature and precipitation (snow: grey; rain: black) (a), simulated canopy snow storage (black) and observed canopy snow cover fraction (grey) (b), and simulated (black) and observed (grey) snow depth (c) derived from the time-lapse photography images at the Lower Russell (LR) 13m forest from February 20 to March 12th, 2007. The temperature threshold for snow vs. rain was 0.25°C
Figure 5.9. Temperature and precipitation (snow: grey; rain: black) (a), simulated canopy snow storage (black) and observed canopy snow cover fraction (grey) (b), and simulated (black) and observed (grey) snow depth (c) derived from the time-lapse photography images at the Lower Russell (LR) old growth (OG) forest from November 11 to December 05, 2007. The temperature threshold for snow vs. rain was 0.25°C.
5.4 Discussion

Based on the physical understanding of these systems, observations from the field and the time-lapse photography images, a number of processes were identified for further investigation to determine how they affected model outputs. Performance when simulating SWE varied greatly by year and location, with cold content and density generally over-predicted. Even where the model performed well, such as during the accumulation phase for snow depth, there were still periods of over and under-prediction. This was likely due to challenges associated with classifying rain vs. snow when temperatures fluctuated above and below freezing. A general over-prediction of SWE was common at most locations, which may have been due to poor representation of layering within the snowpack that assumed matrix flow of meltwater when observations of preferential flow were prevalent. There were also instances during the spring melt period when depth was either over or under-predicted, suggesting that improvement could be made to the albedo module based on observations of litterfall in forested environments. In addition, observations of loading and unloading suggest complex interactions among energy, snow structure and the physical attributes of the trees. The following sections focus on these model deficiencies and assess through sensitivity analysis when possible if changing individual parameters results in improved simulations.

5.4.1 State of precipitation in maritime climates

Predicting snow vs. rain in maritime climates can be difficult when temperatures constantly fluctuate above and below freezing throughout the winter. In continental environments, temperatures tend to be well below freezing, thus there is greater certainty in differentiating state of precipitation. CRHM utilized air temperature to delineate rain vs. snow. The threshold for rain vs. snow was adjusted during the initial modelling stages as a form of calibration between -1 and +1. Figure 5.10 shows the sensitivity of snow depth to even slight changes in threshold air temperatures (0.5°C). It became apparent that adjusting temperature thresholds would improve model simulations at some times of year, but decrease performance at other times. Using wet bulb temperature may improve the model performance when delineating rain vs. snow because air temperature is not the only factor in determining state of precipitation. It is worth noting, that 2007-
2008 was a comparatively cooler year, especially in January and February where average monthly temperatures were well below freezing, while in 2006-2007 the average temperatures were only slightly below freezing (Figure 5.11). These cooler temperatures led to more certainty in classifying precipitation, especially in January and February of 2008 and is a possible factor why the model simulated snow depths better during the accumulation phase in 2007-2008 than in 2006-2007 at a number of locations. These colder temperatures would also have resulted in the stand pipes freezing more often, which may have resulted in the model under predicting SWE at some locations during certain periods. Undercatch was accounted for through a correction factor based on wind speed and precipitation type (Yang et al., 1998), however these values were not derived on site, thus there is uncertainty in precipitation measurements, which may have also contributed to simulated snow depth and SWE diverging from observations.
Figure 5.10. Model sensitivity to snowfall threshold temperature for the Upper Stephanie (US) clear-cut (CC) plots in (A) 2006-2007; and, (B) 2007-2008. Solid black line is for a snow threshold temperature of +0.5 °C, dashed grey line is for a threshold of 0 °C and the dashed black line represents a snowfall threshold of -0.5 °C. Solid grey line is observed snow depth.
Figure 5.11. Average monthly air temperatures from the Upper Stephanie (US) meteorological station (840m a.s.l) for the snow accumulation and ablation months in 2006-2007 and 2007-2008 at Russell Creek Experimental Watershed.

5.4.2 Canopy processes

Generally, the canopy interception module (Ellis et al., 2010) within CRHM performed well at reproducing canopy dynamics at both the old growth and regenerating plots based on the periods assessed. There were some substantial differences between observations and simulations, especially when temperature fluctuated above and below freezing when differentiating snow from rain would have been difficult.

Regardless of problems associated with delineating rain from snow, observations of loading and unloading of snow suggest that the interception module could be improved. Snow structure depends on temperature and crystal formation, thus in certain conditions, snow can be bonded strongly together and form bridges between branches creating pyramid-like snow complexes that exceed what a single branch could hold. Snow in this state is generally stable and requires increased wind speed and/or temperature to break down the structure and cause mass release. Field observations demonstrated that often within minutes of the sun coming out in forested stands, trees
began to shed snow. This was a highly variable process that depended on the position of the sun in the sky, which was not accounted for in the Canopy module. Another process observed using the time-lapse cameras was that of residual snow remaining on the canopy in the form of ice or snow/ice complexes. This often occurred when rain transitioned to snow followed by a rapid drop in temperatures, literally freezing the snow to the branches. On occasion, it was observed that after mass release of snow from the canopy, residual snow and ice remained until the next snowfall occurred or until temperatures remained above freezing over an extended period. The Canopy module was not structured to account for these processes and will need to be updated for use in coastal watersheds, particularly in old growth and mature forests where interception capacity is highly variable due to the complex structure of the trees. The length of time snow remains in the canopy is important in coastal watersheds as most will melt from the canopy, leading to significant differences in snow depth in openings compared to forests (e.g. Stork et al., 2002). In addition, snow and melt water drip released from the canopy will generally have higher albedo than snow that fell directly to the ground.

The Canopy module within CRHM included a simplistic maximum canopy snow load as described in the methods section. Once the threshold was reached, the model unloaded snow from the canopy. The simplicity was appealing, however, loading and unloading of snow from the canopy is extremely complex and the holding capacity of branches is not only a function of the physical characteristics of the branch, but also the physical characteristics of the snow (Lundberg and Halldin, 2001). The capacity of the forest canopy to store snow varies with species, age of tree which has a bearing on branch thickness and the overall bonding of the snow to itself and to the branch. The importance of storage capacity within the canopy was significant as shown in the results. The sensitivity analysis indicated that the higher the storage capacity, the lower the simulated snow depth, with even relatively small differences having major effects (Figure 5.3). This was also the case in maritime watershed in Oregon, where varying the interception capacity resulted in significant changes in SWE (Storck, 2000; Andreadis et al., 2009). This threshold for maximum snow load is extremely important, especially in light of the difficulty in measuring it in the field. There have been attempts to measure interception
with tree-weighing experiments (Pomeroy et al., 1998b; Storck, 2000; Storck et al., 2002) or by weighing individual branches (Schmidt and Gluns, 1991); however, all of these methods are limited by the number and size of trees that can feasibly be used in the experiment. The use of time-lapse photography proved useful in assessing the interception module’s ability to capture the loading and unloading dynamics, but it could not provide information on the actual mass of snow stored in the canopy. It is clear that additional research effort should be focused on the maximum threshold process to improve representation of its dynamic nature.

5.4.3 Preferential flow in coastal snowpacks

While there was no explicit data collected to measure preferential flow within the snowpack at Russell Creek, a number of observations including highly variable lysimeter outputs, identification of multiple ice layers during collection of snow course data, and observations of rilling on the surface of snowpacks following intense rainfall, indicated that it is an important process that was not accounted for in the simulations. The two-layer snowmelt model utilized in SNOBAL may greatly oversimplify the routing and storage dynamics of liquid water in the snowpack, which is a possible reason why simulations for depth and specifically SWE were poor at some locations. Previous research has shown that matrix flow of melt water and/or rain through the snow profile is rare (Furbish, 1988; Harrington and Bales, 1998; Kattelmann, 1985; Kattelmann and Dozier, 1999; Schneebeli, 1995; Singh et al., 1997; Waldner et al., 2004; Williams et al., 2010). During the snow course measurements, numerous ice lenses were observed in the snowpack. Typical of coastal watersheds, ROS events are extremely common, thus it is possible to have 5 to 10+ ice lenses in the snowpack. Distinct ice layers on low gradient slopes can become barriers and increase storage of water in the snowpack (Harrington and Bales, 1998; Singh et al., 1997), or on steeper slopes act as preferential flow paths that can be routed to micro-topographic depressions and isolated drains in the snowpack (Kattelmann and Dozier 1999). All of the above-noted phenomena were observed at Russell Creek. In addition, minor textural differences can result in lateral preferential flow paths (Kattelmann and Dozier, 1999; Mitterer et al., 2011). Rainwater flow through the snowpack can be several times faster than natural snowmelt (Singh et al., 1997).
fact, Kattlemann and Dozier (1999) found that preferential flow paths allowed water to reach the base of the snowpack up to 30 days earlier than the background or wetting front in the deep relatively warm snowpacks of the Sierra Nevada. Snowmelt lysimeters were installed at all plot locations in our study and none of the data were usable due to the extreme amount of variation in outputs, further suggesting that the heterogeneity of the snowpack caused some lysimeters to have excess melt through preferential pathways, and others to be completely bypassed. Observation of preferential flow at Russell Creek was not limited to within the snowpack, with rilling and minor gullying observed on the snow surface at a number of locations during the year.

The heterogeneity of flow paths within the snowpack can lead to a high degree of variation of outflow or SWE measurements. This complexity has long been recognized, but models describing these processes are still in their infancy. Much of the research into preferential flow paths has been done by the snow avalanche community (e.g. Gustafsson et al., 2004; Mitterer et al., 2011). The primary difficulty with attempting to model preferential flow is the dynamic nature of snow metamorphism which can change multiple times throughout a winter season. The hill-slope hydrology community is still improving its understanding of preferential flow, and they work in a medium that is largely static in comparison to snow. While many have taken a physical approach to modelling preferential flow paths and layer development within the snowpack (Gustafsson et al., 2004; Mitterer et al., 2011), due to its dynamic nature, it may be more appropriate to use a stochastic framework to develop preferential flow models (Furbish, 1988).

Problems with simulating depth and SWE were not limited to the over-simplification of liquid water routing in snowpacks. When observed vs. simulated snow temperatures were compared, it was clear that CRHM underestimated temperatures within the snowpack. For example, observed snow temperature measurements from USCC and USOG plots in 2006-2007, remained within 0.5°C of isothermal condition with the exception of the early season in November and December when snowpacks were low (Figure 5.12), however, in almost every instance, the modelled temperatures from the SNOBAL module were many degrees cooler than the observed values. These colder
temperatures translated into higher simulated cold content, especially after sustained cold periods. Differences in cold content would have contributed to divergence of observed and modelled depths and SWE. Furthermore, if rain was falling on snow with a high simulated cold content, the model would re-freeze the water and store it, increasing modelled SWE, when it is possible that the water passed through the snowpack. Differences in modelled cold content in 2006-07 vs. 2007-08 may be a possible reason why the model performed better at the US 06-07 locations than in 07-08 when comparing SWE simulations and observation (Figure 5.13). The modelled monthly cold content of 2006-2007 was much lower from November through to the end of February than in 2007-2008, thus the model would not re-freeze rain or melt water as often leading to simulated SWE values being much closer to observed values. It is likely that the modelled cold content in 2007-2008 in the lower layer was sufficiently high to refreeze water and store it, resulting in high simulated SWE values. Unfortunately there were no snow temperature observations in 2007-2008 to compare to simulated values.
Figure 5.12. Simulated (black) and observed (grey) snow temperature and cold content from the Upper Stephanie (US) clear-cut (CC) and old growth (OG) plots during the 2006-2007 snow season.
Figure 5.13. Modelled monthly average cold content at the Upper Stephanie (US) clear-cut (CC) locations in 2006-07 and 2007-08.

5.4.4 Litter-fall and albedo

The simplicity of the albedo module used in the analysis is appealing; however, in forested environments it may over-estimate albedo by not accounting for litterfall. Images from the time-lapse photography network showed significant litter deposition throughout the season, especially at the OG and 11/13m forest locations (Figure 5.14). Litter deposition has been documented elsewhere (Hardy et al., 2000; Melloh et al., 2001; Winkler et al., 2010) and attempts have been made to incorporate litter-fall into albedo and snowmelt calculations (Melloh et al., 2002). Unfortunately CRHM does not have a module to account for litter-fall. In the absence of albedo measurements, changes can be made to the threshold values (upper and lower limits) for albedo within the existing framework to better represent potential values in the presence of litter-fall. To account for litter-fall at the forested locations, the minimum albedo was adjusted to 0.5, 0.3 and 0.2 based on potential values as derived in Hardy et al. (2000). While this sensitivity analysis indicated a general increase in melt rate (Figure 5.15), the model still over simulated depth at the forested locations. It is important to note that a reduction in minimum albedo had no effect on modelled depth until the beginning of March when
solar radiation comprised a large part of the energy balance. This suggests that albedo plays less of a role in snowmelt during the early season and general accumulation phase of snowpack development.
Figure 5.14. Images of litter-fall at the Upper Stephanie (A) old growth (OG) and (B) 11m forest in the spring of 2007.

Figure 5.15. Comparison of minimum albedo values to account for litter-fall in Upper Stephanie (US) old growth (OG) forest in 2006-2007. Solid gray line is the observed snow depth; solid dark line is the simulated snow depth for a minimum albedo of 0.5, dashed black line is for a minimum albedo of 0.3 and the grey dashed line is a minimum albedo of 0.2.
5.4.5 **Soil heat flux**

Simulated soil heat flux represented between 10 and 90\% of total modelled energy available for melt depending on plot location and month (Figure 5.16). The role of soil heat flux was somewhat surprising since it is generally ignored in most snow-dominated watersheds based on the assumption of frozen soil. In some months at some locations it was the only net positive flux into the snowpack, usually occurring in November and December when snowpacks were minimal (Figure 5.16 and Figure 5.17). High soil heat fluxes in 2006-2007 during the early season were associated with the model greatly under predicting snow depth, implying that soil heat flux was over-estimated during these periods. In 2006-2007 the LR snowpack was relatively low and soil heat was the dominant energy source for much of the year (Figure 5.16). In contrast, at US in 2007-2008 the relative contribution of soil heat was reduced due to the deeper snow (Figure 5.17). Soil heat was always positive throughout the year, where all other fluxes generally had periods of negative input to the snow. These results support recent research showing soil heat to be a major energy source for total melt in a continental snowpack contributing between 3 and 27\% of total snow fall (Smith, 2011) and another application of SNOBAL in a maritime climate showed soil heat contributed between 8 and 25\% of the total melt energy from 1997 - 2003 (Mazurkiewicz et al., 2008).

Despite the relatively high amount of soil heat flux, the model generally over predicted SWE, usually during the accumulation phase. It is possible that energy derived from the soil was unaccounted for during the simulations based on field observations when snow was seen melting from the bottom of the snowpack even when temperatures were well below freezing and no melt was occurring from the top layers. The simple two layer melt model employed by SNOBAL could not account for this melt because it requires ground heat to be conducted to the active layer i.e. melt cannot come directly from the lower layer. In some instances there was a significant gap between the bottom of the snow and the ground, suggesting energy sources other than conduction were driving melt. This implies that an important process is not accounted for in the model.

Soil temperature measurements during this project were limited to one measurement per location, so it is unlikely the variability of heat fluxes were captured at
each site, both in terms of soil temperature and the conductive property of soils. Increased spatial coverage of soil temperature measurements would help to explain the variability in soil heat flux and possibly improve model performance. Even with minimal soil temperature measurements, observations clearly showed that soil heat flux was an important component of the overall energy balance and further research is warranted to gain a better understanding the role soils play in snowmelt. Because soil properties varied between the plots and soil temperature observations were limited, it was difficult to make comparisons between plots and to say definitively that soil heat flux was more important at one location than another. Because of this, and the fact that removal of forest cover will generally have the greatest affect on above-ground energy inputs to melt snow, comparison of energy fluxes among forest types will be limited to above-ground energy sources.

5.4.6 Above–ground energy sources

Turbulent fluxes are often cited as most important in watersheds frequented by ROS events (Beaudry and Golding, 1983; Coffin and Harr, 1992; Harr, 1981; Harr and Berris, 1983; Marks et al., 1998; van Heesjwick et al., 1996). Our results suggest that on a monthly basis, turbulent fluxes played a minimal role, with net radiation in some months dominating the energy balance. In 2007-2008 net radiation became the dominant energy source as the days grew longer in April and May with the melt season extended due to the deep snow (Figure 5.17). High net radiation was not limited to open locations; at some of the forested plots (e.g. LROG) net radiation came close to or exceeded net radiation at the open locations (Figure 5.16). Longwave radiation emitted from the forest canopy moderated temperature extremes and ensured minimal energy loss on cold clear nights compared to the open location. In addition, fog, mist and cloud were prevalent at all the plot locations, especially at US, resulting in high amounts of long wave radiation that was less sensitive to forest cover attenuation than direct solar radiation. There were instances when net radiation was minimal below forested stands such as in February and March 2007 at USOG (Figure 5.16). This was associated with the coldest sustained temperatures of the study period, resulting in a reduction of net modelled longwave radiation emitted from the forest.
The contribution of turbulent fluxes to monthly melt varied depending on location and year. At the LR plots in November and December 2006 sensible and latent heat were a negative net flux (Figure 5.16), while at the US locations in 2007-2008 they were largely positive and made up the majority of the energy flux in some months (Figure 5.17). The locations of the plots were not particularly windy, especially at LR, resulting in minimal differences in turbulent energy flux among forest types. In general, the open locations had increased rates of evaporation, with greater losses of energy due to turbulent exchange than in forested locations. During specific events, turbulent exchanges may have been much higher in the open, but when the monthly inputs were compared, the relatively stable nature of the forested sites resulted in similar net inputs compared to open areas. If different aged forests in more exposed locations were compared, one would expect to see much larger differences in turbulent fluxes between the open and forested plots. Similar to the results from this research, work at the HJ Andrews in the Oregon Cascades has shown that radiation and ground melt are actually the dominant energy sources in open locations (Mazurkiewicz et al., 2008) on an annual basis. Alternatively the 1996 ROS event that caused major flooding in Oregon, turbulent exchanges accounted for 60 to 90% of the total energy (Marks et al., 1998), suggesting that extreme events are likely to show different relative magnitudes of energy fluxes then the more frequent events described in this study. Size of event and exposure of a watershed to wind and resultant turbulent exchange can have a major effect on the energy fluxes and the differences among forest types.
Figure 5.16. Modelled energy available for melt at Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740 m to 770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests in 2006-2007. Note that energy is only calculated when snow is on the ground. Rn is net radiation (black); L is latent heat (dark grey); H is sensible heat (light grey) and G is ground heat (white). Advective heat was negligible so it is not shown.
Figure 5.17. Modelled energy available for melt at Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740 m to 770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests in 2007-2008. Note that energy is only calculated when snow is on the ground. Rn is net radiation (black); L is latent heat (dark grey); H is sensible heat (light grey) and G is ground heat (white). Advective heat was negligible so it is not shown.

5.4.7 Comparison of model outputs to literature

As described in the introduction, campaigns to compare models in different hydrological regimes, particularly in areas where snow plays a major role in the hydrological cycle, have been completed in recent years (Ellis et al., 2010; Essery et al., 2009; Rutter et al., 2009). These serve as testing grounds to determine the strengths and weaknesses of current models and datasets. The results compare well to other models and other locations where SNOBAL and the modules of CRHM have been applied. While a wide range of NS values occurred, with the lowest values occurring for SWE measurements, many of which indicated the average SWE was better than the modelled SWE, NS values greatly improved for snow depth simulations and during ROS melt periods. These values generally fell within the range of NS, MB and RMSE reported in
the literature for current model comparisons (Ellis et al., 2010; Essery et al., 2009; Rutter et al., 2009). The results show it is more difficult to model SWE and depth at forested than in open locations (Rutter et al., 2009), although there were occasions where NS values were higher at forests than in openings. The SWE simulations were weak compared to work done at the HJ Andrews in the Oregon Cascades, an area receiving multiple ROS events (Mazurkiewicz et al., 2008), however, a systematic calibration procedure was used, which likely accounted for these differences. There was a general improvement in NS when simulated values were compared to twice-daily snow depth observations as opposed to snow course observations taken every 2 to 3 weeks. In addition, correlation coefficients were generally above 0.80 on an annual basis and during the snowpack development phases. This suggests the model performs well at simulating snow accumulation and melt dynamics. The early season modelling deficiencies associated with shallow snowpacks and high ground heat energy fluxes identified by others as model weakness (Ellis et al., 2010; Essery et al., 2009; Mazurkiewicz et al., 2008) were also evident for the analysis done at Russell Creek. Results suggest that SNOBAL and the Canopy module performed well at tracking relative differences of snow accumulation and melt at the different elevations and locations.

Recent work using the Variable Infiltration Capacity (VIC) hydrological model at a maritime mountainous site in Oregon, which has ROS events similar to those at Russell Creek, showed similar behaviour to these results (Andreadis et al., 2009). Even with a seemingly more robust canopy model to simulate mass release and canopy drip, the model still over or under-predicted SWE and depth to a similar magnitude as this study. Annual correlation coefficients were similar to Andreadis et al. (2009) which ranged between 0.81 and 0.99. None of the model applications reviewed in the literature for coastal watersheds had snow temperature observations, so no comparisons to simulated cold content could be done. In fact, snow temperature and cold content comparisons seem largely ignored in assessments of model performance.

5.5 Summary and conclusions

A novel approach to collecting snow depth and interception data using a time-lapse photography network combined with traditional snow course measurement within
regenerating and old growth forests in a watershed frequented by ROS was used to evaluate CRHM and its two primary modules for snowmelt and canopy interception. In general, the magnitude of the simulations were much higher than the observed SWE measurements, with the exception of some locations in 2006 and 2007. The high SWE simulations may have been a result of the model over simulating cold content and the simplistic melt-water routing through the snowpack. Observations and research elsewhere suggest preferential flow can be a significant water delivery mechanism. Even though the model over-predicted SWE, it simulated the general pattern of snow accumulation and melt, and captured the relative differences among the forest cover types.

The high temporal resolution snow depth observations collected via the remote camera network provided an opportunity to assess model performance during melt periods. In most instances model efficiency was greatly improved for snow depth compared to SWE and the model tracked snow ablation well during a selection of ROS melt events. Including variability of snow depth collected during snow course measurements suggest the model captured the variability of snow depth at the research plots in most cases. While snow depth was generally over-predicted, it was not as great as the over prediction associated with SWE. Again this suggested that cold content calculations were too high and this may be resolved through improved modelling of energy inputs.

Soil heat flux was shown to be a significant and in some cases dominant heat flux during some months at certain research plots. In addition, net radiation was a major flux, even in the forested locations. Turbulent energy exchanges were relatively small when compared on a month to month basis, with sensible and latent energy generally being greater at the US locations where wind speeds were higher. The results suggest that it is too simplistic to assume that forests always have less energy available for melt than clear-cut locations and differences will depend on overall exposure to the major energy driver, primarily solar radiation and turbulent fluxes associated with wind.

The model was able to detect the major energy inputs to the system and observed snow ablation tracked closely to simulated values. This indicated that above ground
energy components are captured by the model and that differences between forest stands can be accounted for. It is these energy differences, all else being equal, that will be a major factor in stand level snowmelt recovery. From this modelling exercise, the next step will be to select a series of ROS events of differing magnitude and frequency and determine the relative differences among regenerating and old growth forest stands to assess stand level energy flux recovery.
Chapter 6: Comparison of energy available for melt among different forest stands during rain-on-snow in a coastal watershed

6.1 Introduction

Paired watershed studies have been the dominant source of information pertaining to the effects of forest harvesting on streamflows in regions which experience ROS, however, results and conclusions have been inconsistent (Alila et al., 2010; Alila et al., 2009; Berris and Harr, 1987; Beschta et al., 2000; Christner and Harr, 1982; Harr, 1981; Harr, 1986; Harr and McCorison, 1979; Harr et al., 1975; Jones and Grant, 1996; Jones and Grant, 2001; Jones and Perkins, 2010; Lewis et al., 2010; Thomas and Megahan, 1998). Paired watersheds were recognized early on as “black box” experiments that could not fully explain watershed processes nor shed light on the inconsistencies. Of particular interest was how snowmelt dynamics during ROS were altered after forests were removed. This led to a number of stand level studies comparing forest openings and mature forests which generally showed deeper snowpacks and higher melt rates in clear-cuts (e.g. Beaudry and Golding, 1983; Harr and Berris 1983; Berris and Harr 1987; Coffin and Harr, 1992). Melt water drip of intercepted snow was the primary mechanism leading to lower snow-packs in forests than openings (Beaudry and Golding 1983; Berris and Harr 1987; Storck et al., 2002). Increased melt rates during ROS in forest opening were attributed to high wind speeds resulting in large inputs of turbulent energy (e.g. Harr and Berris 1983; Berris and Harr 1987; Coffin and Harr 1992; van Heeswijk et al., 1996; Marks et al., 1998). By and large it was concluded that rain falling on snow produces more water available for runoff (WAR) and that removal of the forest canopy enhances this process. Some went as far to suggest that not only would the magnitude of melt events increase as a result of forest removal, but they would also become more common (e.g. Harr 1981; Berris and Harr 1987; Coffin and Harr 1992). Although statements linking changes in magnitude with frequency of events were made early on in forest hydrology research, methods of analysis prevented a true description of this non-linear relationship.

Traditionally, paired watershed research chronologically pair (CP) individual events between a control and treatment watershed(s) and use either ANOVA or ANCOVA analysis to determine if significant differences in streamflow are present (e.g. Beschta et al., 2000; Christner and Harr, 1982; Harr, 1986; Harr et al., 1982; Harr and McCorison, 1979; Harr et
al., 1975; Jones and Grant, 1996; Jones and Perkins, 2010; MacDonald and Stednick, 2003; Moore and Wondzell, 2005; Thomas and Megahan, 1998; Troendle and Stednick, 1999; Troendle et al., 2001). Many studies using CP analysis concluded that forest harvesting only impacts small to moderate sized events (Beschta et al., 2000; MacDonald and Stednick, 2003; Moore and Wondzell, 2005; Thomas and Megahan, 1998; Troendle and Stednick, 1999; Troendle et al., 2001). This has led to the perception that forest harvesting does not affect high magnitude/low frequency events which are associated with flooding (e.g. Bathurst et al., 2011). Recent work has challenged this perception using a fundamentally different analysis technique known as frequency pairing (FP), showing that forest harvesting affects the magnitude and frequency of all streamflows, including flood events (Alila et al., 2009; Green and Alila, 2012; Kuraš et al., 2012). These studies also suggest that as the magnitude of events increase, so does the effect on frequency. Chronological pairing focuses on differences in magnitude for individual events, which decouples the intrinsic relationship to frequency and masks the effects forest harvesting has on streamflow (Alila et al., 2009). The use of ANOVA and/or ANCOVA analysis further compounds the problems associated with CP. A main assumption of ANOVA and ANCOVA is equal variance – in situations where variance is greater in a pre or post-harvest watershed, the data must be transformed. If variability is reduced through transformation, a fundamental characteristic of the dataset is suppressed. Because of this, ANOVA and/or ANCOVA may suggest that two population mean/medians are equal, when in fact their variance shows significant differences in the frequency of certain events.

While there has been disagreement as to the effects of forest harvesting on streamflows, particularly for “large” or “extreme events”, it is widely recognized that removal of forests will affect streamflows and as trees regenerate, hydrologic recovery of annual water yield and peak stream flows will occur. Post-harvest hydrologic recovery is primarily a function of how regenerating forests interact with precipitation and the sources of energy that melt snow. Transpiration can also play a significant role in watersheds with deep soils, through delayed response to precipitation or snowmelt processes. Basin-wide hydrologic recovery studies have focused on paired watersheds. Some studies show recovery taking as little as 10 years (Thomas and Megahan, 1998) and up to at least 30 years (Jones,
2000) in maritime watersheds. In a western Oregon watershed, water yield recovery took 31 years (Stednick, 2008). A snow-dominated watershed in Colorado showed no signs of water yield recovery 31 years after harvest (Troendle and King, 1985).

The role of hydrological recovery is a main criticism of paired watershed studies identified in both CP (Beschta et al., 2000; Jones, 1996; Jones, 2000; Thomas and Megahan, 1998) and FP analysis (Alila et al., 2010; Alila et al., 2009; Lewis et al., 2010). Regenerating forests introduce non-stationarity to paired watershed studies making it difficult to identify changes to streamflow. Removing this confounding variable can be done through physically-based hydrological modelling (e.g. Kuraś et al., 2012) or through statistical approaches (e.g. Alila et al., 2009). Alila et al. (2009) used FP-based analysis and the same dataset as Troendle and King (1985), albeit with a longer record, and found no recovery 41 years after harvest. In a snow-dominated watershed in central British Columbia, recovery could not be detected 25 years following forest harvest (Green and Alila, 2012). These studies suggest that recovery may take longer than previously thought and any formal assessment of recovery must include changes to the frequency across events of all sizes (Alila et al., 2009; Green and Alila, 2012; Kuraś et al., 2012). In addition, recovery of frequency and recovery of magnitude may occur at different rates. We are not aware of any studies that have assessed watershed recovery using a physically-based hydrological model to perform frequency based analysis in ROS environments. Before this can be done, we need to improve our understanding of stand level recovery by examining the linkage between magnitude and frequency.

Research on stand level hydrological recovery has focused on snow accumulation and melt in both continental and maritime watersheds where studies compared SWE and melt rates among different stand types (Boon, 2007; Coffin and Harr, 1992; Hardy and Hansen-Bristow, 1990; Winkler et al., 2005), and others identified stand attributes (e.g. height, canopy closure, LAI, aspect, species etc) that best explained differences (Buttle et al., 2005; Hudson, 2000a; Hudson, 2000b; Hudson, 2003; Hudson and Horel, 2007; Talbot and Plamondon, 2002; Talbot et al., 2006; Winkler and Moore, 2006). Most research occurred over 2 or 3 winters and compared gross differences in snow course measurements completed in 1 to 4 week intervals. In British Columbia stand level research in areas frequented by
ROS used weekly and bi-weekly snow course data from different aged stands to assess recovery (Hudson, 2000a; Hudson, 2000b). Further research incorporated rainfall data into the snow course measurements, however, high resolution melt data were not collected during ROS (Hudson, 2003; Hudson and Horel, 2007). Rain-on-snow events generally occur over one or two days making it difficult to capture with snow course measurements. Furthermore, snow course comparison studies, with minimal data points and a coarse representation of energy inputs are not conducive to frequency based analysis. The only example found in the literature with high resolution ROS data collected via snowmelt lysimeters used CP to compare ROS events, and while an attempt was made to determine how changes in mean melt due to forest harvesting would affect the frequency distribution of ROS, they stopped short of assessing changes to frequency among stand types (Coffin and Harr, 1992). None of the research reviewed on stand level hydrological research assessed changes to frequency or recovery of this variable. In addition, no studies within ROS environments used energy balance calculations to assess recovery.

Past research in ROS environments has alluded to changes in frequency at the stand level as a result of forest harvesting, with Berris and Harr (1987) stating that “larger inputs of water would become more frequent simply because of the removal of forest vegetation”. Other research suggested that return intervals for 24hr water delivery to the soil profile during ROS from a clear-cut would be reduced by half compared to a mature forest (Coffin and Harr, 1992). While Coffin and Harr (1992) clearly identified the link between increases in melt rate from ROS in forest openings and potential changes to return intervals compared to forested locations, they did not invoke changes in variability. Because of this, changes to frequency may be even greater then suggested, especially if forest removal increased variability of water outputs. Despite this, the logic behind increased snowmelt in forest openings affecting return intervals of water delivered to the soil was correct and supports Alila et al.’s (2009) contention that comparing frequency distributions is critical when discussing the impacts of forest harvesting on streamflow. In maritime watersheds, multiple ROS events of varying magnitude occur through a single winter season, thus with a comparatively shorter dataset, changes to frequency of events can be assessed.
With recent research showing forest harvesting having an effect on the magnitude and frequency of the largest events on record (Alila et al., 2009; Green and Alila, 2012; Kuraś et al., 2012), limiting rates of forest removal within watersheds may be even more important than previously thought. Hydrological recovery will be a key component to setting harvest limits within watersheds, of which snow energy flux recovery will play a critical role. The primary objective of this chapter is to determine how a regenerating forests energy flux during ROS recovers using frequency based analysis. Stand level snowmelt recovery will be assessed using the modelling results from the previous chapter for a selection of ROS events. Analysis will focus primarily on energy flux recovery among the different forest types and compare mean, variability and frequency distributions. Changes to mean and variability for a short term dataset will be placed in context of a longer climate station record to determine how forest removal and subsequent recovery can affect the return interval of ROS events. This will form the basis of a discussion on stand level hydrological recovery in watersheds frequented by ROS. In addition, results will be compared to past studies and suggestions provided for application to forest management and future research.

6.2 Methods

6.2.1 ROS event selection

Meteorological data from 2005 to 2008 were used to select a series of ROS events from the two meteorological stations located at 490m and 840m a.s.l. A ROS event was defined as having a minimum 10mm rain falling on snow with no more than 6 hrs of consecutive rain free periods. An event was only classified as ROS if the time-lapse photography could verify snow was on the ground. Time-lapse photography from 2005-2006 was limited due to power supply issues that were later rectified in 2006-2008, limiting the number of ROS events captured from 2005-2006. Precipitation was delineated based on the threshold values used during the model validation from the previous chapter. Total rainfall for the duration of the ROS event and maximum 24hr rainfall intensity were calculated. We selected 24hr intensity because it is often associated with peak flows and landslides in coastal watersheds (Guthrie et al., 2009).
6.2.2 Snowmelt modelling

The Cold Regions Hydrological Model (CRHM) (Pomeroy et al., 2007) was used to predict snowmelt from 2005 to 2008. Primary modules used in the CRHM platform included SNOBAL (Marks et al., 1999; Marks et al., 1998) and a physically-based canopy interception model (Ellis et al., 2010). The previous chapter provides detailed information about the structure of CRHM. Above ground energy balance components for sensible and latent heat \((H+Lv)\) (turbulent fluxes) and net radiation \((Rn)\) were used to compare maximum energy \((Qn)\) available for melt among the forest stands at LR and US. Advection was ignored due to its negligible role in the total energy balance. While ground heat flux was found to be a significant factor in the modelling from the previous chapter, due to the high amount of variability associated with soil properties and heat fluxes, it was not used in the energy balance calculation in this chapter. In addition, changes in observed depth referred to as \(\Delta z_{obs}\) (as a proxy for snowmelt) captured by the time-lapse photography network were compared to see if the results were similar to the modelled energy components. The primary objective during the analysis was to determine how forest cover removal and regeneration affected energy available for melt during specific ROS events. However, the dataset was relatively short, so there was a need to reduce confounding variables due to differing antecedent conditions among stands such as cold content of the snow-pack, and routing of melt water through different depths. To account for this, it was assumed all snow-packs were ripe, thus all \(Qn\) resulted directly in snowmelt producing a variable known as maximum potential melt \((M_{max})\). Water available for runoff \((WAR)\) was calculated as through-fall rain, assuming no canopy melt and added to the \(M_{max}\). This oversimplifies the melt and interception processes during the ROS events, however, it served to place \(M_{max}\) in context of the total rainfall. Results focused specifically on above ground energy available for snowmelt during ROS because this is the component that forest cover has the biggest impact on.

6.2.3 Statistical comparisons

Two main durations were examined within this study, maximum 24hr energy input and total energy inputs over the duration of the ROS event. These were selected to contrast maximum intensity which is related to both landslide initiation and peak streamflow and the
total energy over the duration of an event which effects the duration of streamflow events and can be associated with channel change. For the remainder of the chapter, “24hr” refers to maximum input of energy, melt or rainfall over a 24hr period, and “total duration” refers to total energy, melt and/or rainfall over the entire length of an event.

With the limitations associated with ANOVA and ANCOVA and importance of comparing frequency distributions, multiple types of statistical analysis were performed to compare energy available for snowmelt among the forest stands. The mean and frequency distributions of energy available for melt were compared using non-parametric approaches to maintain the variability within the dataset and provide further insight into the effects of forest management and regeneration on magnitude and frequency.

6.2.3.1 Comparisons of means

To remove the assumption of equal variance required by ANOVA and ANCOVA analysis, Hotelling’s T-Test (Hotelling, 1934) and Yao’s Approximation were used (Algina et al., 1991; Yao, 1965). All statistical comparisons were done using S-Plus statistical software (Insightful, 2007). Forest stand comparisons were limited to those within their respective elevation ranges (LR: 520-560m a.s.l. and US: 740-770m a.s.l.). Total energy ($Q_n$), turbulent energy ($H+Lv$), net radiation ($R_n$), maximum potential melt ($M_{max}$), Water Available for Runoff ($WAR$) and observed snow depth ablation ($\Delta z_{obs}$) were compared for the total duration and 24hr intensities. While $\Delta z_{obs}$ is not as informative as observed changes to SWE, it was used as a primary comparison because of the high temporal resolution of snow depth observations during ROS which the snow course measurements could not capture. $\Delta z_{obs}$ is a useful variable because it is related to energy fluxes, however it is also a function of densification of snow due to natural settling rates and direct input of water through rainfall, thus observed differences in $\Delta z_{obs}$ may indicate either greater or smaller energy fluxes and resultant melt than actually occurred.

6.2.3.2 Distribution comparisons

One of the limitations of ANOVA/ANCOVA analysis is that results cannot be extended beyond the mean values to discuss the effects of forest management on larger magnitude and/or lower frequency events. Hotelling’s T-Test and Yao’s approximation
allow for the comparison of stands with unequal variance, however, they do not fully describe the differences in variability and distribution.

To account for potential differences in the frequency of energy inputs during ROS, cumulative distribution function (CDF) plots were used to compare the distribution of energy inputs among forests during ROS. The distributions were compared using the two sample Kolmogorov-Smirnov (K-S) test. This test compares the maximum distance between distributions to determine if they are statistically different. This test focuses specifically on differences between the portions of the distribution that are furthest apart which is a function of both shape (variation) and location (mean). In addition to the KS test, CDF’s were visually assessed to provide insight into where distributions differ and changes to frequency may be greatest.

6.2.3.3 Statistical significance and physical reality

A distinguishing characteristic of hydrological science is relatively small sample sizes utilized to describe inherently complex systems where variables cannot be controlled. A key tenant of using statistics to compare treatments or populations is an arbitrarily defined p-value to reject a null hypothesis. It is routine to ignore differences or trends if the p-value is >0.05, or if confidence intervals overlap. However, this is often done without a physical justification and very real processes can be occurring that if ignored may result in fundamental flaws when interpreting outcomes. An over reliance on pre-determined limits of significance is well described by (Elliott and Brook, 2007):

“Yet statistical methods that we would recognize today are less than 100 years old. This raised a question: What were Hooke, Linnaeus, Cuvier and Darwin doing before the development of the p-value?”

Even though a small sample size predisposes studies to poor statistical power, if patterns are evident they should not be discounted if supported by a physical understanding of the system. Furthermore, even statistically insignificant differences can have significant impacts on the return intervals of the largest events (Alila et al., 2009; Lewis et al., 2010).
With the above in mind, results will be presented with statistical significance. However, interpretation of results will utilize differences among mean, variability and frequency regardless of p-value. Depending on where one resides on the use of statistical significance, this can be considered a strength or weakness of the chapter; however, ignoring a probable difference or similarity based simply on a p-value cannot be justified when working outside of a cause and effect study without firm controls on confounding variables. The transition points between separating populations are often the most important because they can represent a shift or change in a physical process that was previously unknown, thus the use of threshold p-values may inhibit our ability to solve some of the most pressing problems by masking where changes occur.

6.2.3.4 Energy and melt recovery

Recovery was assessed for the 4m, 11m and 13m forests with CC representing no recovery and OG forests considered 100% recovered. Recovery will be based on \(Q_n\), \(M_{\text{max}}\), \(\text{WAR}\) and \(\Delta z_{\text{obs}}\) as these are most closely related to snowmelt. A recovery index (HR\%) was calculated based on the mean and standard deviations of the energy and melt variables using the following equation:

\[
HR\% = 100 - \left[ \left( \frac{X_{RF} - X_{OG}}{X_{CC} - X_{OG}} \right) \times 100 \right] 
\]  

where \(X\) is either the mean or standard deviation of the melt or energy variable from the regenerating forest (RF), old growth (OG) or clear-cut (CC). Because mean and standard deviation are directly related to changes in distribution, these will be used as a proxy to compare recovery of the frequency distributions. Since others have suggested that as magnitude increases (Alila et al., 2009; Green and Alila, 2012; Kuraś et al., 2012), potential changes to return intervals also increase, recovery of frequency paired events by their cumulative frequency distribution were plotted for \(M_{\text{max}}\) and \(\Delta z_{\text{obs}}\) to see if any patterns were present.
6.2.3.5 Additional analysis

Partial duration frequency analysis was completed for 24hr maximum rainfall to place the 2005-2008 dataset in context of the entire 16 year record at the 840m climate station using the Cunnane equation:

\[ Tr = \frac{1}{(m - 0.4) \cdot \frac{N}{(N + 0.2)} n} \]  \hspace{1cm} (26)

where \( Tr \) is the return interval for an event of given magnitude, \( m \) is the rank of observation, \( N \) is the number of observations and \( n \) is the length of record in years. Partial duration analysis was selected because many of the largest 24hr rain events occurred in the same year and mean annual maximum frequency analysis would eliminate these events and bias the resulting distribution. The distribution for 24hr rain intensities was assumed to be different than the distribution for ROS melt events over the past 16 years because no data specific to the magnitude of rain and snowmelt for ROS events was available. However, the 24hr rain distribution served as an indicator to determine how sensitive the system was to increases in water input from snowmelt during ROS. The LR (490m) MET station had a similar record length to the US (840m) MET station, however, there were significant gaps in the data outside of the study period that make partial duration frequency analysis difficult, so comparisons were only done for the US plots.

6.3 Results

6.3.1 ROS events

Between 2005 and 2008 a total of 95 ROS events were observed at Russell Creek. There were 54 ROS events recorded at LR and 41 ROS events at US. The largest events occurred from November to January at LR in 2006/07 and 2007/08 (Figure 6.1), while US had all but one of its large (>80mm) events in 2006-2007 (Figure 6.2). The lack of large ROS US in 2007-2008 was a result of cooler temperatures and precipitation falling as snow. In general, large events (>80mm) falling on relatively shallow snowpacks (< 400mm) occurred at LR (Figure 6.1), while US had large rain events (>80mm) falling on snow depths
ranging from 200mm to just over 1200mm (Figure 6.2). Average air temperatures for the
events ranged from just above freezing to a maximum event average of 4.5°C. Average wind
speeds for the total duration of events were below 2m/s at LR (Figure 6.3). The largest rain
events had average air temperatures below 1.5°C and average wind speeds below 1m/s at LR.
Temperatures were slightly lower at US, however, wind speeds were higher for events larger
than 80mm (Figure 6.4). Maximum temperatures of 6 to 10°C were recorded, generally
occurring later in the winter when days were longer (data not shown). In general, there were
a number of events with greater than 80mm of rain falling on snow, however, average
temperatures and wind speeds were relatively low.

Partial duration analysis indicates that the 2005-2008 dataset captured the entire
distribution of 24hr max rainfall intensities (Figure 6.5), including the largest rain event on
record of 112/mm in 24hrs in November 2006 that was associated with more than 100
landslides in the general vicinity (Guthrie et al., 2009)
Figure 6.1. Monthly occurrence and observed maximum snow depth in the open during ROS events from 2005-2008 at Lower Russell (LR) (490m a.s.l.).

Figure 6.2. Monthly occurrence and observed maximum snow depth in the open during ROS events from 2005-2008 at Upper Stephanie (US) (840m a.s.l.).
Figure 6.3. Comparisons of average wind-speed, average temperature and total rain during ROS at Lower Russell (LR) (490 m a.s.l.) from 2005-2008.

Figure 6.4. Comparisons of average wind-speed, average temperature and total rain during ROS at Upper Stephanie (US) (840 m a.s.l.) from 2005-2008.
6.3.2 Energy flux comparison

Energy available for melt during ROS from 2005-2008 generally showed that the CC and 4m forests had consistently higher inputs of total energy at both the LR and US locations (Figure 6.6). Radiation was the dominant form of energy for all forest types at the LR plots, with net turbulent fluxes making up minor components. The LR plots were in a relatively sheltered location from winds so turbulent fluxes were low. In contrast, the US plots were in a windier location and turbulent fluxes made up a greater proportion of the overall energy balance during ROS, just as they did over both winter seasons as described in the previous chapter. The US 11m and OG plots had very low totals for $Rn$, with the OG showing a net cumulative loss for the ROS events. Net turbulent fluxes were similar between all the forest types at the US plots. All plots at both elevations show a decrease in total energy available for melt during ROS as trees become taller and canopies denser.
Examining energy available for melt among the forest stands on a monthly basis provided additional insight in how time of year affected energy fluxes during ROS. The US plots (Figure 6.7) had snow on the ground for longer periods of time than LR, thus March and April energy fluxes during ROS were greater than at LR (Figure 6.6). Energy tended to decrease as forests became taller, however, some months, such as March where US13m (Figure 6.7C) and USOG (Figure 6.7D) had near equal and even greater amounts of energy available for melt than USCC (Figure 6.7A) and US4m (Figure 6.7B). Negative energy fluxes at the LR plots occurred in January and February which were generally the coolest months at the LR elevation. The highest total energy fluxes during ROS at the US sites occurred in March 2007, which was a function of the relatively high number of ROS events that occurred, along with the longer solar days and higher radiation inputs.

Figure 6.6. Summation of energy available for melt at the Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740 m to 770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests during ROS from 2005-2008. Rn is net radiation, Lv is latent heat and H is sensible heat.
Figure 6.7. Monthly summation of energy available for melt at Lower Russell (LR) (520-560 m a.s.l.) during ROS for the clear-cut (CC) (A.), 4m (B.), 13m (C.) and old growth (OG) (D.) forests. Rn is net radiation, Lv is latent heat and H is sensible heat.

Figure 6.8. Monthly summation of energy available for melt at Upper Stephanie (US) (740-770 m a.s.l.) during ROS for the clear-cut (CC) (A.), 4m (B.), 11m (C.) and old growth (OG) (D.) forests. Rn is net radiation, Lv is latent heat and H is sensible heat. NA signifies no available simulations.
6.3.3 Comparisons of mean and variability

As described in the previous section, turbulent energy fluxes made up a small component of total energy at LR, however, CC and 4m still had higher mean and variability than 13m and OG (Figure 6.9). The LR4m stand had the highest mean and greatest variability for mean $H+Lv$ of all plots for both the 24hr and total duration. Turbulent exchanges were much higher at US, with OG having the smallest mean values and least variability. The 24hr $H+Lv$ did not have any negative fluxes and variability was quite similar, however, when turbulent exchanges were accounted for over the duration of the event, the variability at US CC and 4m were considerably higher than 11m and OG, with some highly negative events. This is likely due to the increased rates of evaporation that may have occurred at the start and end of ROS events, as well as the greater changes in temperature in the open vs. closed canopy locations. Significant differences were found among all stands at US for the 24hr $H+Lv$ (except 4m vs. 11m), however, when the total duration was examined, the only significant difference in mean was found between 4m and OG (Table 6.1). This did not occur at the LR plots where 24hr and total duration significant differences remained the same.

Box plots for $Rn$ were somewhat more consistent in their differences among forest stands, with CC and 4m having the highest mean values for the 24hr and total duration periods at both LR and US (Figure 6.9). The US CC and 4m plots show the most distinct differences from 11m and OG of all comparisons for $Rn$. At both LR and US, the CC plots consistently has the largest maximum $Rn$ inputs, with the OG generally having the smallest of the peak $Rn$ inputs and the 4m and 11m and 13m falling in between. The LRCC forest consistently had significantly greater mean inputs of $Rn$ than all other LR plots for both total duration and 24hr, while all other plots had statistically equal means (Table 6.1). At US, CC and 4m had equal mean $Rn$, as did 11m and OG.

The $Qn$ box plots followed a similar pattern to $Rn$, with mean $Qn$ from the CC plots being consistently higher than the OG plots at both the LR and US locations (Figure 6.9). Mean $Qn$ at the 4m plots were also consistently higher than the OG plots, with differences greater at US than LR. All LR plots at both the 24hr and total duration were significantly different, with the exception of 13m and OG (Table 6.1). Means were significantly different
for $Q_n$ at all US plots for both the 24hr and total duration plots with the exceptions 4m and CC plots and 11m and OG. Mean $Q_n$ was significantly different between US 11m and OG over 24hr, but not for total duration.

$M_{max}$ eliminated all the negative $Q_n$ fluxes from the distribution and reduced variability, which in theory affected the frequency distribution. Despite this, relative differences in mean were similar to $Q_n$. The CC locations consistently had the highest maximum $M_{max}$, with the OG plots having the lowest maximums for total duration and 24hr inputs at US and LR (Figure 6.10). In general, the box plots indicated that 11m, 13m and the OG plots had the lowest variability for $M_{max}$, with the CC and 4m plots having the highest. $M_{max}$ for all total durations were significantly different at the LR plots with the exception of 4m and OG for maximum 24hr input and 13m and OG for both the 24hr and total duration periods (Table 6.1). The US CC and 4m plots had equal mean $M_{max}$ over both time periods, and significantly different mean values for all other comparisons.

The larger the ROS event, the greater the proportion of rain in $WAR$, with differences in mean being a function of rain interception and energy available to melt snow. As with the other variables, $WAR$ showed a clear decrease in mean values from CC through to OG at all elevations for both 24hr and total duration (Figure 6.10). As expected, the addition of rainfall resulted in similar variability between all plots. Comparison of mean $WAR$ produced the least number of comparisons with significant differences, with all means being equal at LR except for CC and OG, and CC and 13m (Table 6.1). Significant differences in mean $WAR$ were only found between CC and OG, 4m and 11m and 4m and OG at US.

The $\Delta z_{obs}$ variable provided comparisons using observed data. Changes to depth at the CC and 4m plots had greater means and higher variability than 11m, 13m, and OG at both LR and US (Figure 6.10). The 11m and 13m forests had the smallest variability of all locations. Variability also appears to be comparatively greater for $\Delta z_{obs}$ than the modelled results. Significant differences in mean $\Delta z_{obs}$ were found among all the LR plots with the exception CC and 4m and 13m and OG (Table 6.1). At US, only 11m and OG had equal means, with all others being significantly different.

Table 6.1 displays some patterns among comparisons. The mean values for all variables at LR 13m and OG were statistically equal. Another consistent pattern is that mean
values were significantly different between LR CC and OG for all variables. At US, CC and 13m were significantly different for all variables except $H+Lv$ for 24hr and total duration of 
$WAR$. The US locations showed similar patterns for mean differences with CC and OG, and 4m and OG having significant differences for all variables expect for one. All other 
comparisons have equal means or significant differences depending on variable.

In general, the LR and US CC and 4m forests had greater mean and variability of 
energy inputs during ROS than the 11m, 13m and OG forests, with the observed $\Delta z_{obs}$ showing similar differences in mean and variability. In many instances both mean and
variability were similar between the 4m and CC locations. The 11m, 13m and OG forests had similar means across all variables, with variability being generally higher at 11m and 13m than the OG forests. Both mean and variability was generally greater over the total duration of events than for the peak 24hr energy inputs. Negative energy fluxes were a substantial component at all locations, especially over the total duration. The highest potential $M_{max}$ rates occurred in the CC locations.
Figure 6.9. Maximum 24hr and total duration energy sources during ROS events from 2005-2008 at Lower Russell (LR) (520-560m a.s.l) and Upper Stephanie (US) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests.
Figure 6.10. Maximum potential melt ($M_{\text{max}}$), water available for runoff ($WAR$) and observed snow depth ablation ($\Delta z_{\text{obs}}$) during ROS events from 2005-2008 at the Lower Russell (LR) (520-560m a.s.l) and Upper Stephanie (US) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests.
Table 6.1. Comparison* of mean energy sources, maximum potential melt ($M_{\text{max}}$), water available for runoff (WAR) and observed snow depth ablation ($\Delta z_{\text{obs}}$) during ROS events from 2005-2008 at Lower Russell (520-560m a.s.l) and Upper Stephanie (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Period</th>
<th>LOWER RUSSELL</th>
<th>UPPER STEPHANIE</th>
</tr>
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<tr>
<td></td>
<td></td>
<td>CC vs. 4m</td>
<td>CC vs. 13m</td>
</tr>
<tr>
<td>$H + L_v$ (MJ)</td>
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<td>0.04 0.19</td>
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<tr>
<td></td>
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<td>0.08 0.26</td>
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<td></td>
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<td>1.28 0.20</td>
</tr>
<tr>
<td>$Q_n$ (MJ)</td>
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<td>0.90 0.33</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.87 1.27</td>
<td>1.37 0.40</td>
</tr>
<tr>
<td>$M_{\text{max}}$ (mm)</td>
<td>24hr</td>
<td>1.42 2.37</td>
<td>2.12 0.95</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2.06 3.60</td>
<td>3.47 1.54</td>
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<tr>
<td>WAR (mm)</td>
<td>24hr</td>
<td>4.40 7.20</td>
<td>9.95 2.80</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5.10 8.10</td>
<td>11.30 3.00</td>
</tr>
<tr>
<td>$\Delta z_{\text{obs}}$ (mm)</td>
<td>24hr</td>
<td>-2.40 45.20</td>
<td>42.40 47.60</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-3.70 42.90</td>
<td>40.60 46.60</td>
</tr>
</tbody>
</table>

*Hotelling’s T-Test and Yao’s approximation was used to identify significant differences among the forest types. **Bold** values are significantly different at $P<0.05$. 

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6.3.4 Frequency comparisons

Distinct shifts in distributions were evident at all locations, especially between the CC and OG forests. When simply comparing mean turbulent exchanges across all events, differences were small, and could even be considered negligible to total melt, however, the CDF’s indicated substantial differences in turbulent inputs for some events (Figure 6.11). Even though turbulent exchanges were not very high at the LR plots, a departure of the higher magnitude events occurred at LR4m, with bigger turbulent inputs occurring more frequently. This corresponds to LR4m having a significantly different distribution than all other plots over the 24hr period (Table 6.2). The CDF’s at US show a much higher input of turbulent energy than LR with a wide range of values for both 24hr and total duration (Figure 6.11). The 24hr CDFs have a fairly consistent separation between US CC and OG across the entire distribution, with 4m showing no distinct departures from the other CDFs. At US the 24hr maximum intensities had no negative net turbulent exchanges, while over the total duration negative turbulent fluxes make up a large proportion of the CC and 4m CDFs. As the magnitude of turbulent energy increased for the 24hr periods, the CDF’s converged for all forest types at US, however, for the largest event, there was substantial spread among all plots. This same pattern was evident over the total duration among US plots for events with positive turbulent fluxes. The US CC and 4m plots had equal distributions and both were significantly different from the OG forest for both 24hr and total duration periods (Table 6.2). None of the other comparisons had consistently significant differences between 24hr and total duration.

The CC locations had the highest $Rn$ inputs across the entire distribution at all locations and periods, with instances of US4m having the greatest inputs over portions of the distribution, specifically the small to moderate sized inputs (Figure 6.11). At LRCC for both the 24hr and total duration periods, $Rn$ was almost always positive, while the other LR sites had a substantial number of events with negative $Rn$, with OG showing the greatest magnitude of negative 24hr $Rn$. For events with positive $Rn$, LR OG and 13m had similar distributions and overall showed no significant differences over the 24h or total duration (Table 6.2). The LRCC $Rn$ distributions were significantly different than all other forest types. The LR 4m forest had an equal distribution to 13m and only showed a significant difference from OG for the total duration of the event. There was a clear separation of $Rn$
CDFs for CC and 4m from 11m and OG at US. The US 11m and OG plots had a similar
distribution for the small to moderate sized events, but as magnitude increased, the larger
events began to diverge, especially for the 24hr period. Even with this separation, results
suggest the distributions were not significantly different at both the 24hr and total duration
for 11m and OG (Table 6.2). The \( Rn \) distributions for US CC and 4m were significantly
different than 11m and OG during all periods.

With \( Rn \) being the dominant energy source at LR, the \( Qn \) CDF was similar; however,
there were some differences (Figure 6.11). The main difference was LR CC and 4m were
significantly different than 13m and OG and the 4m CDF was equal to CC for the total
duration. Another difference was LR 13m and OG had significantly different \( Qn \) for the total
duration. With both positive and negative turbulent fluxes being greater at US, more
substantial changes to the \( Qn \) CDFs were observed compared to the LR plots, although
relative differences among plots appeared to remain the same. The US CC and 4m plots had
significantly different distributions than US 11m and OG (Table 6.2). The 11m and OG
forests also had significantly different distributions at \( p<0.10 \). The US CC and the 4m plots
had equal distributions.

The results for the \( Qn \) distribution comparisons showed the exact same significant
differences and equalities for \( M_{max} \) (Table 6.2). Because all negative energy sources were
eliminated, the CDF plots for \( M_{max} \) had numerous ROS events where no melt occurred
(Figure 6.12). At US, the higher the magnitude, the greater the difference among plots,
especially between CC and OG. In addition, as \( M_{max} \) became larger at US, the 11m
distribution moved away from OG towards 4m and CC, especially during the 24hr period.

Similar to comparisons of mean \( WAR \), the distributions became similar and relative
differences were decreased. Even so, a clear separation of CC and OG can be seen for all
durations and locations (Figure 6.12). As the events became larger and total rain made up a
greater proportion of \( WAR \), all the distributions converged, although the highest magnitude
events still appeared to have different frequencies, especially the CC vs. OG forests. At both
LR and US, the CC and OG plots were significantly different for the 24hr and total duration
(Table 6.2). In addition, LR13m and US11m had equal distributions to their comparative OG
forests. At LR, the 4m and 13m distributions were equal for both 24hr and total duration, as
was the 4m and 11m at US. As a whole, differences among forest types using WAR were far less distinct.

Differences in distribution among forest plots for $\Delta z_{obs}$ were generally more pronounced than the other variables. The LR CC and 4m tracked closely for most of the distribution and diverged for the largest $\Delta z_{obs}$. The highest magnitude events had substantial differences in the distribution, especially between the CC and OG plots (Figure 6.12). Similar to the US locations, initially the 4m forest $\Delta z_{obs}$ were closer to the OG and 11m distributions, however, as the magnitude of $\Delta z_{obs}$ increased the distribution diverged and tracked closer CC. The US 11m and OG plots had similar CDFs, with some portions of the 11m CDF having smaller magnitude changes in depth at comparable frequencies. The $\Delta z_{obs}$ CDFs confirmed many of the similarities and differences identified by the other energy and melt parameters, with LR CC and 4m being significantly different than 13m and OG, and US CC and 4m being significantly different than 11m and OG. At both LR and US, the CC and 4m plots had equal distributions. The $\Delta z_{obs}$ distributions were equal between LR 13m and OG for the total duration, but not at 24hrs. When the US 11m and OG $\Delta z_{obs}$ distributions were compared, the 24hr was equal, but the total duration was significantly different.

All distributions for all variables were significantly different between CC and OG plots at both LR and US over 24hr and total duration periods. This was consistent with the significant differences found when comparing means and suggests that the mean, frequency and by default variability for CC plots are significantly different/greater than the OG forests. All distributions were significantly different between LR CC and 13m, except for the turbulent exchanges over the total duration, and suggest CC had significantly higher mean, variability and more frequent large events than 13m. All other forest types at LR had varying numbers of equal and significantly different distributions suggesting that some level of recovery occurred between the LR 4m and 13m forests relative to the OG. Distributions for all parameters at US CC and 4m were equal, with the exception of the 24hr $Rn$, suggesting that both mean, frequency and variability were the same for most parameters, except mean $\Delta z_{obs}$. By and large all US CC variables had significantly different distributions than 11m, with the exception of WAR and 24hr turbulent exchanges, with the same being true for the 4m and 11m. Other than WAR and 24hr turbulent exchanges, all US CC and 4m energy and
melt variables had significantly different distributions and means than the 11m and OG forests, suggesting little to no recovery has occurred between these plots.

To summarize, the CC and 4m locations tended to have the greatest energy inputs across the entire distribution at both LR and US. Relative differences among stands were greater over the total duration than the 24hr periods. There was also greater separation of distributions among the forest types at the US plots compared to the LR locations. The LR13m forest had a similar CDF to LROG for all variables, while the US11m CDF diverged from USOG as the events became larger and less frequent. Results from this section generally matched comparisons of mean and standard deviation, but provided a clearer representation of where differences were greatest.
Figure 6.11. Cumulative distribution function (CDF) distributions for latent and sensible ($H+Lv$), net radiation ($Rn$) and total energy ($Qn$) during ROS events from 2005-2008 at Lower Russell (LR) (520-560m a.s.l) and Upper Stephanie (US) (740-770m a.s.l) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests.
Figure 6.12. Cumulative distribution function (CDF) for maximum potential melt ($M_{\text{max}}$), water available for runoff (WAR) and observed snow depth ablation ($\Delta z_{\text{obs}}$) during ROS events from 2005-2008 at Lower Russell (LR) (520-560m a.s.l) and Upper Stephanie (US) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests.
Table 6.2. Distribution comparison* of energy sources, maximum potential melt ($M_{\text{max}}$), water available for runoff ($WAR$) and observed snow depth ablation ($\Delta z_{\text{obs}}$) during ROS events from 2005-2008 at Lower Russell (520-560m a.s.l) and Upper Stephanie (740-770m a.s.l) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests.

| Variable | Period | LOWER RUSSELL | | | UPPER STEPHANIE | | |
|----------|--------|---------------|-----------------|-----------------|---------------|-----------------|-----------------|-----------------|---------------|-----------------|-----------------|-----------------|---------------|
|          |        | CC vs. 4m     | CC vs. 13m      | CC vs. OG       | 4m vs. 13m     | 4m vs. OG       | 13m vs. OG       | CC vs. 4m       | CC vs. 11m     | CC vs. OG       | 4m vs. 11m      | 4m vs. OG       | 11m vs. OG      |
| $H+L_v$  (MJ) | 24hr  | 0.34          | 0.22            | 0.21            | 0.35           | 0.36           | 0.10            | 0.17            | 0.14           | 0.32           | 0.20            | 0.33            | 0.28            |
|          | Dur    | 0.17          | 0.14            | 0.32            | 0.20           | 0.33           | 0.28           | 0.10            | 0.30           | 0.30           | 0.29            | 0.32            | 0.09            |
| $R_n$ (MJ) | 24hr  | 0.32          | 0.40            | 0.35            | 0.12           | 0.12           | 0.24           | 0.41           | 0.54           | 0.52           | 0.66            | 0.66            | 0.20            |
|          | Dur    | 0.31          | 0.42            | 0.42            | 0.16           | 0.18           | 0.15           | 0.14           | 0.50           | 0.60           | 0.60            | 0.67            | 0.16            |
| $Q_n$ (MJ) | 24hr  | 0.29          | 0.39            | 0.34            | 0.17           | 0.19           | 0.12           | 0.15           | 0.359          | 0.48           | 0.45            | 0.58            | 0.223           |
|          | Dur    | 0.15          | 0.36            | 0.48            | 0.45           | 0.58           | 0.22           | 0.09           | 0.33           | 0.37           | 0.36            | 0.39            | 0.19            |
| $M_{\text{max}}$ (mm) | 24hr  | 0.29          | 0.39            | 0.33            | 0.17           | 0.18           | 0.10           | 0.15           | 0.36           | 0.50           | 0.45            | 0.58            | 0.22            |
|          | Dur    | 0.15          | 0.36            | 0.50            | 0.45           | 0.58           | 0.22           | 0.08           | 0.31           | 0.39           | 0.36            | 0.39            | 0.19            |
| $WAR$ (mm) | 24hr  | 0.23          | 0.29            | 0.39            | 0.15           | 0.27           | 0.19           | 0.14           | 0.25           | 0.36           | 0.18            | 0.32            | 0.21            |
|          | Dur    | 0.14          | 0.24            | 0.29            | 0.14           | 0.21           | 0.16           | 0.14           | 0.25           | 0.35           | 0.20            | 0.31            | 0.23            |
| $\Delta z_{\text{obs}}$ (mm) | 24hr  | 0.14          | 0.46            | 0.45            | 0.56           | 0.50           | 0.210          | 0.22           | 0.51           | 0.50           | 0.37            | 0.31            | 0.16            |
|          | Dur    | 0.22          | 0.51            | 0.50            | 0.37           | 0.31           | 0.16           | 0.27           | 0.47           | 0.43           | 0.33            | 0.51            | 0.23            |

*Kolmogorov-Smirnov used to compare distributions among forest types. Values correspond to the maximum distance (D) between the two distributions. **Bold** values are significantly different at $P<0.05$ and *italicized* are significantly different at $p<0.10$. 

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6.3.5  Stand level energy flux and snowmelt recovery

Recovery of regenerating forests depended on variable, statistical metric, location and period (total duration vs. 24hr). The LR4m forest showed signs of recovery for mean and standard deviation of \( Q_n \) and \( M_{max} \) over the 24hr (Table 6.3) and total duration (Table 6.4). The recovery of mean \( Q_n \) and \( M_{max} \) were less similar over the 24hr period than the duration at LR4m. The LR4m forest showed signs of over recovery for standard deviation for \( Q_n \) over 24hrs, while \( M_{max} \) standard deviation recovery was similar to mean recovery. Recovery was also observed for mean and standard deviation of WAR, with over recovery in some instances at LR4m. Recovery of standard deviation for all LR4m variables was lower over the total duration than over 24hrs. In contrast to \( Q_n \), \( M_{max} \) and WAR, the observed \( \Delta z_{obs} \) at LR4m for both 24hr and total duration showed mostly negative recoveries (i.e. greater than CC) for mean and standard deviation.

The LR13m stand had the highest recovery of all forest types showing signs of over recovery of 24hr mean \( M_{max} \), with partial or full recovery of mean \( Q_n \), WAR and \( \Delta z_{obs} \) (Table 6.3). Standard deviation was over recovered at LR13m for \( Q_n \) and \( M_{max} \) and partially recovered for WAR and \( \Delta z_{obs} \). When recovery was assessed for the 13m forest over the total duration of events, all stands were fully or slightly over recovered for both mean and standard deviation across \( Q_n \), \( M_{max} \), WAR and \( \Delta z_{obs} \) (Table 6.4).

At US4m no signs of recovery were evident for mean over the 24hr period except for \( \Delta z_{obs} \) (Table 6.3). Standard deviation showed signs of recovery at US4m for \( \Delta z_{obs} \), \( Q_n \), \( M_{max} \) and WAR, with recovery generally below the LR locations. Over the total duration of ROS events at US4m, there were no signs of recovery for mean or standard deviation of \( Q_n \), \( M_{max} \) or WAR, however, \( \Delta z_{obs} \) showed signs of recovery (Table 6.4). Over the 24hr period, partial to full recovery occurred at US11m for mean \( Q_n \), \( M_{max} \) and \( \Delta z_{obs} \). Standard deviation showed signs of recovery for \( Q_n \), \( M_{max} \), WAR and \( \Delta z_{obs} \) and were much lower than recovery at LR13m despite similar LAI’s. For the total duration of ROS events, US11m showed signs of partial or full recovery for mean and standard deviation of all variables other than WAR. Mean \( Q_n \) and \( \Delta z_{obs} \) were fully recovered, while \( M_{max} \) showed partial recovery. Standard deviation at US11m was partially recovered for \( Q_n \) and \( M_{max} \), whereas \( \Delta z_{obs} \) was over
recovered. In general, recovery was lower across all variables and durations at US compared to LR.

The above assessment of recovery focused on mean and standard deviation for all events within each forest type. An assessment of recovery for frequency paired events showed some forest types had a reduction in recovery as events became larger and less frequent. At plots with similar means and/or standard deviations, there was minimal effect on the size of event and its overall recovery. The LR13m forest showed a slight increase in $M_{\text{max}}$ recovery as events became larger for both the 24hr (Figure 6.13) and total duration periods (Figure 6.14), with the lowest frequency events ranging between 75 and 160% recovered (values over 100% are considered “over-recovered”). Variability of recovery increased as events became more frequent for $M_{\text{max}}$. While mean and standard deviation indicated LR4m was 88 and 63% recovered for $M_{\text{max}}$ over 24hrs, there was a significant reduction in recovery as events became less frequent and larger in magnitude with many of the larger events between 40 and 60% recovered. The relationship between magnitude and recovery over the total duration was weaker than the 24hr, however, recovery was much lower with many events showing 30% or lower recovery, compared to 59 and 48% for the mean and standard deviation.

When recovery based on frequency of events were plotted for $\Delta z_{\text{obs}}$ at LR (Figure 6.13 and 6.14), there were weak negative relationships between frequency and recovery for both the 24hr and total duration periods, with the 13m plots showing 100% and greater recovery. $\Delta z_{\text{obs}}$ for LR4m had much lower recovery than $M_{\text{max}}$ indicated, with most values showing negative recovery (i.e. higher than the CC).

There were relatively strong negative relationships at US between frequency and recovery for $M_{\text{max}}$ at the 11m forest with some of the largest events being approximately 40% recovered compared to 80 to 100% for the most frequent events over 24hrs (Figure 6.13). This same relationship was evident over the total duration at US11m, however, it was weaker with the largest events being between 60 and 100% recovered (Figure 6.14). The US4m stand showed a small decrease in recovery as frequency decreased, with recovery slightly lower over the total duration.
Recovery for $\Delta z_{obs}$ had a slight increase as events grew larger at US11m for the 24hr and total duration with the least frequent events generally showing over recovery (Figure 6.13 and Figure 6.14). The US4m forest also had a weak negative relationship between recovery and frequency of event with recovery being much higher than for $M_{max}$. This may be a function of the considerably smaller sample size of $\Delta z_{obs}$ captured due to camera malfunction and burial of the camera due to deep snow.
Table 6.3. Recovery of peak 24hr total energy ($Q_n$), maximum potential melt ($M_{max}$), water available for runoff (WAR) and observed snow depth ablation ($\Delta z_{obs}$) for the Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests for ROS events from 2005-2008.

<table>
<thead>
<tr>
<th>Site</th>
<th>N</th>
<th>$Q_n$ (MJ)</th>
<th>$M_{max}$ (mm)</th>
<th>WAR (mm)</th>
<th>$\Delta z_{obs}$ (mm)</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>$\mu$</td>
<td>$\mu_%$</td>
<td>$\sigma$</td>
<td>$\sigma_%$</td>
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<tr>
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<td>0</td>
</tr>
<tr>
<td>US4m</td>
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<td>-11</td>
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<td>48</td>
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<tr>
<td>US11m</td>
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<td>65</td>
<td>1.66</td>
<td>34</td>
</tr>
<tr>
<td>USOG</td>
<td>41</td>
<td>0.89</td>
<td>100</td>
<td>1.11</td>
<td>100</td>
</tr>
</tbody>
</table>

Table 6.4. Recovery of total duration total energy ($Q_n$), maximum potential melt ($M_{max}$), water available for runoff (WAR) and observed snow depth ablation ($\Delta z_{obs}$) for the Lower Russell (LR) (520-560m a.s.l.) and Upper Stephanie (US) (740-770m a.s.l.) clear-cut (CC), 4m, 11m, 13m and old growth (OG) forests for ROS events from 2005-2008.

<table>
<thead>
<tr>
<th>Site</th>
<th>N</th>
<th>$Q_n$ (MJ)</th>
<th>$M_{max}$ (mm)</th>
<th>WAR (mm)</th>
<th>$\Delta z_{obs}$ (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
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<td>$\mu_%$</td>
<td>$\sigma$</td>
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<tr>
<td>USOG</td>
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<td>0.37</td>
<td>100</td>
<td>2.00</td>
<td>100</td>
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</tbody>
</table>
Figure 6.13. Relationship between cumulative frequency and recovery for 24hr maximum potential melt ($M_{\text{max}}$) and observed snow depth ablation ($\Delta z_{\text{obs}}$) at the Lower Russell (LR) 4m, and 13m forests and the Upper Stephanie (US) 4m, and 11m forests. Lines represent trend in data.
Figure 6.14. Relationship between cumulative frequency and recovery for total duration maximum potential melt ($M_{\text{max}}$) and observed snow depth ablation ($\Delta z_{\text{obs}}$) at the Lower Russell (LR) 4m, and 13m forests and the Upper Stephanie (US) 4m, 11m forests. Lines represent trend in data.
6.4 Discussion

6.4.1 Energy balance contributions

Much of the work at the stand level has focused on individual ROS events of interest, based purely on data availability (Beaudry and Golding, 1983; Berris and Harr, 1987; Harr, 1981; Harr and Berris, 1983) or on opportunistic extreme events that occurred over a specific long term research watershed (e.g. Marks et al., 1998). Research has focused primarily on the differences between clear-cut and old growth or mature forests, with turbulent fluxes being identified as the primary generator of melt during ROS (Beaudry and Golding, 1983; Berris and Harr, 1987; Harr, 1981; Harr and Berris, 1983). The results of past research have created the perception that wind is essentially what drives snowmelt during ROS and forest openings will produce more melt than forested stands largely due to higher turbulent fluxes. It also indirectly suggests that other energy fluxes are a minor component to melt. More recent research has been able to focus on a greater variety of ROS events due to longer data records (Mazurkiewicz et al., 2008), and similar to these results found that radiation fluxes were the dominant form when energy was summed for the entire season or series of events. However, this may be somewhat misleading as the results show fluxes can transition from negative through to positive during ROS. Summing these fluxes over a monthly or annual basis may make it appear as they play a small role, when in fact certain fluxes may be important during specific periods. Other research also found that it is rare, at least during study periods, to have high winds that would generally result in high melt rates (Berris and Harr, 1987; Storck, 2000; van Heesjwick et al., 1996). This may be a result of research plot location within watersheds, but it is worth noting that wind speeds are not necessarily high during all ROS events, and when this is the case, differences in melt rates between forested and clear-cut locations may be physically insignificant. What became apparent when examining the distributions and box plots were the number of events that had very little or even negative fluxes during ROS events, for both $R_n$ and turbulent energy. It was clear from the results that not all ROS are equal, and specific antecedent conditions and available energy sources must be present for high melt rates to occur (e.g. Harr, 1981; Perkins and Jones, 2008; Jones and Perkins, 2010). Simply put, it is not that amount of rain that drives snowmelt, it is the amount of energy that accompanies it.
Even though there were more recorded ROS events at the LR forest types, in general, the total net energy during ROS for all events captured from 2005-2008 were greater at the US forest types. This was a function of increased turbulent exchanges at the US forest plots, due to generally higher wind speeds. In addition, the prevalence of cloud/fog at this elevation was much greater than the lower elevations, due to cooler temperatures and lower dew point temperatures, resulting in increased condensation and higher latent heat exchange. Another factor leading to higher net energy fluxes at the US sites was that snow was on the ground for longer periods of time, into the late spring when days were generally longer and warmer, thus $R_n$ accumulated to a greater degree. Differences based on elevation and location within a watershed highlight the importance of location when examining the effects of forest harvesting and subsequent regeneration on energy balance contribution for snowmelt during rain events. Certain portions of watersheds will respond differently to ROS events and differences among forest types will depend on location (Harr, 1981). Similar behaviour based on watershed position was observed at the HJ Andrews research forest in Western Oregon, with both aspect and exposure to prevailing winds being major factors in melt during ROS (Mazurkiewicz et al., 2008).

### 6.4.2 Melt during ROS in clear-cut and old growth forests

The previous section focused on net cumulative differences among forest types and location within Russell Creek and provided some generalizations. When it comes to flooding and its relation to forest removal, we are most interested in how ROS augments the magnitude and frequency of events at both the stand/hillslope and watershed scale. The results clearly showed mean differences in CC and OG locations were significant and persistent across all variables. If only the mean differences were compared, it may have been concluded that because melt rates differed by only 2 to 6mm that they were physically insignificant. This was clearly not the case as revealed by the CDF plots, where differences in magnitude increased as events became less frequent. The CC and OG forests had clear separation between the frequency distributions with no sign of convergence for the largest melt events. These results show that large energy inputs during ROS occur more frequently in CC than OG forests, and while they cannot be directly scaled to the watershed, it stands to reason that changes to magnitude and frequency of snowmelt as a result of forest harvest will translate to changes in streamflows. Results from this stand level research are similar to
those reported at the watershed scale using frequency based analysis (Alila et al., 2009; Green and Alila, 2012; Kuraś et al., 2012) and illustrates that relying on CP analysis and ANOVA/ANOCOVA would have overlooked a fundamental characteristic of the ROS dataset. Results also support previous work at the stand level which suggested that the frequency of melt events would increase after forests were harvested (Berris and Harr, 1987; Christner and Harr, 1982; Coffin and Harr, 1992; Harr, 1986).

Convergence of the frequency distributions for WAR occurred as events became larger and would seem to contradict the other variables which generally showed higher magnitude events as more frequent in CC than OG forests. The primary reason for this convergence was that large rain events were not accompanied by high energy inputs during the study period. Some have attributed convergence of high magnitude events when comparing harvested to un-harvested watersheds to a reduction in the ratio of melt to rain as total rain increases (e.g. Coffin and Harr, 1992; Jones, 2000). This logic is based on chronological pairing and does not hold when using frequency based analysis provided the dataset captures the entire spectrum of events. If the ROS events observed in this study had high amounts of energy associated with them, the divergence that occurred as events became larger for snowmelt would have transferred to the WAR distributions, although not to the same extent as melt alone.

The $\Delta z_{obs}$ showed the largest relative differences for the CC and OG plots compared to the other variables. This can be partially explained by the fact $\Delta z_{obs}$ were observed values and included all variables associated with melt, whereas the modelled outputs were a simplification of reality. The $\Delta z_{obs}$ included ground heat, as well as components of canopy melt that were not included in the energy flux comparisons. The $M_{max}$ appeared to greatly underestimate melt compared to the $\Delta z_{obs}$ observations, however, there were a number of factors which effect changes in depth not fully accounted for in the model. Beside not accounting for ground heat flux there where situations where the model failed to account for the high initial settling rates of cold, fresh snow prior to initiation of melt. Differences in density would also have contributed to the greater differentiation in mean and variability between the open (CC/4m) and forested stands (11m, 13m, and OG) with fresh snow on the ground under a forest generally having higher density due to its interaction with the canopy prior to mass release. There were many instances during the image analyse from Chapter 4
where mass release was observed, but no changes in depth were recorded because the lighter snow on the ground was displaced or compacted by the denser snow from the canopy.

While the results indicated that clear-cut locations generally had greater net energy fluxes, they also showed that during certain events forested sites may produce more melt than open locations. This has been documented elsewhere in both ROS (Beaudry and Golding, 1983; Berris and Harr, 1987; Coffin and Harr, 1992) and snow-dominated environments (Winkler et al., 2005). The high negative turbulent fluxes for certain events in open areas at US over the total duration of events could result in increased cold storage within the snowpack. Combine this with loss of net radiation during cold clear nights, and much deeper snow-packs and situations could arise where more energy is required to initiate melt than is available during a ROS event, thus short duration ROS may produce minimal melt from open locations. Contrast this against shallow snowpacks in forested locations without negative turbulent fluxes leading into ROS and long wave radiation emitted from forested stands. Combined with high soil heat fluxes, which would be conducted much more quickly through the snow profile of a shallow snowpack under a forest, and snow below a forest canopy may melt more readily than in an opening. Because trees hold a finite amount of snow, it has been argued that melt water drip only contributes a small proportion to the overall melt and/or rainfall and has a minimal effect on the overall magnitude of events (Beaudry and Golding, 1983; Berris and Harr, 1987; Harr and Berris, 1983). Results from the previous chapter indicate that snow does not remain in the canopy for long once temperatures rise, and research elsewhere indicated up to 72% will fall as melt water drip (Storck et al., 2002). The time-lapse photography showed maximum snow storage up to 27mm and if 70% was released as melt water drip, an additional 19mm of water could be added to an event, certainly enough to cause a shift in the frequency distribution towards the clear-cut locations. Others have described release of water from snow-packs under forested stands not only due to a lower cold content, but also from snow melting from the canopy priming the snowpack by bringing it close to saturation resulting in rapid runoff with any additional melt or rainfall (Beaudry and Golding, 1983; Berris and Harr, 1987). All of the above processes when combined could result in forested plots producing significantly more melt than clear-cut locations.
None of the above scenarios were accounted for in our analysis and these results should be viewed in this light. As mentioned earlier, by assuming snowpacks were ripe when calculating $M_{max}$, the variability of all datasets were reduced, which would have altered both the mean, and frequency distributions. Again this may have affected the overall differences detected between forests, however, the differences in mean, standard deviation and the frequency distributions using observed snow depth ($\Delta z_{obs}$) were similar, adding confidence to the results. The results also matched the physical understanding of snow accumulation and melt and provided a means to assess when forest harvesting had the greatest potential to impact melt rates and streamflows.

6.4.3 **ROS, forest harvesting and return intervals**

Long term data for return intervals of WAR which would include interception and snowmelt properties were not available. Because of this, it was not possible in an absolute sense to assess how forest harvesting affected return intervals of ROS outputs. However, the short term dataset clearly showed how forest removal affects the mean, variability and frequency of energy available for melt during ROS. Because return intervals are related to changes in both mean and variability, a relatively simple analysis was performed to illustrate that increases in melt during ROS can have significant impacts on return intervals. To determine how an addition of snowmelt to a 24hr rain event would affect the return intervals based on the 24hr partial duration analysis, a change to the mean was applied based on the stand level results for $M_{max}$. Coffin and Harr (1992) used a similar approach, but rather than applying a change in the mean, they increased the 24hr rainfall distribution by a percentage based on proportion of melt in total WAR measured using a CP approach. This type of analysis has numerous assumptions, however, it provides an indication of how sensitive the watershed may be to increases in 24hr maximum water input.

Changes to return intervals were only assessed for the US forests because they were in close proximity to the 840m climate station with the 16 year record. Averages of mean $M_{max}$ were added across the entire dataset for 24hr rainfall intensity at US CC, 4m, 11m and OG. There was not a significant relationship between magnitude of rain and $M_{max}$ or $Qn$, which was somewhat surprising, but can be explained partially by the fact that none of the large magnitude rain events were accompanied by high wind speeds and/or air temperature.
This implies that the probability of the largest calculated $M_{\text{max}}$ was equal no matter how much rain fell. Based on this logic, we also took the maximum $M_{\text{max}}$ from each plot and added it to the 24hr rainfall partial duration series as above.

Adding the relatively small mean values for $M_{\text{max}}$ (3 to 8mm), resulted in substantial shifts of the return intervals (Figure 6.15A). What is revealing was the shift in return interval increased the higher the intensity. For example, a 90mm rain event occurred approximately once every 1.5 years and when melt was added from the CC, the return interval shifted to once every 0.95 years. When we look at the largest 24hr rain event in the watershed (112mm), corresponding to a return interval of once every 23.3 years, this same amount in the CC would occur every 3.8 years. Even the relatively small amount of melt (3mm) from the OG increased return intervals for the largest event to once every approximately 9 years. The change in return interval was even more dramatic when we applied the maximum potential melt rates (Figure 6.15B), where the clear-cut would produce 112mm once every 1.5 years.

It is probable that given a long enough record or simply a change in location to a windier portion of the watershed, differences between CC, 4m, 11m, 13m and OG forests would have been much greater at Russell Creek. Because of the comparatively small difference in mean from the dataset, a sensitivity analysis was done using the 24hr rainfall partial duration frequency curve by increasing melt compared to the OG by 5%, 10% and 20%, all of which are physically plausible increases due to snowmelt (Figure 6.16). As above, there was a dramatic shift in the return intervals, with a 23.3 year return interval occurring once every 3.89 years with only a 5% increase in melt. A further 10 and 20% increase reduced the return intervals to 2.5 and 1.5 years. This equates to an additional 10 to 30mm of melt for the largest rainfall amounts, which is probable in an exposed windy location even if temperatures are only a few degrees above freezing.

In the above example only increases in mean were applied, however, in most cases variability of energy available for melt increased in the CC compared to the OG forest, with standard deviations at the 4m, 11m and 13m forests generally falling in between. An increase in variability will steepen the slope of a CDF, resulting in comparable magnitudes becoming more frequent, thus if included in the above comparisons, changes to return
intervals could have been even greater. Because the distribution of total water delivery to the soil profile for ROS events was unknown, it was assumed to be the same as the 24 rainfall intensity; however, it is likely steeper across the entire range due to the higher probability of substantial melt events occurring during intense ROS. Regardless of the limitations described, the analysis clearly showed the importance of changes in mean (and variability) beyond simply magnitude of water input. Differences in mean and variability and how this relates to frequency must be included in any discussion of hydrological recovery and/or the effect of forest harvesting on snowmelt and streamflow.
Figure 6.15. Changes to 24hr rainfall intensity return interval (Tr) at Upper Stephanie (US) when mean (A) and maximum (B) potential melt ($M_{max}$) from the clear-cut (CC), 4m, 11m and old growth (OG) forests are added to the 24hr maximum rainfall intensity.
Figure 6.16. Changes to return interval ($T_r$) as a result of a proportional increase of melt during ROS in the Upper Stephanie (US) clear-cut (CC) compared to the old growth (OG) forest.
6.4.4 Recovery

Despite calls for research to determine how regenerating forests affect snow accumulation and melt in ROS environments (Berris and Harr, 1987; Coffin and Harr, 1992), surprisingly little work has been done on this topic. This may be a result of the general perception that forest harvesting does not affect high magnitude streamflows and/or the inherent difficulties with capturing enough ROS events in environments which are operationally challenging to work in (e.g. Coffin and Harr, 1992). This project attempted to remedy this through an innovative study design but also fell victim to some of the challenges of working in these environments. Despite this, the results clearly showed significant differences between forested and clear-cut locations. The previous section indicated that changes to return intervals increase as magnitude of events becomes larger, even with relatively small increases in melt. The results indicate that recovery has occurred for the regenerating forests, with some parameters being fully recovered and others showing partial to no recovery. It is important to put these results in context of other research examining recovery in both snow and ROS watersheds at the stand level.

Much of the work related to snowmelt hydrological recovery at the stand level has occurred in continental watersheds where spring snowmelt is the primary driver of peak flow events, thus recovery comparisons are based on peak SWE and melt rates during the freshet period (e.g. Buttle et al., 2005; Hardy and Bristow, 1990; Winkler et al., 2005). Even though melt in these environments is generally associated with solar radiation, they should be compared to Russell Creek, especially with the results showing the importance of radiation during ROS. Using the recovery equation (24) from the methods section data were used from studies across North America to calculate recovery rates for peak SWE and melt rates where applicable (Table 6.5). There were a number of studies suggesting no recovery in short stands and in some situations snowmelt was actually enhanced, similar to the 4m forests in this study (Buttle et al., 2005; Hardy and Bristow, 1990). In all studies, as trees became taller, peak SWE and snowmelt recovery increased. It is clear from Table 6.5 that recovery depends on factors other than simply height and will depend on location and the “type” of recovery being monitored. The recovery rates for mean and standard deviation from this study generally fell within the reported ranges in both
continental and ROS environments. However, none of the studies in continental or ROS environments assessed how the frequency of melt events at the stand level were altered due to forest removal nor how the frequency of energy inputs recovered.

When frequency paired events were compared by rates of recovery using data from this research, it became obvious that in stands where the mean and/or standard deviation were not equal, recovery depended on size of event, and as events became less frequent and increased in size, recovery decreased - something that has not been documented in the literature. Recovery based on frequency of events also differed over the total duration compared to 24hr peak intensity, with the 24hr input showing much lower rates of recovery the lower the frequency of the event. This highlights the importance of assessing recovery based on the size of the event or the process of interest, especially if many different factors cause channel or watershed disturbance. For example, landslide hazard may increase after a threshold for 24 hour peak rainfall intensity is passed, whereas channel forming events may be a function of flood duration rather than the instantaneous peak streamflow.

The inverse relationship between recovery and size of event can be further illustrated by adding the mean differences of $M_{max}$ from US to the 24hr rainfall partial duration plot (Figure 6.15). For example, for events up to 90mm, recovery based on return intervals ranged from 60 to 82%. After this threshold, return interval recovery began to decrease with a 100mm event in an OG forest being 2.7 years, 2.1 in the 11m forest and 1.63 in the CC, representing 48% recovery. The largest event in the OG forest was 115mm occurring every 23.3 years, with comparable events every 8.75 and 6 years in the 11m and CC, indicating the 11m stand was only 15% recovered. Even with the relatively small differences in mean, there was a clear decrease in recovery as magnitude increased that was similar to the frequency paired recovery rates based on the CDFs. We recognize these recovery rates are not conclusive because of the numerous assumptions with the methodology, however, they show that rates of recovery may change as events become larger.

An important component of stand level recovery not fully addressed in the results section is the effect of slope, aspect and other watershed characteristics. The aim of the
study design was to control for slope and assess recovery rates within two elevation ranges in the transient snow zone with an assumption that differences in temperature due to elevation would result in distinct antecedent snow conditions. The sites were not selected based on their exposure to turbulent energy, but rather to ensure a range of regenerating forests could be monitored in close proximity to a clear-cut and old growth forests at a similar elevation. The plot locations inadvertently resulted in differences in exposure to wind, with the US locations having higher turbulent exchanges. As already discussed, position in a watershed is important to rates of recovery. This is analogous to research in both ROS (Harr, 1981; Mazurkiewicz et al., 2008; Perkins and Jones, 2008) and snow-dominated catchments (e.g. Talbot et al., 2006; Winkler et al., 2006) where location played a major role in snowmelt dynamics and differences among forest types. Our locations were compared based primarily on height and canopy closure, however, others have shown neither to be a strong predictor of snow accumulation and melt, suggesting additional factors need to be accounted for when assessing recovery (Hardy and Bristow, 1990; Buttle et al., 2005; Winkler and Moore, 2006; Talbot et al., 2006).

For all intensive purposes LR13m showed near to full recovery for all attributes, with results suggesting over-recovery as events became larger and less frequent. This recovery must be placed in context of position in the watershed, the range of events that occurred during the study and the physical characteristics of the individual trees and the stand itself. The LR13m forest was located in a sheltered portion of the watershed. If it was in an exposed location, melt rates would likely have been higher due in part to the physical structure of the trees. Relatively short heights of the trees would make them more susceptible to increased wind speeds impacting the stand 20 to 30 meters closer to the snow surface compared to an old growth forest. Combined with weaker branch/trunk strength and shallow roots, and wind speeds could overwhelm the trees ability to attenuate turbulent energy. The weaker branches also cause mass release to occur sooner resulting in generally deeper snowpacks that can sometimes persist longer, increasing the probability that rain will fall while snow is on the ground.

Another important component to consider when basing recovery on tree height is to include information related to canopy closure. This becomes especially important when snow-packs become moderately to very deep and persist well into the growing
season. Much of the parkland areas in coastal watersheds are maintained by deep snowpacks and short growing seasons, thus trees are found naturally clumped together with openings in between. This same process can occur in newly harvested areas in snow accumulation zones. It is not uncommon at higher elevations at Russell Creek to have 3 meter snowpacks in new clear-cuts compared to 1.5 to 2m in old growth forests. This change in microclimate can severely hamper regeneration and result in pseudo-parkland forests with gaps interspersed among groups of regenerating trees. Often snow depths will not decrease to pre-harvest levels in high snow areas until the structure (e.g. branch diameter) of the regenerating forests start to match that of old growth trees, which takes much longer to recover than both height and canopy density. Until the regenerating forests completely fill in the gaps and/or stand heights became tall enough to reduce turbulent exchanges at the snow surface, significant recovery of melt rates will not occur. It is analogous to the minimal rates of recovery observed for the 4m stands with significant gaps. If the gaps are equal to or greater than the heights of the surrounding trees, total accumulation will be enhanced with possibly much higher turbulent exchanges depending on exposure to prevailing winds, both of which will delay stand level hydrological recovery.
Table 6.5. Summary of stand level hydrological recovery studies for both snow and rain-on-snow-dominated watersheds.

<table>
<thead>
<tr>
<th>Paper</th>
<th>Area/ Process/Mature Tree Height (m) (Spp*)</th>
<th>Regen. Tree Height (m); (Spp*)</th>
<th>Recovery Parameter</th>
<th>Recovery (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hardy and Bristow, 1990*</td>
<td>Montana/Snow/ 18.8 – 26.6 (Fd)</td>
<td>0.5-4 (Pl)</td>
<td>Peak SWE</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td></td>
<td>10.4 – 14.0 (Pl)</td>
<td>Mean Daily Melt</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hudson, 2000a; 2001</td>
<td>British Columbia – Coast/Snow-ROS/32</td>
<td>4 (Hw; Cw, Hm)</td>
<td>Mean Daily Melt</td>
<td>53</td>
</tr>
<tr>
<td></td>
<td>(Hw, Hm, Yc)</td>
<td>8 (Hw; Cw, Hm)</td>
<td></td>
<td>75</td>
</tr>
<tr>
<td></td>
<td></td>
<td>14 (Hw; Cw, Hm)</td>
<td></td>
<td>90</td>
</tr>
<tr>
<td>Talbot and Plamondon, 2002</td>
<td>Quebec/Snow/ 15 (Bb)</td>
<td>4 (Bb)</td>
<td>Mean Daily Melt</td>
<td>50</td>
</tr>
<tr>
<td></td>
<td></td>
<td>10 (Bb)</td>
<td></td>
<td>100</td>
</tr>
<tr>
<td>Buttle et al., 2005*</td>
<td>Ontario/ Snow/ 16.2 (Ba; Sw; Ep)</td>
<td>4-6 (Sb, Ba, Sw)</td>
<td>Peak SWE</td>
<td>80</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mean Daily Melt</td>
<td>50</td>
</tr>
<tr>
<td>Winkler et al.*, 2005</td>
<td>British Columbia – Interior/ 23 (Se, Pl, B1)</td>
<td>4.5 (Pl)</td>
<td>Peak SWE</td>
<td>43</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mean Daily Melt</td>
<td>29</td>
</tr>
<tr>
<td></td>
<td></td>
<td>6.4 – thinned (Pl)</td>
<td></td>
<td>43</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mean Daily Melt</td>
<td>13</td>
</tr>
<tr>
<td>Hudson and Horel, 2007</td>
<td>BC Coast/ROS/ 30+ (Hw, Hm, Yc)</td>
<td>4</td>
<td>Mean Daily Melt</td>
<td>55</td>
</tr>
<tr>
<td></td>
<td></td>
<td>11</td>
<td></td>
<td>75</td>
</tr>
<tr>
<td></td>
<td></td>
<td>13</td>
<td></td>
<td>84</td>
</tr>
</tbody>
</table>

*Pl – Pinus contorta (lodgepole pine); Fd – Pseudotsuga menziesii (Douglas fir); Hw – Tsuga heterophylla (Western hemlock); Hm – Tsuga mertensiana (mountain hemlock); Yc – Chamaecyparis nootkatensis (yellow cedar); Cw – Thuja plicata (Western red cedar); Bb – Abies balsimfera (balsam fir).

* Recovery derived using data using equation 24 from methods section (Hudson, 2000a).
6.4.5 From the stand to the watershed

It is important to compare recovery at the stand level to other research completed at the watershed scale in ROS environments. Of particular interest is research in ROS watersheds in Oregon that suggested recovery would occur in 10 years following post-harvest using CP analysis (Thomas and Megahan, 1998). This would seem to contradict our results that show forests up to 11m tall not being fully recovered, which even in a productive land base would take longer than 10 years. Other research using CP analysis suggested that events with return intervals greater than 5 years would not be affected by forest cover removal (Beschta et al., 2000). Again our results seem to contradict this assertion showing a decrease in rates of recovery as the frequency of events decrease and magnitude increases. Scaling stand level results to the watershed can be difficult, however observations from Russell Creek suggest it is probable that stand level results will translate directly to streamflow when certain conditions are met.

Previous research at Russell Creek from a hillslope in its lower reaches found preferential flow within the soil profile dominated during storm events with rates multiple magnitudes higher than matrix flow and among the highest documented in the world (Anderson et al., 2009). This can lead to rapid response to water inputs, especially once the watershed becomes near saturated (Beckers and Alila, 2004). Figure 6.17 provides an example of stream height response to rainfall and snowmelt input for differing antecedent conditions from November 2007 through January 2008 at two stream gauges in Russell Creek (30km²), one at the mouth (300m a.s.l.) and one located approximately 3km upstream draining an area 8km² (500m a.s.l.). For all events, peaks in depth at the two gauges were within 30 to 60 minutes of each other, suggesting portions of the watershed respond at a similar rate. In addition, presence of snow on the ground did not appear to substantially alter timing of peak between gauges, implying inputs due to snowmelt were rapid during ROS. There were rapid rises in streamflow regardless of how much snow was on the ground, however, once a moderately deep snow-pack was established, the duration of flows was extended. Anderson et al. (2009) found that preferential flow velocities were strongly related to 1 hr rainfall intensities, so processes such as snowmelt which increase intensity of water delivery would speed up preferential flow rates. In environments that have extensive preferential flow networks with rapid movement of
water, it seems logical that increased melt rates will translate directly to streamflow. It also stands to reason that road networks will have less of an influence on enhancing steam flow than in watershed where matrix flow dominates.

![Graphs showing air temperature, rainfall, and water depth over time.](image)

Figure 6.17. Rainfall, and air temperature (840m a.s.l), snow depth at the Upper Stephanie (US) clear-cut (CC) and old growth (OG) forests, and water depth at Stephanie Creek (8km²; 500m a.s.l) and Russell Creek (30km²; 300m a.s.l.) gauging stations from November 12, 2007 to January 15th 2008.

It is a medium term goal to utilize a fully validated hydrological model to assess changes due to forest harvesting and subsequent regeneration on the magnitude and frequency of streamflow at Russell Creek, but as described in Chapter 5, improvements to our general understanding of some processes must be overcome. Even so, generalizations can be made of how changes to water delivery as a result of forest harvest and subsequent regeneration at the stand and hillslope may impact streamflows at the
watershed scale by examining peak flow frequency distributions. Russell Creek has 11 years of streamflow records from 1992 to 2003, unfortunately there were no reliable flow measurements for the duration of this study due to significant changes in the rating curve as a result of a series of high magnitude flow events. However, the Tsitika River watershed (360km$^2$) of which Russell Creek is a sub-basin, has a continuous streamflow record from 1975-2010 collected through Environment Canada’s Water Survey Program (www.ec.gc.ca/rhc-wsc/), providing an opportunity to put potential increases in the frequency of streamflows at Russell Creek in the context of a much larger watershed. In addition, McAllister Creek (37 km$^2$) is located approx 55km NE of Russell Creek with a streamflow record from 1998-2012 that is also maintained by Environment Canada. McAllister Creek’s peak elevation is 1550m, thus it is slightly lower than Russell Creek, but it has similar climate, topography and parent material and with its more northern latitude, we would expect it to have a slightly lower transient snow zone which the lower overall elevation should compensate for. A sensitivity analysis was performed by augmenting the flood frequency distribution of the 3 watersheds by 5, 10 and 20%. All of the watersheds have extensive forest harvesting within them, thus it is likely the frequency distributions have already shifted.

It is typical for slopes of flood frequency distributions to decrease as events become larger in magnitude, and Russell Creek is no exception (Figure 6.18B). The Tsitika River frequency distribution also shows a relatively shallow slope for the largest events, but in general, the slope of the entire frequency distribution is steeper (Figure 6.18D). There is a sharp increase in slope at approximately the 4 year return interval at the Tsitika that is not evident at either McAllister or Russell Creek. McAllister Creek followed a similar pattern to Russell Creek peak streamflow. At Russell Creek, minimal increases in magnitude of peak streamflow affected the frequency distributions similar to the 24hr stand level comparisons. For example, the largest flow on record had a return interval of 18 years, however, with only a 5% increase, the return interval became 4.2 years and at 20% it was 1.8 years. It is revealing that studies using CP often state increases of 10 to 20 percent as insignificant or minor (e.g. Jones and Perkins, 2010), yet when assessing changes in a frequency framework, these seemingly small increases can result in the larger events becoming 10 times more common. An interesting
characteristic of the frequency distribution at McAllister Creek and the 24hr rain from Russell Creek is how flat the slope is between return intervals of 1 and 2 years. This suggests return intervals in this range, which are often associated with channel forming/maintaining events, are even more sensitive to increases in flow. Combine this with extreme flows becoming more common and the impact on both stream channel morphology and sediment transport could be dramatic. This sensitivity will also result in substantial decreases in the frequency of flows as forests regenerate, increasing the importance of reforestation and using silviculture methods to enhance growth. It is important not to over-emphasize specific changes in slope over short sections of the frequency distributions due to the high amount of uncertainty in streamflow calculations, especially at higher flows – however, they should not be ignored, especially if they have physical implications.

The rapid response of streamflow to rainfall and snowmelt in coastal watersheds with shallow soils, preferential flow paths and steep gradients implies that an increase in the variability of time of input at the stand or hillslope level will be mirrored in the variability of peak flows, thereby steepening the CDF, resulting in much greater effects on the frequency of large events compared to small and medium flows. Previous lines of reasoning in forest hydrology have suggested that removal of the forest canopy has minimal effects on large or extreme events because any increases from melt or interception are minimal compared to overall inputs – i.e. the system is “overwhelmed” such that any mitigation provided by interception or deficits created by transpiration are negligible (e.g. Bathurst et al., 2011). However, if an increase in variability leads to a decrease in slope of the entire flood frequency distribution, than the same increase in magnitude will have a much greater increase on return intervals of specific events. Rather than the effects of forest harvesting being “overwhelmed” by the background meteorological and snowmelt inputs from forested stands, the entire watershed is responding within a completely new frequency distribution that may be even more sensitive to increases than before. This shift in the frequency distribution discounts the importance of the system being “overwhelmed” during individual events and places more importance on how watershed sensitivity changes with increased variability as a result of
forest harvesting. This will be an important aspect to consider when scaling stand level hydrological recovery to the watershed.

It must be highlighted that there is not always a direct relationship between the 24hr rain return intervals and peak streamflow, suggesting that in many instances stand level results will not translate to the watershed. An excellent example of this is the 24hr rain storm of record at Russell Creek on November 15th, 2006. The frequency analysis suggested an event of this magnitude occurs once every 23.3 years. It was associated with 100’s of landslides in the vicinity of Russell Creek (Guthrie et al., 2010), with at least two documented debris flows within the watershed and a third off the backside that closed Highway 19. There was also a relatively shallow snow-pack across the middle and lower elevations of the watershed, however, both photo analysis and energy balance calculations indicated snowmelt was minimal. Despite the rainfall intensity and the resultant high rates of landslides, streamflows were not extraordinary, with the Tsitika River peak flow having a return interval of approximately 1.05 years and McAllister 0.28 years. Even if records were compared over the same period for the 24hr rain and Tsitika River peak flows (1995-2010), the return interval was still only 1.63 years. There were examples with a direct link to 24hr rainfall intensity and peak flow return intervals as illustrated on November 12, 2007 in which both the Tsitika River peak flow and Russell Creek 24hr rainfall intensity had return intervals of approx 8.6 years when the same period of record was compared (1995-2010). In contrast, the return interval was only 0.74 years at McAllister Creek. As in 2006, a shallow snow-pack was present across all elevations of the watersheds and minimal snowmelt was observed, however, streamflows responded differently. These examples illustrate the importance of scale when assessing forest harvesting impacts on watershed processes. They also demonstrate how important synchronization of watershed processes are to the generation of peak flows during ROS (Jones and Perkins, 2010; Perkins and Jones, 2008), while at the same time showing extreme ROS events can have wide spread impacts on watershed processes without causing extreme flows.
Figure 6.18. Return intervals (Tr) for (A) 24hr maximum rainfall from the Upper Stephanie (US) (840mm a.s.l) climate station (1995-2010), (B) peak discharge at Russell Creek (30km²) from 1993 to 2001, (C) McAllister Creek (37km²) from 1998-2010 and (D) Tsitika River (360km²) from 1975 to 2010.
6.5 Summary and Conclusions

Through the use of snowmelt modelling and detailed observations in eight forest stands it was shown that both the mean and variability of energy available for melt was highest in clear-cut locations. In addition, the frequency of high energy events were greater in clear-cuts compared to old growth stands. When recovery was assessed based on frequency, there did not appear to be a significant reduction in energy during some ROS events in forests 4m in height. By the time forests were 11m and 13m in height, energy inputs during ROS appeared to be approaching similar mean and distributions to the OG stands, however, rates of recovery in many situations were reduced as events became larger and less frequent.

The results indicated that in relatively sheltered areas, absolute differences in energy available for melt among different forest types depends on exposure to winds and incoming energy fluxes. Because of this, more extensive forest harvesting may be permissible in areas that do not receive high amounts of wind. This will affect the general accumulation of snow within open vs. forested areas, resulting in potentially more water delivery to the soil profile, especially during long duration events, however, for individual, short duration high intensity events, the effects on streamflows may be limited. Restrictions on harvesting or use of alternative silviculture systems such as single tree selection or partial cutting in areas exposed to high winds could minimize the potential impacts of increased snowmelt during ROS. It is likely that it will take longer for these areas to recover as trees will be less able to attenuate winds due to weaker physical structure compared to OG stands. Time to recovery will also be delayed when harvesting at elevations with deep snowpacks that can suppress regeneration of trees until they are taller than the snowpack for most of the growing season. Particular attention should be paid to limit harvesting on unstable terrain where ROS is possible, for even if an event does not greatly affect streamflow, melt can be sufficient to augment water input and increase landslide hazard.

This research is limited to stand level energy flux recovery and the type of events that occurred during the study period. It may be used to improve the current methods surrounding application of cut control and stand level hydrological recovery to minimize the effect on peak flows in watersheds frequented by ROS. While there is not a direct linear
relation between forest harvesting, stand level hydrological recovery and streamflow, our physical understanding of forested watersheds dictates there will be increases in streamflows when certain conditions are met. If harvest levels are extensive in portions of watersheds which receive high amounts of energy, increases in the frequency of large magnitude flood events are probable. As forests regenerate, the flood frequency distribution should begin to shift back towards its original shape and location.

Results from this chapter show that forest cover removal alters the frequency of energy inputs during ROS by increasing mean and variability. This provides a clear link between physical processes and frequency distributions that has not been fully assessed in past stand level research. It can be argued that the frequency of an event depends on record length and thus one cannot make decisive conclusions about changes to return intervals, especially in non-stationary environments. While this is true, flood frequency distributions are a function of the physical characteristics of a watershed and the dominant climate regime (Eagleson, 1972). No matter the length of streamflow or climate record, if changes to mean and variability alter frequency, return intervals will shift and result in changes to physical processes. Just as there is concern over climate change altering the frequency of extreme events through changes in mean and variability (Katz, 1993; Katz and Brown, 1992), an equal emphasis must be placed on how landuse changes such as forest cover removal alter frequency distributions. The results clearly showed that what might seem like an inconsequential amount of additional melt, can completely alter the frequency distribution causing “problem” events to happen more often. When this occurs, mitigation strategies will include determining rates of recovery that must be calculated in a frequency framework. Return intervals are often the characteristic of greatest importance, especially for bridge design, flood plain mapping or changes in channel morphology – how much bigger an event is of minor importance to how often a bridge is destroyed or a community flooded.
Chapter 7: Summary and conclusions

Rain-on-snow (ROS) is a process that has garnered much interest in the hydrological sciences due to its relation to floods and landslides (e.g. Marks et al., 1998; Guthrie et al., 2010). Of particular interest to the field of forest hydrology is the effect of forest removal on ROS processes and streamflows (e.g. Harr, 1986; Berris and Harr, 1987). Research has generally shown an increase in melt rates from forest openings, but this has not always translated into significant increases in streamflows at the watershed scale. Recent research using a fundamentally different analysis technique known as frequency pairing has suggested that forest harvesting may have greater impacts on streamflow than previously thought (Alila et al., 2009; Kuraś et al., 2012). This increases the importance of assessing hydrological recovery in watersheds frequented by ROS and hydrological modelling provides a means to accomplish this. However, uncertainties still exist in model structure and the data used to both run and validate simulations (e.g. Rutter et al., 2009). An initial step towards reducing this uncertainty will be to test existing models in ROS environments within regenerating forests to assess model strengths and weaknesses and determine which outputs can be used for further analysis. This dissertation used an innovative approach to collecting ROS data to test an interception and snowmelt model using previously unavailable observations. Outputs from the model were then used to compare energy available for snowmelt among different forest types during ROS using frequency based analysis. These comparisons were used as a basis for discussion on stand level energy flux recovery.

Collecting data in ROS watersheds has always been a challenge. A primary objective of this dissertation was to assemble a dataset that was previously unavailable, and while some problems were not overcome, an unprecedented amount of data were collected through an innovative design, primarily from a time-lapse photography network to monitor snow depth and snow in the canopy. Automatically processing the time-lapse images was challenging due to changing illumination which made it difficult to select consistent thresholds to identify snow. This problem was overcome by selecting times of day with relatively stable lighting, along with a substantial amount of manual processing. These data were vital during the model validation process.

The observations derived from the time-lapse photography, climate stations and snow temperatures identified key deficiencies in current model structure of the Cold Regions
Hydrological Model (CRHM), specifically the SNOBAL module (Marks and Dozier, 1992), which had poor simulations of SWE at some locations and during different times of year compared to the bi-weekly to monthly snow course observations. This may result from the overly simplistic representation of the snowpack that assumes two layers and matrix flow of rainfall and melt, when both field observations and research elsewhere suggest preferential flow is a dominant mechanism of water delivery, especially in snowpacks with multiple ice layers. The model over-estimated cold content when compared to observed values which may have contributed to over estimation of SWE when rain was falling on snow, causing it to freeze and remain in the snowpack. In addition, defining thresholds for snow vs. rain was challenging due to temperatures around the freezing mark and sensitivity of the dataset to thresholds differing by even 0.5°C. The time-lapse photography network provided additional insight into snow interception dynamics and model performance of canopy loading and unloading. The canopy module (Ellis et al., 2010) was able to simulate the timing of loading and unloading of snow from the canopy, however, there were some periods where the amount of snow remaining in the canopy differed between simulations and observations. The snow depth observations derived from the time lapse photography network proved very useful to identify model strengths and weaknesses, suggesting much better model performance compared to the limited observations of SWE. When simulating snow depth, the model worked best at all locations during the snow accumulation phase and the period of sustained snowmelt associated with the freshet. When snowpacks were shallow during the early and late season, model performance was generally poor, similar to other applications of SNOBAL and CRHM (e.g. Mazurkiewicz, et al., 2008; Essery et al., 2009; Rutter et al., 2009). Even though the model did not consistently produce accurate depths and SWE, it performed well at capturing the changes in depth during ROS at all locations, suggesting it was able to effectively simulate above ground energy fluxes. While turbulent exchanges were important during ROS, both net radiation and ground heat fluxes were the primary contributor to snowmelt during some periods of the year, contrary to other studies where sensible and latent heat fluxes dominated the energy balance (e.g. Berris and Harr, 1987; Marks et al., 1998).

The outputs from CRHM, specifically snowmelt energy during ROS, were used to compare potential melt rates among different forest types. Fortunately, a relatively high
number of events from 2005-2008 were captured creating a dataset suitable for frequency based analysis. This provided a unique opportunity to assess how forest removal and regeneration of trees impacts the mean, variability and frequency of melt energy during ROS. The results showed that forest harvesting can have significant impacts on the magnitude and frequency of melt energy during ROS by changing the distribution of events through increases in both the mean and variability. If only the means were compared, differences would have been considered minor between the forest types, however, including measures of variability and frequency showed that differences were greatest during the largest magnitude and lowest frequency events. When put in context with the 24hr maximum precipitation record, it is probable these increases will result in significant changes to return intervals of water delivery to the soil profile. In watersheds with shallow soils and high runoff coefficients, it is likely that increases in water available for runoff (WAR) at the stand or hillslope scale will translate directly to increases in streamflow.

Measures of recovery depended on the variable being compared and location within the watershed. In portions of watersheds exposed to higher winds and turbulent energy, recovery may take longer and differences between stands will be greater. At the lower elevation forests that were sheltered from high winds, stands 13m in height were 100% recovered for energy inputs during ROS, while an 11m tall forest at the higher elevation location with exposure to higher winds and turbulent exchange was only partially recovered, despite having very similar canopy closure to the 13m stand. Forests that were 4m tall and had large spaces between trees showed signs of recovery for some energy sources, however, there were instances where potential melt rates were higher than the clear-cut plots, especially at the higher elevation location. It was also revealed that rates of recovery differed for mean, variability and frequency, a result that has not been previously reported (e.g. Hudson, 2000a; Buttles et al., 2005; Talbot et al., 2006; Talbot, and Plamondon, 2002; Winkler et al., 2005). By comparing the frequency of events, the effects of differences in mean and variability was more fully explored. In situations where the mean and variability were not equal to the old growth forests, such as the 11m stand, energy flux recovery decreased as events became larger, which has not been documented in previous research.
7.1 Study limitations

There were a number of limitations in the analysis which included uncertainty in both the model inputs and observations, and use of a short term dataset to make inferences about return intervals of importance. These deficiencies cannot be overlooked and the results must be considered in this context. As is often the case, the dataset used was of short duration, spanning 2 years for model testing and 3 years for comparison of ROS events. Even though the dataset covered the entire range of 24hr maximum rainfall intensities on record from 1995-2011, there was no information related to the frequency of ROS events over that period, including the extent of antecedent conditions for snow depth, SWE and cold content, which would affect the amount of snowmelt from the varying forest types. Melt from the canopy was also not included in the calculations for WAR, thus it is possible that differences among forest types may have been less than the results suggest. These limitations impact how far the results can be extended in a flood or intensity duration frequency framework because the underlying distribution of WAR as a result of ROS was unknown.

Sample sizes were limited, a common problem in field research. It was assumed that 8 locations, 4 within each elevation range, would capture the variability in the dataset specific to forest types. This was done through controlling aspect, slope and elevation among forest plots to limit confounding variables. Because of this, no information for energy flux recovery was collected at different aspects or elevations, nor in forests between 4m and 11m or greater than 13m, thus the measures of energy balance were limited. Uncertainty was not explicitly accounted for in the analysis due in part because it was unknown for many of the model input variables. Factors such as under-catch at the precipitation gauges and frozen gauges may have introduced uncertainty into the dataset which would be difficult to quantify. In addition, inputs to CRHM were on an hourly time step and used average wind speed. Wind speed can be highly variable through a one hour period and the overall inputs of turbulent energy may have been suppressed by using averages. Opportunities exist to assess some of these deficiencies in future research at Russell Creek Experimental Watershed and at other study basins in maritime climates.
7.2 Future opportunities

The initial analysis from this dissertation has already led to additional weather station installations at Russell Creek in a variety of locations exposed to different types of energy, from windy to sheltered and sunny to shaded across all elevations. A primary goal is to test and validate CRHM at these locations through improvements to the interception modules and the development of a multi-layer snowmelt model that accounts for preferential flow. Efforts are underway to reduce uncertainty in streamflow data through the development of an automated salt dilution stream gauging network to scale stand and hillslope processes to the watershed. Because multiple peak flows can occur in a single winter, stream cross sections constantly change. The automated salt dilution network allows for a rapid rebuild of streamflow rating curves following an event. The automated measurements also provide an opportunity to measure peak discharges remotely to target high flows which are lacking in the record. The watershed was recently flown using Light Detection and Ranging (LIDAR) to create a digital elevation model and describe the forest canopy, providing a baseline to re-acquire LIDAR at peak snow accumulation to gain a better understanding of snow depth variability in complex terrain. The increasing length of the data record at Russell Creek and planned work to improve model simulations will allow for additional frequency based analysis at the watershed scale to determine the effects of forest harvesting on the mean, variability and frequency of streamflows and landslide events at Russell Creek. As the dataset lengthens, and the climate station and observation network expands, the probability of capturing extreme events will increase. In addition, more measurements both spatially and temporally will reduce uncertainty. Improving the model structure and reducing uncertainty will allow for a more complete assessment of watershed scale hydrological recovery and provide an opportunity to assess how climate change and forest management will affect coastal snowpacks in British Columbia.

This research focused on a single aspect of hydrological recovery – above ground energy available for melt during ROS. Additional factors must be included in hydrological recovery assessments, such as water yield (e.g. Stednick, 1996); changes in timing of melt during the spring freshet (e.g. Moore and Wondzell; 2005) and alteration of nutrients and temperature regimes (e.g. Danehy et al., 2011). Recovery of stream channels as a result of increased streamflow, riparian logging and landslides is of major importance (e.g. Bilby and
Ward, 1991; Gomi et al., 2001; Gomi et al., 2003). In productive forests, stand level energy flux recovery may occur within two or three decades of disturbance, however, other processes may take a century or more, such as the requirement for large diameter trees in riparian areas for stream bank integrity and large woody debris input (e.g. Reeves, 2003). While the original flow regime in these watersheds may return to pre-harvest levels simply by allowing forests to regenerate, recovery of disturbed channels and riparian areas may require intervention through targeted restoration. Future research must focus on combining all aspects of measuring and detecting watershed recovery to allow for sustainable management of forest resources.

7.3 Contributions to forest management and hydrology

Despite limitations, results from this dissertation can be directly applied to current operational applications of recovery within coastal British Columbia. It can be used to update current recovery curves in watersheds that routinely experience ROS to minimize the impacts on streamflows and landslides. This research will guide improvements to modelling in both Russell Creek Experimental watershed and others with similar characteristics. This thesis has made the following contribution to forest hydrology:

- First application in British Columbia of a physically-based interception and snowmelt model in clear-cut, regenerating and old growth forests in a watershed frequented by ROS;
- Tested the model using previously unavailable observations via a time-lapse photography network;
- Identified key model deficiencies in simulating snow water equivalent which may be improved through development of algorithms to account for preferential flow in coastal snowpacks with distinct ice layers;
- The first study to assess stand level hydrological recovery using frequency based analysis, showing that when mean and variability are different among regenerating forests, energy flux recovery decreases as events become larger in magnitude.

Two general themes in this dissertation were addressed which are vitally important to moving forward in the science of forest hydrology: utilizing emerging technology to collect more and better data, and; using these data along with the best available methods to answer
both emerging and long standing questions (Kirchner, 2006; Soulsby et al., 2008). Collecting data in mountainous, remote environments will always be a challenge, however these datasets are critically important. Integral to development of these datasets will be expansion of current networks and collection of new observations through increased instrumentation of existing weather stations, application of remote sensing products such as LIDAR and utilization of emerging technology such as wireless networks to increase spatial coverage of observations (Soulsby et al., 2008). As our datasets grow in both time and space, we will be able to use new methods of analysis, including advancements in data processing modelling and statistics to improve our physical understanding of systems and develop operational tools for practitioners. This will provide opportunities to test current paradigms which may be inadvertently suppressing the advancement of forest hydrology.

This dissertation contributes to a growing body of research that challenges the use of well established analysis techniques in forest hydrology, mainly chronological pairing and ANOVA/ANCOVA analysis. While the results from this research are not conclusive, they point to the strengths of using frequency based analysis to detect differences and similarities among forest types and have produced different results from previous studies. It is well established in the hydrological sciences that landuse changes introduce non-stationarity into flood frequency distributions (e.g. Hosking and Wallis, 1997). If changes to the magnitude and frequency of streamflows and landslides as a result of forest harvesting are of primary concern, these results should encourage others to compare hydrological variables using this approach. Furthermore, frequency analysis is not the only method to analyze data and others may be considered as data availability dictates, however, the strengths and weaknesses of all methods should be considered to ensure they can answer the questions of concern and/or the hypothesis being tested. We are about to reach the end of the decade long initiative of Predictions in Unguaged Basins (http://pub/iahs.info), whose primary goal was to reduce uncertainty in hydrological predictions. How much progress we have made in forest hydrology towards this is difficult to measure, however, based on the level of debate that still occurs over the most fundamental of questions, we are still far from this goal. As Kirchner (2006) so succinctly stated: “We need to develop better models and better analysis tools; we also need to create better data to model and analyze”. It is my hope that the results from this
thesis met some of these needs and will contribute to future work on stand and watershed scale hydrological recovery.
Works cited


Green, K. and Alila, Y., 2012. A paradigm shift in understanding and quantifying the effects of forest harvesting on floods in snow environment: IN REVIEW. Water Resources Research.


Kirchner, J.W., 2006. Getting the right answers for the right reasons: Linking measurements, analyses, and models to advance the science of hydrology. Water Resources Research, 42(3): W03S04.


Talbot, J. and Plamondon, A.P., 2002. The diminution of snowmelt rate with forest re-growth as an index of peak hydrologic recovery, Montmorency Forest, Quebec. In: J. Hardy


