Snow Ablation Processes and Associated Atmospheric Conditions in a High-Elevation Semi-Arid Basin of Western Canada

by

Scott Isaac Jackson
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Abstract

Snow surface energy balance was studied along an elevational gradient and under varying forest cover types during the ablation season of 2007 in the Coldstream Basin, Okanagan, British Columbia, Canada. During the snowmelt period, 1-4% of the peak annual snow-water equivalent (SWE) was lost to sublimation in open sites – averaging 0.4 mm d^{-1}. Melt and sublimation rates increased significantly with elevation, and were higher and more variable in the open sites than under forest canopies. Melt rates were driven almost entirely by sensible heat fluxes and exceeded 30 mm d^{-1} during large-scale advection events. The melt and sublimation processes observed at the snow surface were significantly linked to conditions in the atmospheric boundary layer. From these linkages, a proxy record of historical ablation season energy fluxes for the period 1972-2007 was created. Significant trends towards earlier dates of snowmelt and freshet onset were detected, as was a trend towards increasing ablation-season temperatures at the 850 mb height. Significant correlations between estimated historical ablation-season melt and sublimation and the regionally dominant teleconnection indices were also found. This study significantly advances the understanding of ablation season snow-surface energy exchanges, and the links to the driving atmospheric conditions in the Okanagan Basin.

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CHAPTER 1: INTRODUCTION

1.0 Introduction

Seasonal snow cover is inextricably linked and integral to the maintenance of ecological, hydrological, geomorphological and climatic systems in the Northern Hemisphere (Gray and Male, 1981). Most of Canada is covered in snow for several months of the year, and in the Northern Hemisphere, this snowpack exerts the strongest feedback to the Earth's radiation balance in the spring period. This radiation balance is the primary driver of the Earth's atmospheric circulation system (Groisman *et al.,* 1994). When extensive snow cover melts it recharges soil moisture, aquifers, river systems, ecosystems, and lakes and can cause flooding (Gray and Male, 1981). The seasonal snowpack in western Canada at high elevations lasts much longer on average than that at lower elevations, is deeper, and stores the majority of annual precipitation inputs – to be released over a longer (relative to central and eastern Canada) spring melt period. In British Columbia, and the unique semi-arid Okanagan Basin in particular, this snowmelt provides water inputs to lower elevations at a critical period for agricultural activities (Bonsal *et al.,* 2003; Cohen *et al.,* 2006).

There has been extensive research into the physical processes that control snowmelt energetics in various environments and under differing climatic conditions and many models have been developed, both as explanatory and predictive tools. However, despite recognition of the importance of snowmelt processes to natural and human systems, large knowledge gaps still exist including: the processes controlling the partitioning of energy and water fluxes within and between hydrologic systems (Bales *et al.,* 2006); ablation season processes in forested areas other than the boreal forest (Buttle *et al.,* 2000); turbulent energy fluxes over snow – particularly in heterogeneous terrain (Helgason and Pomeroy, 2005) and, the potential effects of an intensifying hydrological cycle (Huntington, 2006).

The Okanagan Basin contains British Columbia's largest agricultural centre, however, the intensive irrigation required to maintain its productivity, coupled with rapid development make it one of Canada's most water stressed large catchments. Climate projections generated by global circulations models (GCMs) suggest that as winter temperature and precipitation increase over the next century, less precipitation will fall as snow, the snowmelt season will occur 4-6 weeks earlier, and conditions conducive to snow sublimation/evaporation losses will become more frequent resulting in "considerable reductions" in annual and spring flow volumes (Merritt *et al.,* 2006). These findings are reflected in other studies of western North American watersheds fed by alpine snow (Lemke *et al*., 2007). To date, most hydrological research in the Okanagan has focused on forestry impacts and the modelling of hydrologic processes (Winkler and Moore, 2006; Merritt *et al*., 2006). Research literature and basin meteorological data concerning the climatic drivers of snow energy and mass balance in the higher elevations of the basin are scarce – a serious gap considering that the hydrology of the semi-arid Okanagan Basin is driven by the accumulation and ablation of the snowpack (Merritt *et al*., 2006). Knowledge of the physiographic, vegetative and atmospheric controls on energy and mass balance on snow ablation will provide water managers and researchers with the necessary data to formulate more robust methods of adapting to water shortages exacerbated by predicted climate change. At the initiation of this research, field based measurements of the turbulent and radiative fluxes driving snow ablation processes (melt, sublimation/evaporation and condensation) had not yet been undertaken in the semi-arid Okanagan Basin. In addition, anthropogenic climate change and land cover changes at high elevations (i.e., mountain pine beetle salvage and forest fires resulting in larger openings) require that the effects of elevation and forest cover on the late-winter and early spring snow energy balance be more completely understood. Furthermore, knowledge of the upper atmospheric conditions driving these energy and mass exchanges is extremely limited. Therefore, in order for realistic projections to be made of climate change effects on water supply within the Okanagan Basin, and science based water management policies to be effective in the future, these knowledge gaps must be addressed.

1.1 Research Area

The Okanagan Basin (8046 km^2) is an excellent example of an under-studied semi-arid environment that is heavily dependent on the spring snowmelt to recharge its freshwater resources. It extends approximately 185 km from north to south and comprises a portion of the northern extent of the Columbia River Basin, a trans-boundary watershed. The valley bottom is occupied by a mainstem river-lake system that drains to the south, comprising Okanagan Lake (351 km^2), and the smaller Kalamalka Lake, Wood Lake, Skaha Lake, Vaseux Lake and the Okanagan River (Summit, 2005). These lakes are recharged by 31 main tributaries flowing from the surrounding uplands. Topographic relief from the valley bottom to the surrounding peaks averages 1100 m. The basin is characterized by a dry continental climate, with precipitation averaging 250-300 mm yr^{-1} in the valley (85% of which is estimated to be lost to evapotranspiration) and >1000 mm at the higher elevations (Cohen and Kulkarni, 2001). Summer rainfall is largely driven by local scale convection, with winter precipitation resulting from synoptic systems originating in the Pacific Ocean. Average annual temperature decreases and precipitation increases along a south to north transect within the basin, with the southern portion containing Canada's only true desert ecosystem (Cohen and Kulkarni, 2001).

The primary water sources in the Okanagan are the tributary streams, most of which are at license capacity and listed as "fully recorded", or "water shortage", leading to water allocation conflicts, particularly in dry years such as the drought of 2003 (Cohen and Neale, 2006). A comprehensive summary of the groundwater component of the basins water balance is provided by Neilson-Welch and Allen (2007). Its freshwater resources in 45 community watersheds are under increasing stress from a convergence of several pressures, including increases in development and population (projected to rise to 450 000 in 2031 from 210 000 in 1986), intensive agricultural irrigation (70% of diversions), extensive logging at higher elevations (partially a result of mountain pine beetle salvage operations), and projected climate change. Two of the three main impacts of future climate change in the Okanagan Basin identified by Cohen and Neale (2006) are: reductions in annual water supply resulting from reduced storage in snowpacks and, earlier peak flows in spring, with uncertainty regarding changes in timing and magnitude.

These predictions mean that water resources will experience their greatest annual stress during the periods of highest demand during the growing season, and that the sensitive snowpacks at high elevations will play a proportionately larger role in future water availability. The studies to date highlight two serious shortcomings in the current data that need to be addressed in order for more accurate predictions to be made: establish long-term, high elevation climate stations, and begin detailed studies of precipitation/elevation relationships (Cohen and Kulkarni, 2001; Merritt *et al.*, 2006).

1.2 Research Needs

Several areas requiring further investigation regarding snow mass/energy balance studies are recognized by the scientific community for high elevation areas in general, and the Okanagan in particular. While the understanding of snow processes at the point scale, and the modelling of snow processes has progressed a great deal in the past two decades, all papers cited here note the need for further ground-truthing of model outputs, and more research on the spatial variability of the relevant processes. In particular, the degree of spatial variability resulting from high-relief topography and the associated modifications of the turbulent boundary layer is still poorly understood. This is especially true in the Okanagan Basin, where high-elevation climate stations are virtually non-existent. Only one climate station includes a radiation sensor (Summerland CS), and 21 snow courses (bi-monthly) and pillows (hourly) provide indexes of snow water equivalent (SWE) at a coarse spatial resolution. Finally, as 100% of the potential license capacity of all surface water in the Okanagan is currently utilized, and as this is based on an "average" water year, it is vitally important that the basins water balance be more completely understood. As most of the surface water in the Okanagan is recharged by snowmelt, a more complete understanding of the climatic and physiographic variables driving snow ablation is urgently needed.

1.3 Objectives and Thesis Format

A literature review was conducted to assess the state of knowledge in the field of study, and is presented in Chapter 2. To address the research needs listed above, the following objectives were identified.

Objective 1 is addressed in Chapter 3, written as a stand-alone journal style manuscript.

1) *Quantify ablation season snow-energy and mass-balance processes that characterize melt conditions along an elevational and forest-cover gradient in a selected catchment of the Okanagan Basin.*

Objectives 2 and 3 are similarly addressed in a stand-alone journal style manuscript, presented in Chapter 4.

2) *Assess how well atmospheric boundary layer variables are correlated with ablation-season snow-surface energy exchanges, and determine if these relationships are useful predictors of snowmelt and sublimation events during past ablation seasons.*
3) *Analyze time-series of hydro-meteorological variables for trends in magnitude and frequency, and define the relationships with winter and spring teleconnection indices.*

This thesis concludes with Chapter 5; a summary of the research findings presented in Chapters 3 and 4, and identifies future research avenues in snow ablation and linkages to prevailing climatic conditions.

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CHAPTER 2.0 LITERATURE REVIEW

To guide and frame the research presented within this thesis, a review of the pertinent literature was conducted. The following sub-sections focus on seven main areas: 1) snow accumulation patterns; 2) snow energy and mass balance; 3) canopy snow interception and sublimation processes; 4) snow distribution and process modelling; 5) climate and snow linkages; 6) upper atmosphere and surface energy flux relationships; and 7) trends in hydrologic and atmospheric boundary layer variables. The literature cited herein is not an exhaustive survey, but a summary of the current state of knowledge and recent advances in these areas of research. A statement at the end of each section describes how the summarized knowledge was applied to the research presented in this thesis. Note that due to the presentation style of this thesis (manuscript vs. traditional thesis format); there will be necessary repetition between the introduction, literature review, and the two manuscripts contained within.

2.1 Snow Accumulation

While the studies reviewed have utilized different approaches in many dissimilar environments, there is general agreement about the factors controlling snow distribution in open high elevation areas. They fall into two general categories: variables that modify the wind field and speed (most important), and those that affect radiation fluxes. The variables affecting these two processes are: topographic variation including roughness length, elevation, slope and aspect; vegetation type, distribution and height (Martinec and Sevrup, 1992; Winstral and Marks, 2002). In summary, any feature (topographic or vegetative) that causes air flow divergence will result in increased deposition (i.e. drifting on the windward side of a tree, or cornices on the leeward side of a ridge). Windward slopes, flat topography and unvegetated areas will act as a source of snow, while leeward slopes, depressions and vegetated areas will act as sinks. As the net radiation and turbulent heat flux inputs to an area increase, the accumulated snow will decrease, and all studies reviewed state that the effects of wind have a much greater impact on snow distribution and spatially variable melt rates than solar radiation (Luce *et al.,* 1998; Winstral and Marks, 2002; Erickson *et al*., 2005).

In forested areas, snow accumulation is heavily influenced by the density and distribution of forest cover, with higher accumulation and ablation rates in open areas than under canopy cover (López-Moreno and Latron, 2008). Some of the first work that quantified these relationships in Canada found that mid-size openings in the Alberta foothills accumulated the most snow, as they constitute a balance between elimination of interception, disturbance of air flow over the canopy, and wind speed in the openings (Golding and Swanson, 1978). In southeastern BC, Toews and Gluns (1986) reported that SWE in clearcuts was 4 – 118% (mean difference 37%) greater than that under nearby forest cover. In southern BC, Winkler *et al.* (2005) found that the April 1 SWE was 32% and 14% less under mature and juvenile forest canopies respectively than in the clearcut. Snowmelt began first in a juvenile-thinned stand, and 30% of the April 1 SWE still remained under the mature canopy when the other 3 sites were snow-free. These differences in accumulation with varying forest cover are directly related to the efficiency with which the canopy intercepts snow (directly related to crown closure), and exposes it to higher rates of evaporation and sublimation (Pomeroy *et al*., 1998). The spacing and distribution of silvicultural treatments plays an important role in forested basins subject to harvesting. For example, in Montana it was found that grouped stand treatments accumulated significantly less snow and the associated SWE was three times more variable than SWE in evenly spaced stands (Woods *et al.,* 2006).

Within-stand variability in accumulation was given greater attention by Pomeroy *et al*. (2002) who report that the standard deviation of SWE was not associated with mean stand leaf area index (LAI), maximum accumulation in small clearings, or seasonal snow interception. However, at the stand scale an inverse relationship was found between snow accumulation and LAI. This relationship was assumed to hold given that mid-winter melt events, wind distribution and surface evaporation are minimal.

Disturbance events such as fire, wind-throw, disease and insect outbreaks can substantially alter stand structure and composition, and the distribution and size of open areas within the forest. In British Columbia's arid interior, the ongoing mountain pine beetle epidemic was found to influence snow accumulation and ablation in lodgepole pine stands (Boon, 2007). Accumulation in a beetle-killed stand was closer to that of a cleared stand than a live one, but ablation rates tracked those in the live stand more closely than the cleared stand. Similar relationships are likely to hold true in the semi-arid Okanagan Basin, as beetle salvage logging dramatically increases the size and number of forest openings. Therefore, this research examined the differences in the distribution of SWE between open and forested areas in a basin currently undergoing salvage logging of beetle killed lodgepole pine.

2.2 Snow Energy and Mass Balance

The energy and mass balance of snow are inextricably linked through complex transfers of energy and water vapour, which determine the rate and characteristics of snow ablation processes (melt, sublimation/evaporation). For example, Kuusisto (1986) identified the following characteristics of snow surface energy exchanges that are generally applicable in most areas studied:

1) The radiation balance and turbulent exchange processes play a major role; the contributions of heat from precipitation or heat exchange at the ground surface are small or negligible.
2) The radiation balance and sensible heat exchange are almost always positive during snowmelt periods.
3) Both evaporation and condensation may prevail during snowmelt; thus the latent heat flux may be negative or positive.
4) In forest environments the radiation balance is usually the most important energy component.
5) On cloudy or rainy days turbulent heat transfer dominates.
6) A very intense snowmelt usually also requires a large turbulent transfer.

Male and Granger (1981) provided the standard reference on the calculation of these exchanges, stating that the two most important energy exchange processes are radiation transfer (short- and long-wave) and the turbulent exchange process (sensible and latent heat transfer). The snow energy balance is expressed as

$$-\frac{dH}{dt} = R_{\mathrm{N}} + H_{\mathrm{S}} + H_{\mathrm{L}} + G + M \tag{1}$$

where *dH/dt* is the net rate of change of the snowpacks internal energy per unit area, and R_N, H_S, H_L, G and M are net radiative, sensible, latent, conductive (ground) and advective energy fluxes respectively. *dH/dt* is negative when the combined energy fluxes are positive, resulting in the ablation of the snow cover. The relative dominance of these variables changes with, elevation (Moore, 1983), continentality (Granger and Male, 1978), regional circulation types (Moore and Owens, 1984), forest cover (Golding, 1978; Suzuki *et al.*, 1999; Winkler *et al.*, 2005) and air temperature and humidity (Cline, 1997). The meso-scale spatial variability of these fluxes can also be quite high – even for unforested terrain, which complicates the scaling up of site specific values to an entire basin (Pomeroy *et al*., 2003; Pohl *et al*., 2006). In general, radiation is the most important contributor to snowmelt at continental, polar and high elevation sites; sensible heat exchanges dominate during chinook or föhn winds, and at some high elevation sites (e.g., Sierra Nevada); and latent heat transfers dominate under conditions of high humidity (including rain-on-snow events) and at maritime sites.

There are several methods currently in use that allow estimation of SWE vapour exchanges and melt: simple gravimetric measurements using lysimeters of various sizes and configurations (e.g., Golding, 1978; Zhang *et al*., 2004); bulk aerodynamic estimates made from single wind speed, temperature and humidity measurements above the ground surface (e.g. Moore, 1983; Suzuki *et al*., 2006), or along a logarithmic profile (e.g. Hood *et al*., 1999; Zhang *et al*., 2008) and; estimates derived from eddy covariance measurements of turbulence, temperature and water vapour (e.g. Pomeroy and Essery, 1999)

The use of bulk aerodynamic formula for these estimates assumes a stable boundary layer (often the case over melting snow), similarity of transfer coefficients and that any turbulence within this layer is a result of surface roughness. When more than one measurement level is present, the bulk transfer coefficients that depend on surface roughness and atmospheric stability can be derived using the Monin-Obukhov surface layer similarity theory, which states that the dimensionless vertical gradients for mean temperature and wind speed are functions of a non-dimensional stability parameter (Grachev, 2000; King *et al*., 2008). The Monin-Obukhov length is defined as the height above ground where mechanical turbulence is in balance with buoyant forces due to free convection, or where the Richardson number is equal to one. There are several methods used to calculate these stability functions, some of which are outlined and tested by Gellens-Meulenberghs (2005).

However, when only one level of measurement is available, the Richardson number (Ri) is commonly used to correct for atmospheric stability variations using empirical terms to account for non-similarity of the diffusion coefficients. This is done by relating the relative roles of mechanical forces (i.e., snow surface roughness) and buoyancy driven by convection in boundary layer turbulent flow. Therefore, under unstable conditions convection dominates, Ri is negative and increases (decreases) with the magnitude of the temperature (wind speed) gradient. Ri is positive under stable conditions, and under neutral conditions Ri is zero because the first assumption of the bulk transfer formulations is not violated and therefore requires no correction (Oke, 1987).

This stability correction approach is still widely used particularly where only one level of measurements is available, although Male and Granger (1981) thought it a poor estimator of vertical vapour flux. Shook and Gray (1997) clarified this and outlined four factors that may result in this method producing poor estimates of turbulent transfers over snow:

1) Questionable validity of a constant flux layer – a temperature maximum 10-50 cm above the snow surface from radiative heating can cause a reversal in the vertical heat flux at the height of this raised maximum.
2) Turbulent mixing above melting snow is dampened considerably by the dominance of stable conditions in the boundary layer.
3) Assumed equality of the three eddy diffusivities for momentum, sensible and latent heat may be false (e.g., Mawdsley and Brutsaert, 1977).
4) The turbulent heat fluxes are larger for bare surfaces than for snow covered areas due to differences in roughness length, and therefore the influence of small scale advection during patchy snow conditions is not accurately represented. In addition, an environment with high topographical relief and lower uninterrupted fetch lengths violates the last assumption of the bulk transfer method. This is that the generation of turbulence generated by the local topography instead of by the surface roughness of the snowpack (Helgason and Pomeroy, 2005; van den Broeke *et al.*, 2005).

One method used to account for these problems is the use of gravimetric measurements which, although labour and time intensive, allow direct quantification of melt and vapour fluxes between the snow cover and the overlying atmosphere, and provide a physical baseline for evaluating the accuracy of bulk transfer estimates (Nakai *et al*., 1999a; Suzuki *et al*., 1999; Zhang *et al*., 2004). Gravimetric measurements have the added benefit of allowing the roughness length (z_o) to be calibrated to the study site, instead of relying on empirical functions derived from studies undertaken in different land cover regimes.

The use of eddy covariance methods is gaining prominence for snow surface energy balance studies, as it allows the determination of turbulent fluxes of water vapour, momentum, sensible heat, or any other admixture from covariances between the vertical wind velocity and the concentration of the variable of interest (i.e., water vapour, sensible heat). Some benefits of this approach are that there are no moving parts in instrumentation and therefore no friction, shortwave radiation effects on the measured values are small, and the measurements have a high temporal resolution. However, the assumptions that predicate its use are: uniform surface, the boundary layer is in a steady state where vertical fluxes dominate turbulent exchange with the surface, a minimum 100:1 fetch length:height ratio and, that the properties of the air flow vary slowly. The instrumentation is also fragile, and unsuited to environments where extreme wind speeds and temperatures are often encountered, restricting the application of this method from some areas that still require much study (i.e., polar and high elevation areas) (Male and Granger, 1981).

Most formulations of snow energy balance assume a complete snow cover. This can lead to reduced accuracy of estimates of mass and energy transfers under patchy snow conditions where the local, small scale advection of sensible heat from bare ground to the snow cover begins to dominate. Research to date suggests that the bulk of the sensible heat transfer occurs along the leading edge of a snow patch, and that the fraction of sensible heat advected to the snow surface decreases exponentially with decreasing snow cover fraction (Neumann and Marsh, 1998). Boundary layer growth over a snow patch has been found to follow a power function, increasing in height with distance from the edge of the snow patch, and increasing upwind surface roughness (Granger *et al*., 2006).

An important sub-component of the snow energy balance consists of the flux of water vapour between the surface boundary layer and the snow surface, in the form of sublimation or evaporation, the phase change of water directly from solid or liquid to vapour from the snow surface, or condensation. Both are a function of wind speed, air temperature, humidity, snow particle size and solar radiation (Marsh, 1999). Depending on the specific site and climate characteristics, this process can account for the loss of 15-47% of SWE over a given snow season (Pomeroy *et al.,* 1997; Hood *et al.,* 1999). Sublimation can occur in one of three ways: from the snow cover itself (Cline, 1997); during blowing snow events, where the time and surface area exposure of a snow particle is greatly increased (Pomeroy and Essery, 1999; Liston and Sturm, 2004); or from snow intercepted by a forest canopy (see Section 2.3). It is interesting to note that some studies actually found an overall *decrease* in sublimation during blowing snow events; a result of a negative feedback between the increased initial sublimation during such an event, and the concomitant rapid increase in saturation of the air column (Mann *et al*., 2000; Déry and Yau, 2001). Condensation usually dominates at night or under high humidity, both highly stable conditions over snow cover. In general for open sites, high magnitude sublimation events are characterized by relatively low atmospheric water vapour concentrations and high wind speeds (Hood *et al*., 1999; Zhang *et al*., 2004).

The first work that attempted to estimate sublimation from a snow surface was conducted in Svalbard, Norway by Sverdrup and Ahlmann (1936), who found potential snow sublimation to be a function of vapour pressure deficits, wind speed and surface air pressure. Sublimation of the ground snowpack is influenced by canopy cover, radiation inputs, wind speed, aspect, snowpack characteristics, temperature and humidity (Schmidt *et al.,* 1998). Thorpe and Mason (1966) conducted a laboratory study of sublimation of ice crystals and sphere for various shapes, diameters, wind speeds and Reynolds numbers, and found that rates increased with increasing ventilation, and were influenced by the geometric shape of the ice crystal. This approach was subsequently applied by Pomeroy *et al*. (1998) while studying forest snow interception and sublimation. In a study of clearing size effect on snow evaporation, Bernier and Swanson (1992) found that in the smallest opening, night-time radiative heat transfer from the canopy increased evaporation, while evaporation in the larger openings was driven by greater turbulent heat transfer; a result of increased fetch lengths. The intermediate sized clearings were found to have the lowest evaporation rates, as they were not large enough to allow radiative heat transfer from the canopy, but were too small to allow high enough wind speeds to develop. A similar study in eastern Siberia reported that ~8% of the spring snow ablation was due to sublimation and evaporation, with no difference between an open site and one with a sparse canopy (Suzuki *et al*., 2006). In Colorado, Schmidt *et al*. (1998) noted that the snowpack to sublimation index decreased in proportion to the time since snowfall, and that rates varied from 0.43 – 0.61 mm/day for north and south aspects respectively.

This research seeks to incorporate the above knowledge gained from the study of snow surface energy balance, and apply it to a previously unstudied high-elevation, semi-arid environment in western Canada.

2.3 Interception and Sublimation from Forest Canopies

While wind and radiation are the most important factors influencing snow accumulation patterns in open environments, interception by forest canopies dictates SWE distribution more than any other factor in forested environments. Snow storage in the canopy is often an order of magnitude larger than that for rain, and the sublimation of intercepted snow is the most difficult term of the winter water-balance equation to quantify. An extensive review of this process, including its measurement and modelling is provided by Lundberg and Halldin (2001). The relationship between snow accumulation and forest cover is well documented, with an average of 40% greater SWE in clearings greater than 5 tree heights in diameter than under the canopy (Winkler *et al.*, 2005). This is due to the ability of the canopy to intercept, store and facilitate sublimation (and melt if temperatures greater than 0˚C) of fallen snow. The efficiency of interception and sublimation is directly related (in order) to storm snow density, canopy density, and time since snowfall (Hedstrom and Pomeroy, 1998; Nakai *et al*. 1999a; Lundberg and Koivusalo, 2003).

The first study to directly quantify interception used suspended cut trees to weigh intercepted snowfall during storms (Satterlund and Haupt, 1967). They found that interception storage increases sigmoidally with snowfall. This was confirmed by both Woods *et al.* (2006) and Schmidt and Gluns (1991), who determined that this relationship was produced by branch area and flexibility, and storm snow density. Interception efficiency is low when branches are snow free, as the particles fall through the gaps between needles. As a storm event progresses, interception efficiency increases as more snow is trapped by the branch, until it is completely snow covered, at which point interception efficiency decreases as snow particles now bounce off the branches bent under the weight of snow. The cohesive strength of the intercepted snow is determined by crystal size, type, riming, temperature and humidity (Bunnell *et al.,* 1985). At some point, depending on snow density, tree species and time since snowfall, the weight of snow overcomes the ability of the branch to support it and it falls to the ground (Hedstrom and Pomeroy, 1998).

The interception of snow by a canopy increases the surface area to mass ratio of newly fallen snow and exposes more of the snow to turbulent transfers of energy, which in turn facilitates sublimation and evaporation. Sublimation rates of intercepted snow have been reported to be several times greater than nearby open snow-covered areas, but taper off as the intercepted snow consolidates (Pomeroy *et al.,* 1998; Nakai *et al.,* 1999b; Lundberg and Halldin, 2001). Various methods have been tested to calculate sublimation from intercepted snow: Pomeroy and Gray (1995) applied fractal geometry and equations describing snow particle thermodynamics, turbulent and radiative exchange; Pomeroy *et al.* (1998) used interception and snow sublimation algorithms and scaled them up to the canopy scale to calculate snow mass balance and surface snow accumulation; and several studies have used direct measurements via the eddy correlation method to determine the surface energy balance of intercepted snow (Nakai *et al.,* 1999a; 1999b; Molotch *et al*., 2007). All studies reported that sublimation from canopy-intercepted snow ranges from

0.4 – 5 mm/day, and that sublimation was higher from the canopy than from nearby clearings, totalling as much as 100 mm SWE over the entire winter (Storck *et al*., 2002). Intercepted snow also results in an effective increase of LAI, which reduces the sub-canopy specific humidity gradient and radiation balance and acts to suppress sub-canopy snowpack sublimation (Lee and Mahrt, 2004; Molotch *et al*., 2007). A comprehensive review of forest snowfall interception and accumulation research conducted in Canada is provided by Buttle *et al*. (2000; 2005).

Because sublimation from snow intercepted by forest canopies largely occurs in the period following snowfall, the above-canopy energetics were not factored into this research. However, as these processes have a direct influence on the distribution of SWE on the ground, snow courses and gravimetric measurements were conducted beneath various canopy types and at various elevations within the study basin.

2.4 Modelling Snow Processes

Due to the ecological and economic importance of predicting the magnitude and timing of snowmelt and spring freshet, there have been concerted efforts to develop models that will allow water managers to predict these processes with accuracy. There are many energy and mass balance models available, varying from those that represent the snow cover as a single layer, to multi-layer models that incorporate algorithms dealing with grain size evolution (Jordan, 1991; Brun *et al.,* 1992; Link and Marks, 1999), and those that use coupled mass and energy balance routines to estimate snowmelt (Marks *et al.,* 1999). The traditional model used for most operational applications is the degree-day method, where snowmelt is calculated as a function of accumulated degree-days above 0°C (Rango and Martinec, 1995). While this method works adequately in regions where sensible heat is the main driver of snowmelt, it fails to accurately predict the magnitude and timing of snowmelt in environments where insolation dominates (e.g., Arctic tundra and high elevations). Thus, many modelling efforts focus on distribution of solar radiation with respect to topography and forest cover (Pohl *et al*., 2006; Ellis and Pomeroy, 2007).

There has been much progress in the development of models that are calibrated for a particular basin (Winstral and Marks, 2002; Thyer *et al.,* 2004), or those that deal with one or two of the processes in great detail (Pomeroy *et al.,* 1998; Marsh, 1999). However, there is still a gap between site and basin scales that remains insufficiently examined, although several models have begun to address this problem (Marks *et al.,* 1999; Lehning *et al.,* 2006; Pomeroy *et al*., 2007). In many of these models, error introduced due to uncertainty in the underlying processes (i.e., wind redistribution of snow, sublimation and radiation exchange) is removed by calibrating certain components of the model to match the basin hydrograph. In addition, these models are often run on less than five years of data (Pomeroy *et al.,* 1997; Winstral and Marks, 2002), leaving open the question of whether they would accurately replicate higher magnitude, lower frequency events or seasons, or whether they would account for a changing climate and the associated changes in process linkages.

The issue of scaling can complicate the application of models developed for processes at the site scale to larger areas where small errors at the site scale become exaggerated once extrapolated to the basin scale, reducing the accuracy of the modelled outcomes (Blöschl, 1999). This is particularly true of blowing snow processes, where sublimation from entrained snow begins to dominate over longer fetch lengths (Pomeroy and Gray, 1995). A number of studies have attempted to address this problem. For example, the ISNOBAL model has been tested in basins ranging from 1-2500 km^2, and was developed specifically for mountain basins; although the resolution of the grid cells isn't high enough to account for rugged high mountain topography, and it still requires intensive instrumentation to provide the necessary data to calculate the spatial variability of the processes and fluxes of interest (Garen and Marks, 2006). The ALPINE3D model is one of the most recent attempts at a process based snow mass and energy exchange model that operates at a fine enough resolution to address this issue (Lehning *et al.,* 2006). Another model designed for use in cold regions, where the hydrograph is representative of a nival regime, removes the issue of basin specific calibrations being required to produce accurate hydrographs. Instead it relies on physically based algorithms for each process of interest and the users understanding of the hydrological system to assemble the necessary parameters and structure the model accordingly (Pomeroy *et al*., 2007).

To put the large amounts of research effort that have been directed towards refining models over the past couple of decades into perspective, it might be instructive to take the approach that while all models are wrong (to varying degrees), some can be useful.

2.4.1 SNTHERM

Due to this studys focus on snow surface energy exchange, the point energy balance model SNTHERM was employed to model snow surface energy fluxes over the late-winter and spring of 2007. SNTHERM is a one-dimensional mass and energy balance model that was originally developed to simulate snow surface temperature following the passage of tanks (Jordan, 1991). It is adaptable to a full range of meteorological conditions, including precipitation events and a transition between snow covered and bare ground, and includes heat fluxes from the underlying soil. The model accounts for the metamorphism of multiple snow layers, and the accompanying movement of energy and water vapour between these layers. The model is initialized with the number of snow and soil layers, and the density or water content of the snow, temperature and grain size of each layer. Also required are the conditions in the atmospheric boundary layer; at a minimum, air temperature, relative humidity, wind speed, precipitation and incoming solar radiation provide the meteorological inputs to the model. Incoming longwave and outgoing solar radiation can be computed from inputs of fractional cloud cover, type and height, as well as the physical position of the site of interest.

SNTHERM has been tested for many applications and in numerous regions including: slush and brine on sea ice in Antarctica (Andreas *et al*., 2004); calibration tests of simpler snow energy balance models in Sweden (Gustafsson *et al*., 2001), and Finland (Koivusalo and Heikinheimo, 1999); verification of bulk aerodynamic estimates of snow energy and mass balance (Suzuki *et al*., 2006); and for use in GCMs (Jin *et al*., 1999; Yang *et al*., 1999).

All authors are in agreement that SNTHERM is a robust, physically based point-energy and mass-balance model that performs well in a wide variety of conditions and applications. As with all other methods used by investigators interested in parameterizing the energy fluxes at the snow surface, SNTHERM tends to estimate latent heat transfers poorly relative to those of sensible heat and radiation (Jin et al., 1999). However, latent heat fluxes are often very small compared to sensible heat and radiative fluxes, and for many applications, this is not an issue. SNTHERM has also been found to underestimate snow surface temperature relative to field measurements, which leads to overestimation of sensible heat fluxes, an important consideration when studying snowmelt in environments where this heat flux dominates (Jin *et al*., 1999; Yang *et al*., 1999; Andreas *et al*., 2004).

Due to the broad applicability of SNTHERM, it was employed to model snow energy and mass balance in this study. The noted performance deficits of this model (estimates of latent and sensible heat transfers) were taken into account by using multiple snow courses and gravimetric measurements to provide a physical baseline for the modelled estimates.

2.5 Climate – Snow Linkages

Snow has the distinction of possessing the highest albedo of any natural surface and thus plays a crucial role in the energy balance of the Earth's climate system. Snow cover and surface climate are linked: the retreat of Northern Hemisphere spring snowpack and the exposure of lower albedo ground cover have resulted in spring temperatures rising faster than any other season (Groisman *et al*., 1994). The accumulation and ablation of the seasonal snowpack in the Northern Hemisphere is the primary driver of the hydrological system at high elevations and latitudes, via the release of the water stored in the snowpack, and the modification of large scale atmospheric circulation (Cohen, 1994; Vavrus, 2008). Additionally, synoptic climate and teleconnections have been linked to snowpack variability in Europe (Bednorz, 2004; Scherrer and Appenzeller, 2006), Siberia (Iijima *et al*., 2007), and Eurasia in its entirety (Clark *et al*., 1999; Bamzai and Shukla, 1999).

More germane to this study, within British Columbia, snow accumulation on glaciers has been found to be controlled by the frequency of circulation patterns associated with heavy precipitation that occur during the fall, winter and spring, with summer melt a result of patterns that bring clear skies, and increased temperatures and solar radiation (Yarnal, 1984). This conclusion has since been investigated further, with particular attention paid to the influence of climate cycles on snow accumulation and melt. The three main patterns that have been identified as exerting a strong influence on annual snowpack fluctuations in British Columbia are the: El Niño Southern Oscillation (ENSO) (Philander, 1990), Pacific Decadal Oscillation (PDO) (Trenberth and Hurrell, 1994), and the Pacific-North America pattern (PNA).

There is wide agreement between studies on the effects these teleconnection patterns have on snow accumulation and melt in western North America and British Columbia. In the western USA, Jin *et al.* (2006) report that the cold ENSO (La Niña) phase generates increased snowpack in the Pacific Northwest, with the negative phase of the PNA producing the same effect, but independent of ENSO. More specifically, in Oregon ENSO is most highly correlated with annual discharge variability, spring snowmelt timing, and magnitude, whereas timing of annual floods is best correlated with the PDO (Beebee and Manga, 2004). In southern Canada, a positive phase of the PNA exerts the strongest influence on snow cover variability, associated with reduced snow cover in western Canada. ENSO produces the same effect but with a much weaker correlation (Brown and Goodison, 1996; Moore and McKendry, 1996; Hsieh and Tang, 2001). La Niña has also been found to result in increased snow accumulation, with more pronounced anomalies than El Niño years – a result of the mid-latitude circulation anomalies during El Niño years being located 35˚ east of those in La Niña years (Clark *et al.*, 2001). A particularly strong signal has also been reported to emerge as elevation increases (Hsieh and Tang, 2001). In general, it can be stated that temperature is higher (lower) in British Columbia and precipitation lower (higher) during El Niño (La Niña) years (Stahl *et al.,* 2006).

The effects of constructive (amplitude of event increases when cycles are in sync) and destructive (amplitude decreases when cycles are out of sync) phasing between the various climate cycles have been explored and the conclusions are consistent across studies. An El Niño (La Niña) occurring during a positive (negative) phase of the PDO is associated with increased (decreased) runoff and temperature in the Columbia River basin (Barton and Ramirez, 2004). In a similar study, positive PDO and PNA phases occurring in tandem were found to be associated with warm, dry winter anomalies in mainland BC, with negative phases possessing the opposite association (Stahl *et al.,* 2006). Links to ENSO were less pronounced, with greater spatial variability, precipitation in particular having a stronger response in interior BC than on the coast. Similar associations have been reported in the Peace River Basin, derived using eigenvector-based map-pattern classification by Romolo *et al*. (2006a). This work was extended to include snow ablation, with positive (negative) phases of the PNA associated with high (low) spring temperatures. La Niña events were found to be significantly correlated to late-melt initiation dates, indicating that this phase of the ENSO pattern not only results in deeper snowpacks in western North America, but greater duration due to cooler than average spring temperatures (Romolo *et al*., 2006b).

Due to the importance of seasonal snowpacks in the functioning of many terrestrial and climatic systems (Selkowitz *et al*., 2002), much effort has been directed at predicting the effects that anthropogenic climate change may have on snowpack variability (Räisänen, 2008; Vavrus, 2008). A consistent conclusion among studies examining long-term variability in North American snowcover and ablation is that over the 20th century; variability has increased, duration has decreased at lower elevations, ablation is occurring earlier and in some cases more rapidly (Dyer and Mote, 2006). These changes have been linked to an increase in spring temperatures (Groisman *et al*., 1994; McCabe and Clark, 2005). However, these trends are weaker in the mountainous areas of western North America; a result of an elevational threshold between solid and liquid winter precipitation. At higher elevations, the increase in temperature is accompanied by an attendant increase in precipitation, thus annual SWE is actually increasing in some high elevation areas (Mote, 2003; Howat and Tulaczyk, 2005). One study that used 20 GCMs to simulate 21st century climate in the N. Hemisphere reported that the threshold between increasing and decreasing mid-winter SWE coincides with the -20˚C isotherm (November - March mean temperature) (Räisänen, 2008). The trends noted above are predicted to continue as climate change progresses; the largest source of uncertainty is the position and relative importance of climatic thresholds (i.e., precipitation-elevation relationships and timing and magnitude of ablation) that are certain to be crossed, and the severity of the associated changes in future water availability (Barnett *et al*., 2005).

In Canada, Karl *et al*. (1993) reported a decrease in the snow to total precipitation ratio south of 55N, and Brown (2000) noted that while winter snow cover extent *increased* in North America over the 20th century, areal SWE decreased significantly in March and April. Again, this change is most likely due to increased rates of warming in the spring relative to other seasons, and the associated dramatic advance of the 0˚C isotherm over the past 20-30 years in western Canada (Bonsal and Prowse, 2003).

The influence of the prevailing seasonal climate on snow accumulation, distribution and ablation was incorporated into this study, by linking the dominant teleconnections to time-series of snow ablation variables.

2.6 Upper Atmosphere – Surface Energy Flux Relationships

The calculations and studies reviewed in Section 2.2 focus on the surface boundary layer, largely confined to the first several metres above the ground surface. One of the main assumptions of the methods used by these studies is that the turbulence generated at the Earth's surface is due to ground obstacle height (roughness length). However, this assumption is violated in areas of high relief topography, and even in areas of homogenous land cover the variables driving ground level energy fluxes are dominated by air mass characteristics (i.e., potential temperature, water vapour and wind profiles) (Granger and Male, 1978; Helgason and Pomeroy, 2005) . Therefore, to understand the synoptic scale conditions that drive energy exchanges at the site scale, the corresponding upper atmosphere conditions must be considered as well.

In New Zealand, robust links between snowpack energy balance and synoptic climate have been made, where Moore and Owens (1984) found that air mass characteristics and regional circulation patterns explained 75% of the variance in sensible and latent heat fluxes during snowmelt. Further work found that the magnitude of snowmelt was strongly influenced by north-westerly storms (net radiation) and anticyclonic circulation (sensible heat transfers via large scale advection) (Neale and Fitzharris, 1997), which supported the conclusions of Prowse and Owens (1982).

Links between snow energy balance and upper-atmosphere conditions in North America have largely focused on synoptic map pattern analyses (e.g. Yarnal, 1984; Romolo *et al*., 2006b), and defining atmospheric water vapour lapse rates for basin scale snow-melt modelling exercises (i.e., Garen and Marks, 2005). Granger and Male (1978) reported that sensible heat flux over a melting snowpack in Saskatchewan was more closely related to the 850 mb geopotential height temperature than near-surface temperatures, but that net radiation was still the dominant melt-inducing flux.

Granger and Male (1978) show that one very useful source of data for water vapour transfers in the atmospheric boundary layer is provided by radiosonde measurements made worldwide at 0000 and 1200 UTC. Although the main purpose of radiosonde observations is to provide data for operational weather forecasting, these data have proven useful in past boundary layer energy and water vapour studies.

The work of Wilfried Brutsaert and his colleagues provides the theoretical and physical foundation for the transfer of temperature, water vapour and wind fields from the atmospheric boundary layer to the surface under all stability conditions (Brutsaert and Mawdsley, 1976). Early work found the similarity functions for water vapour to be smaller than those for sensible heat in Nebraska, but that the method still provided general agreement with ground based flux measurements on a monthly basis (Mawdsley and Brutsaert, 1977). Kustas and Brutsaert (1986) expanded this work to complex terrain, and determined that the roughness height and zero plane displacement values for a hilly area in the Pre-Alps of Switzerland were 2 to 4 orders of magnitude larger than previous studies conducted in flat, homogenous cover terrain, and that the relationships with obstacle height and distribution remained reasonably consistent. The derived regional evaporation was moderately well correlated with the ground estimates, and mechanical turbulence was found to far outweigh convective turbulence (Brutsaert and Kustas, 1987). These methods were further refined, combined with remotely sensed surface temperatures and compared to eddy covariance estimates in Kansas, where high correlations were obtained between the two methods, although evaporation was still underestimated by ~5% (Sugita and Brutsaert, 1991; 1992). Finally, these methods were applied over a mixed forest/crop land cover in rugged terrain, and high correlations between the radiosonde derived evaporation and surface estimates were reported (Brutsaert and Parlange, 1992).

The importance of the overlying atmospheric boundary layer conditions on snow ablation processes was addressed in this research by making statistical linkages between the two during the ablation season of 2007.

2.7 Trends in Hydrologic and Atmospheric Boundary Layer Variables

2.7.1 Atmospheric Boundary Layer

Because the radiosonde archives contain data of high temporal resolution, and in spite of the fact that the data collection methods have changed with operational needs and therefore are not ideal for long term trend identification (Luers and Eskridge, 1998), recent studies have attempted to discern trends in water vapour and temperature in the atmospheric boundary layer. In one of the first studies utilizing these data, the 500 mb geopotential height was analyzed for variations and trends in both its thickness and height over the period 1946-88. A significant positive trend in the 500 mb thickness was found to be positively correlated to the trends in hemispheric mean temperature (Wallace *et al.*, 1993). In a direct hydrological application, Stewart *et al*. (2005) analysed 700 mb height anomalies over western North America, and found positive (negative) anomalies to be associated with earlier (later) centre of hydrograph mass timing.

Ross and Elliot (1996) found significant (and increasing with geopotential height) positive trends in surface – 500 mb precipitable water vapour over the period 1973-93 in the western Hemisphere north of the equator. They also noted an increase in dewpoint that outweighed the concomitant increase in temperature, although spatial variability was higher on the seasonal scale than on an annual basis. The authors expanded this work to include the entire Northern Hemisphere and “change-points” demarcated by sudden shifts in long term means were taken into account (Ross and Elliot, 2001). They report that tropospheric trends for the Northern Hemisphere from 1973-95 show positive trends in surface to 500 mb precipitable water, 850 mb specific humidity, dewpoint and temperature. Water vapour increases are larger, more uniform and more significant over

North America than Eurasia, and are partially attributable to the late 1970s climate shift (PDO) that primarily affected North America. Finally, specific humidity at the 850 mb height showed small increases from 1958-95, with most of the increases occurring since 1973. Trenberth *et al*. (2005) extended this analysis by using the NCEP and ERA-40 data-sets, with the caveat that the water vapour values are suspect over the oceans, but reasonable over land where constrained by radiosonde measurements. The variability in atmospheric water vapour from 1988-2001 was dominated by the evolution of ENSO, particularly by the 1997-98 El Niño event. Recent trends in precipitable water vapour are positive (1988-2003), driven largely by an increase in SST over the corresponding period. A similar study in Greenland found opposing trends at different heights from 1964-2005 - the troposphere was found to be warming, most notably from 1994 onward, while the stratosphere is cooling (Box and Cohen, 2006). This conclusion is supported at a global level by Lanzante *et al*. (2003).

As a direct result of the operational nature of upper atmosphere soundings, the instruments, data reporting standards, and water vapour conversion algorithms vary considerably over time, and across jurisdictions. Radiosonde derived humidity values for the US (Elliot and Gaffen, 1991) and Canada, Europe and the US (Garand *et al*., 1992) were examined for variations in accuracy and precision. The authors found large differences in the low and high range of water vapour values, due largely to differences in instrumentation and quality control procedures. Overall, the Canadian data were found to provide more realistic values of water vapour at the extreme ends of the scale, and particularly in the cold, dry conditions that prevail at high altitudes and latitudes. Gaffen *et al*. (2000) explored the issue of radiosonde data quality and its effect on upper-air temperature trends in two upper-air datasets. Their analysis focused on the sensitivity of identified trends to alteration of the data set at key points where the mean changed abruptly, either at all points, or solely those that coincide with changes in instrumentation. The effect of the latter was to reduce tropospheric warming trends, and overall trends were found to be more sensitive to these adjustments than any other radiosonde data quality issues.

To address some of the above mentioned problems with spatial and temporal inconsistencies in radiosonde data archives, the Integrated Global Radiosonde Archive (IGRA) has been assembled (Durre *et al*., 2006; Durre and Yin, 2008). This data set consists of more than 1500 stations worldwide, and includes measured as well as derived variables describing upper atmosphere conditions on a twice daily interval from the 1960s to present. The data have been subjected to rigorous quality control procedures, and as such this archive is used for the research presented herein.

2.7.2 Hydrologic Shifts

One of the expected results of a warming climate is a shift in hydrologic variables including the date of maximum streamflow, centroid of the hydrograph and annual mean discharge. One of the first such studies in Canada found that spring temperature departures were strongly linked to earlier dates of spring runoff, particularly during the period corresponding to the recent positive phase of the PDO (1977 onwards), although this was not explicitly mentioned (Burn, 1994).

An expansion of this work by Zhang *et al.* (2001) concluded that annual mean streamflow has decreased significantly in the southern regions of the country, with the strongest declines in August and September, whereas March and April discharge exhibited significant increases, particularly in British Columbia. The authors noted that while the earlier warming results in earlier snowmelt, the ablation is more gradual due to the lower solar azimuth and shorter daylight periods. Also, the local significance of these trends was found to be relatively weak, possibly due to the influence of rainfall events on springtime river levels, and because the magnitude of high streamflow events is not increasing. Additionally, in some areas higher spring precipitation still falls as snow at the higher elevations, prolonging the ablation season.

An analysis of streamflow variability in Western Canada found that areas that had high correlations with April, May and June Southern Oscillation Index (SOI), PNA and the Multivariate ENSO Index (MEI) corresponded to areas with significant correlations between winter SOI and annual precipitation (Woo and Thorne, 2003). In the Okanagan, the October-March average PNA and MEI had the strongest correlations with June streamflow. It has been found in two separate studies that shifts towards earlier peak flows and spring pulse onset of 10-30 days from 1948-2002 are common in basins less than 2500 m in elevation (Regonda *et al.*, 2005; Stewart *et al.*, 2005). Increased ENSO activity, and the phase shift of the PDO in 1976 have clearly influenced hydrologic trends in snow dominated basins through their effect on ground level temperatures and precipitation patterns. There is increasing evidence that the long term trends supersede the effects of these decadal climate oscillations and are due to large-scale increases in winter and spring temperatures of 1-3˚C over the past 50 years (Stewart *et al.*, 2005). However, trends in SWE in the western United States associated with precipitation variability have been found to be controlled by decadal climate variability, and the PDO in particular (Hamlet *et al*., 2005).

In general, decreases in SWE and an increase in the rain to snow precipitation ratio are strongest in mid-elevation snow dominated basins in the western North America, regions that lie closest to the annual 0˚C isotherm, and are therefore most sensitive to slight increases in temperature (Knowles *et al*., 2006; Lemke *et al*., 2007). Higher elevation basins show few statistically significant shifts towards earlier peak flows, likely a result of the increased precipitation still falling as snow during the winter, and the moderating effect of spring rainfall events (Hamlet *et al*., 2005; Regonda *et al.*, 2005; Stewart *et al*., 2005). These same findings have been reported in the Swiss Alps by Laternser and Schneebeli (2003), and apply to latitudinal gradients as well (Stewart *et al*. 2004). Future increases in temperature are likely to outweigh increased precipitation falling as snow at higher elevations under a warming climate (Stewart, 2009). Higher winter SWE is strongly associated with delayed snowmelt and centre of mass dates in Canada, and the same is true for the April 1 SWE. The PDO was found to be inversely related to the hydrograph center of mass.

SWE for British Columbia from 1956-2005 were regressed against the ENSO and PDO indices, and the residuals tested for trends (Chapman, 2007). In total, 86% of the snow courses examined showed an average decrease in April 1st SWE of 18% (14-47%). A large portion of this decline is explained by the PDO, however a climate warming trend is still evident, particularly in the dry interior regions of the province, where shallow snowpacks are unable to reabsorb water resulting from mid-winter melt events. The influence of the regionally dominant teleconnections on ablation season processes in the

Okanagan Basin was addressed in this research; the results of which are presented in Chapter 4.

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CHAPTER 3: SPATIAL VARIATION OF SNOWMELT AND SUBLIMATION IN A HIGH-ELEVATION SEMI-ARID BASIN OF WESTERN CANADA[1]

Abstract

The Okanagan Basin, a semi-arid region of western Canada, is currently experiencing rapidly increasing pressure on its water resources from development and population increases, exacerbated by changes in climate. The major source of freshwater in the region originates from the melt of high-elevation snowpacks, about which little is currently known, including the proportion of the peak snowpack lost to sublimation. To better understand the hydrologic regime of this snow resource, a detailed field program was conducted during the 2007 snowmelt season. Specifically, peak annual snow distribution, ablation-season surface-energy exchange and mass balance were measured in a forested high-elevation catchment of the Okanagan Basin. During the snowmelt period, 1-4% of the peak annual snow-water equivalent (SWE) was lost to sublimation in open sites – averaging 0.4 mm d^{-1}. Melt and sublimation rates increased significantly with elevation, and were higher and more variable in the open sites than under forest canopies. The largest sublimation events (>0.25 mm d^{-1}) were associated with low atmospheric vapour pressure, temperatures below 0˚C, and higher than average wind speeds. Condensation occurred under highly stable conditions in the boundary layer when sensible heat fluxes exceeded net radiative inputs to the snow surface. Melt rates were driven almost entirely by sensible heat fluxes and exceeded 30 mm d^{-1} during large-scale advection events. The results from this study will allow water managers to better predict the amount of water available for ecological, agricultural and municipal needs. This work

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also provides the basis for assessing changes in snow surface energetics due to ongoing salvage cutting in forested areas affected by the current mountain pine-beetle outbreak.

Keywords: Snow hydrology; Energy balance; Okanagan Basin; Elevation gradient; Forest Cover; SNTHERM

3.1 Introduction

Seasonal snow cover is integral to the function of ecological, hydrological, geomorphological and climatic systems in cold-regions environments. The melting of this snow recharges soil moisture, aquifers, rivers and lakes, and a variety of aquatic ecosystems (e.g., Gray and Male, 1981). In the case of high-elevation snowpacks of western Canada, they tend to persist on average much longer than those at lower elevations, are deeper, and store the majority of annual precipitation, which is then released over a longer (relative to central and eastern Canada) spring to early summer melt period. This snowmelt runoff is especially important for semi-arid climatic regions including, in particular, the Okanagan Basin within the interior of the western province of British Columbia, Canada. Here, snowmelt is the dominant source of water to lower elevations, specifically during the low-precipitation summer period when agriculture and municipal demands are highest (e.g., Cohen *et al.,* 2006).

The importance of climate-change impacts on snowpacks has been recognized and studied extensively in many cold regions (Lemke *et al*., 2007). While broad-scale trends in spatial and temporal patterns of snow-cover and melt have begun to emerge, there is still a great deal of uncertainty surrounding the nature and magnitude of regional changes. Climate scenarios generated by global circulation models (GCMs) for western North America suggest that as winter temperature and precipitation increase over the next century, less precipitation will fall as snow, the snowmelt season will occur 4-6 weeks earlier, and conditions conducive to snow sublimation/evaporation losses will become more frequent resulting in “considerable” reductions in annual and spring flow volumes (Stewart *et al*., 2004). These projections are consistent with observed rates of change over the past half-century, including a study of streams in the Okanagan Basin (Merritt *et al.,* 2006). These hydrologic changes also include a substantial shift since 1990 towards a 2 days/decade earlier occurrence of the spring 0˚C-isotherm at the Vernon Coldstream Ranch station in the northern portion of the Okanagan Basin (Bonsal and Prowse, 2003). Overall, water resources will experience their greatest annual stress during the periods of highest demand during the growing season, and high-elevation snowpacks will play a proportionately larger role in future water availability.

To date, most scientific literature regarding hydrologic-change in the Okanagan has focused on groundwater distribution and downscaled modelling of climate-change effects on basin-scale water budgets (Merritt *et al*., 2006; Neilson-Welch and Allen, 2007). By contrast, scientific literature and micro-meteorological data concerning the snow energy and mass balance (and sublimation losses in particular) in the higher elevations of the basin are virtually non-existent – a serious gap considering that the hydrology of the Okanagan Basin is dominated by the accumulation and ablation of the snowpack. At the time of writing, no studies have been made of the turbulent and radiative fluxes driving snow ablation processes (melt, sublimation, evaporation and condensation) anywhere in the Basin.

In addition to the effects of changing climate, the Okanagan Basin is experiencing pronounced land-cover changes at high-elevations, some of which have also been ascribed to the effects of changing climate (e.g., Carroll *et al*., 2006). Specifically, mountain pine-beetle salvage operations and forest fires are leading to an increasing number and size of openings in the forest cover. Hence, to fully understand the hydrology of the high-elevation snowpack and how it may change, the effects of forest cover and elevation on the late-winter and early spring snow energy balance needs to be assessed. Obtaining an improved understanding of the physiographic, vegetative and atmospheric controls on the energy and mass balance of snow ablation will provide water managers with invaluable information about the nature of the current water resource and guide adaptive planning for its use under the effects of climate change.

3.2 Background

While snow-sublimation research has been conducted in many different hydroclimatic regimes globally, most of the work to date in Canada has focused on losses due to canopy interception, blowing snow in prairie and sub-arctic areas, and Chinook or föhn wind driven events (e.g., Bernier and Swanson, 1993; Pomeroy *et al*., 1998). To place this work in context, a brief summary of the findings of previous international work is provided below.

In general, snow-surface energy balance is governed by the turbulent exchanges of latent and sensible heat, incoming and outgoing radiation, and small scale advection between (and within) the snowpack, and to a lesser extent by the conducted heat energy from the ground surface and precipitation. The meso-scale spatial variability can also be quite high – even for unforested terrain, which complicates the scaling up of site specific values to an entire basin (e.g., Pomeroy *et al*., 2003). Overall, radiation is the most important contributor to snowmelt at continental, polar and high-elevation sites; sensible heat exchanges dominate during Chinook or föhn winds, and at some high-elevation sites (i.e., Sierra Nevada); and latent heat transfers dominate under conditions of high humidity (including rain-on-snow events) and at maritime sites.

Depending on the specific site and climate characteristics, sublimation accounts for the loss of 15-47% of snow water equivalent (SWE) over a given snow season (Pomeroy *et al.,* 1997; Hood *et al.,* 1999). Sublimation can occur from an in situ snow cover on the ground or one intercepted by vegetation (e.g., Cline, 1997; Lundberg and Halldin, 2001), or when in transit such as during blowing snow events when the time and surface-area exposures of a snow particle are greatly increased (e.g., Liston and Sturm, 2004). The opposite flux of condensation usually dominates at night or under high humidity (i.e., highly stable conditions).

In forested environments, sublimation and melt rates are typically lowest under dense canopies and highest in the openings and increase with opening size and effective fetch length. An increase in wind speed and/or roughness length acts to increase the size of the turbulent eddies developing within the near-surface boundary layer, thereby increasing the magnitude of melt and sublimation rates. Therefore, high magnitude sublimation events in open areas are characterized by relatively low, atmospheric water-vapour concentrations and high wind speeds (Hood *et al*., 1999; Zhang *et al*., 2004).

Given the above noted need to more fully understand regional changes in snowpacks, particularly in semi-arid catchments, and more specifically the dearth of information about sublimation and melt rates from high-elevation snowpacks in the Okanagan region, the main objective of this study is: to quantify snow-energy and mass-balance processes that characterize melt conditions along an elevational and forest-cover gradient in a selected catchment of the Okanagan Basin. The resulting knowledge will provide the basis from which to evaluate historical changes in climatic conditions affecting these snow processes, and highlight potential future changes in a non-stationary system.

3.3 Study Site Description

The Okanagan Basin contains the province of British Columbia's largest agricultural centre; however, the intensive irrigation required to maintain its productivity coupled with its rapidly expanding population make it one of Canada's most water-stressed large catchments. The primary water sources are the tributary streams, most of which are at license capacity and listed as "fully recorded" or "water shortage", which has led to water allocation conflicts, particularly in dry years such as the drought of 2003 (Cohen and Neale, 2006).

Covering an area of 8046 km^2, the Okanagan Basin extends approximately 185 km from north to south and comprises a portion of the northern Columbia River Basin (Figure 1). The region is characterized by a dry continental climate, with precipitation averaging 250-300 mm in the valley, 85% of which has been approximated to be lost to evapotranspiration, and >1000 mm at the higher elevations. Summer rainfall is driven largely by local-scale convection, with winter precipitation influenced primarily by synoptic weather systems originating in the Pacific Ocean. Average annual temperature decreases and precipitation increases along a south to north transect through the basin, with the southern portion containing the only true desert ecosystem in Canada (Cohen and Kulkarni, 2001). SWE-distribution within the Basin follows the same pattern (Figure 2).

The sub-catchment selected for study is the Coldstream Basin, primarily because it has the best set of combined hydrometric and meteorological records available within the larger Okanagan Basin. Of particular importance was the high-elevation access and additional meteorological data available from a ski resort situated above the upper portions of the catchment, and the high-resolution radiosonde record from the nearby Kelowna and Vernon Airports.

Located approximately 20 km east of Vernon, The headwaters of the Coldstream River cover 58.5 km^2 and intersects with the larger Coldstream Valley drainage, which has an area of 205 km^2 and flows into Kalamalka Lake (Figure 1). On average, 76% of the annual discharge is allocated to various water-license holders within the catchment. Kalamalka Lake is situated in the northeast corner of the Okanagan Basin, and in 2007 provided 32% of the nearby city of Vernon's water supply (Cotsworth pers. comm., 2008). The Coldstream Valley is primarily agricultural, and the surrounding slopes are subject to widespread logging, with the current operations focusing on mountain pine-beetle salvage. The dominant tree species at the higher elevations are lodgepole pine (*Pinus contorta*), and Western hemlock (*Tsuga heterophylla*), with replanted Engelmann Spruce (*Picea Engelmannii*), and scattered Western red cedar (*Thuja plicata)* in the low-lying wet areas. Mountain Alder (*Alnus tenuifolia*) is the most common species in recently disturbed areas. The basin and the surrounding area are well instrumented, with a variety of high-quality long-term hydrologic, climatic and radiosonde records available (Table 1). Climate normals for the period 1971-2000 from the Vernon Coldstream Ranch station include a mean air temperature of the coldest month (January) at -5.0˚C, the warmest month (July) 19.1˚C, and the average annual precipitation of 356.5 mm, with

127.9 mm of that falling as snow (Environment Canada, 2008). At higher elevations, the precipitation regime is dominated by winter snowfall, with the snow course at Silver Star Lodge resort recording the highest mean April 1 SWE value in the Okanagan Basin (Figure 2). The historical streamflow record reveals that the Coldstream Creek basin is characterized by an earlier onset to maximum spring discharge (5 days/decade) and lower fall and winter discharge over the last four decades (Whitfield, 2001).

3.4 Methodology

The methods used to meet the objectives of this study involved three separate components: snow courses to assess spatial distribution of SWE and changes in magnitude, a suite of micro-meteorological measurements of melt and vapour fluxes used to model changes in snow mass at selected elevational and vegetation sites, and an intensive set of broad-scale gravimetric measurements of changes in snow mass for comparison to modelled estimates. SNTHERM (Jordan, 1991), a process driven, one-dimensional energy and mass-balance model was employed for the energy-balance calculations of changes in snow thermal structure and ablation. Data were collected during the spring months of 2007, from Feb. 13 – Apr. 19, 2007 (Julian day [JD] 44 – 109), with micro-meteorological measurements beginning on JD 44 and the gravimetric and areal SWE measurements starting somewhat later on March 9 (JD 69). Compared to past years, the spring of 2007 was warmer than the long term average. March-April mean temperature at Coldstream Ranch was 6.9˚C compared to the 20th century average of 4.7˚C, and the April 1 SWE value at Silver Star Resort was 98% of the historical average.

3.4.1 Study Sites

Four sites were selected along an elevational gradient in the Coldstream Basin, situated at 650 m (*Cold Low*), 980 m (*Cold Mid*), 1143 m (*Cold Bonzai*), and 1450 m (*Cold Up*) (Table 1). The *Cold Bonzai* site was added once the *Cold Low* and *Cold Mid* sites lost their snowcover and logistics permitted the augmentation of a higher elevation site. Each site along this transect comprised a representative open and forested site, situated on a south-east aspect to capture the predominant, winter-precipitation storm track. The snow on this aspect was also expected to melt before other aspects, and at a higher rate, thereby enabling upper bounds to be set on high magnitude melt events.

3.4.2 Snow Courses

To obtain high spatial and temporal resolution of SWE distribution and ablation values for all sites, SWE and snow depth were surveyed every 4 days using a Standard Federal 2 m snow tube along snow courses with a maximum of 90 points in each of the open and forested sites. A depth measurement was made every 2 m and a depth and density measurement every 10 m to ensure that the spatial variation in snow cover was captured with a high degree of statistical confidence (Winkler *et al*. 2005). The sites with a lower number of sampling points (due to site and terrain constraints) are *Cold Low* Open (n = 35), *Cold Low* Forest (n = 75), *Cold Mid* Forest (n = 15 – all measurements are depth and density), *Cold Up* Forest (n = 70).

3.4.3 Estimates of Melt and Vapour Fluxes Using SNTHERM

Two open sites (*Cold Mid* and *Cold Up*) were instrumented with automatic weather stations that recorded air temperature and relative humidity with Vaisala HMP45C temperature probes. Net radiation (W m^{-2}) was measured with Kipp and Zonen NRLite net radiometers at a height of 2 m above the snow surface, wind speed (m s^{-1}) and direction (˚) were measured with Texas Electronics TV-110-L320 anemometers and TD-106-5D wind meters at 3.5 m. Snow-temperature profiles were measured with an array of four Campbell Scientific 107B thermistors distributed evenly from the snow to ground surface (Table 1). At *Cold Up*, the thermistors were installed at 0.9 m (near snow surface), 0.6 m, 0.3 m and at the ground surface; at *Cold Mid*: 0.6 m (snow surface), 0.4 m, 0.2m and at the ground surface prior to the onset of snowmelt. All data were collected at 10-second intervals and averaged every hour. Melt and vapour fluxes were estimated using SNTHERM driven by the data collected at these two stations from Julian Day (JD) 46 to 107 (Table 1).

Temperature lapse rates were calculated between each of the stations, and the average lapse rate (no inversions occurred) between *Cold Mid* and *Cold Up* (-0.4 ˚C 100 m^{-1}) was used to infill 46 hours of missing data at the latter. Missing values of wind speed were estimated by multiplying the value from the operating station by the average wind speed ratio (2.1) between these two stations. Missing values of net radiation were infilled with data from the *Cold Mid* station, as both stations showed similar radiation regimes (r^2 = 0.78) during the snow-cover period. The temperature lapse rate was never equal to zero; therefore barometric pressure for *Cold Mid* and *Cold Up* was calculated using the measured pressure at *Cold Ranch* and the elevation of the estimated station (U.S.

Standard Atmosphere, 1976). Missing values of relative humidity were calculated by averaging the difference between recorded values at *Cold Mid* and *Cold Up* ($r^2 = 0.88$) for the same 24-hour period for the previous and following day at an hourly time step. This captured the diurnal cycles of, and relationship in relative humidity between, the two stations. The data from *Cold Mid* were adjusted by this hourly average to estimate the missing RH values at *Cold Up*.

Daily precipitation totals were measured each morning using the lysimeters described below, and these measurements were checked for validity against the precipitation lapse rate calculated from the Silver Star Resort station. Incoming longwave (↓L) and shortwave (↓S) radiation (W m^{-2}) were measured at the *Cold Ranch* station with a Campbell Scientific CNR1 all-wave radiometer. The values of ↓S were adjusted for aspect, elevation and slope angle for application to the *Cold Mid* and *Cold Up* sites, according to methods outlined in Pohl *et al*. (2006). Hourly cloud-cover fraction was estimated following Brock and Arnold (2000) using the ratio of measured incoming to maximum possible ↓S. The estimated cloud-cover fractions were checked against values observed in the field to ensure accuracy. ↓L was estimated using the derived cloud-cover fractions according to Arnold *et al*., (1996), and checked against the measured values at the Coldstream Ranch station.

Site-specific roughness lengths were set in SNTHERM so that the calculated vapour fluxes and melt mirrored the ablation of the snowpack correctly, as measured by the snow courses. The roughness lengths for *Cold Mid* (0.5 m) and *Cold Up* (0.09 m) fall at the upper end and even exceed the roughness lengths summarized by Moore (1983). At *Cold Mid*, this is due to the closely spaced juvenile spruce; here, therefore the roughness length represents the vegetation above the snowpack and not the snow surface directly. At *Cold Up*, the surrounding high relief terrain generates local turbulence that can exceed common values for the snow surface alone by an order of magnitude, as also noted by Helgason and Pomeroy (2005). The model was initialized with temperature and bulk density values for 12 layers within the snowpack, taken from snow pits excavated on the same day as the instrument arrays were installed. All other default parameters in SNTHERM were retained, as they have been found to be robust for application under a wide range of site and climatic conditions (i.e., Fox *et al*., 2008).

The snowpack mass balance (mm d^{-1}) was derived from the preceding energy balance estimates and is defined by the sum of all vapour fluxes (condensation [+ve], evaporation [-ve] and sublimation [-ve]), melt (+ve) and precipitation (+ve), where a negative (positive) value indicates loss (gain) at the snow surface.

3.4.4 Gravimetric Measurements

Due to the assumptions associated with the use of point-scale energy-balance measurements, and the lack of full micro-meteorological equipment under the forest canopy, an array of snow lysimeters was also installed at each site to provide comparative gravimetric data of changes in daily, snow mass balance. Snow vapour losses were measured using 40 clear, 20 cm diameter high-density polyethylene pails filled with a representative sample of the surrounding snowpack and set flush with the surface (Suzuki *et al*., 1999). Five pails were installed at each site – four to measure vapour fluxes and one to provide a control by accounting for trace precipitation. Holes in the base of the four snow-filled pails allowed separation and percolation of melt-water into a catch pail below. The pails were measured once each morning to capture changes in mass over the previous day (including night-time condensation) due to melt and vapour flux transfers; tared; refilled with a representative snow sample (including any layers present); and then buried flush with the snow surface. Occasionally, rapid melt exposed the lysimeter edges, which potentially affected the near surface turbulence regime, and therefore melt and sublimation rates. These data were quality controlled by removing obvious outlier values or when melt exceeded the lysimeter capacities.

Since this study focuses on the ablation season, and because snowfall was minimal during the study period, the surveyed SWE represented the snow available for melt after the preceding winters canopy interception processes had modified the under canopy snowpack. Therefore, the main purpose of the gravimetric measurements was two-fold: first, to provide complementary melt and vapour flux rate data for the SNTHERM estimates; and second, to provide insight into the relative differences in magnitude of these fluxes between open and forested sites of similar elevation, slope and aspect.

3.5 Results and Discussion

3.5.1 Spatial Variation in SWE

Results indicate that the snow energy balance at the *Cold Mid* and *Cold Up* sites switched to a positive overall balance on JD 64. Melt-out occurred first at the lower elevations on JD 84; the open sites were snow free 2-5 days earlier than the forest sites at the same elevation despite higher areal SWE – a negative covariance also noted by Faria *et al*. (2000). Snow cover was extensive at all elevations within the basin at the initiation of snow surveys on JD 65, but varied widely between open and forested sites, and along an elevational gradient (Table 2). The relatively higher ratio of under canopy SWE at the *Cold Up* site might be explained by the more open nature of the forest, lower level of crown closure and thus lower canopy interception efficiency compared to the denser stands at the lower elevation sites. SWE showed a clear linear relationship with elevation at the open and forest sites, with average increases of 22.7 mm 100 m^{-1} and 12.4 mm 100 m^{-1}, respectively ($r^2 = 0.88$). Snow density was lower on average under the forest canopy (280 kg m^{-3}) than in the open sites (339 kg m^{-3}), and did not show a clear elevational trend.

3.5.2 Gravimetric and SNTHERM Estimates of Melt and Vapour Flux – Open Sites

Prior to comparing the results from the gravimetric and modelled approaches, a sensitivity analysis was conducted of each by assessing various sources of potential error in measurement (Table 3). Overall, the potential error associated with the micro-meteorological instrumentation was higher than that for the lysimeter method, and the error associated with estimation of melt (57% cumulative error) was higher than that for vapour fluxes (45%).

Table 4 summarizes the diurnal variation in measured micro-meteorological variables at each site and the correlations between sites. The temperature lapse rate averaged 0.5˚C 100 m^{-1} over the entire study period and was highly correlated between stations, relative humidity was similar at all elevations and also highly correlated between stations, and wind speed increased with elevation, with weak correlations between stations. The wind speed value at the *Cold Ranch* station was higher than at *Cold Mid*, which was contrary to what would be expected given that wind speeds typically increase with elevation. This was due to the sensors higher placement relative to the other two stations (10 m vs 3.5 m above the surface), and night-time katabatic winds. Net radiation differed substantially between sites and is poorly correlated, with the exception of the *Cold Mid* and *Cold Up* sites ($r^2 = 0.78$).

Daily totals of hourly values for the two methods are plotted in Figure 3 and show good correspondence for both melt and vapour flux values at both sites. Regression lines offer better explanation for vapour loss ($r^2 = 0.37$) than melt rates ($r^2 = 0.25$) at *Cold Mid*, with gravimetric values consistently overestimating those modelled by SNTHERM (Figure 3a, b). Possible causes include incomplete drainage of melt water from the lysimeters and/or advection influences from nearby exposed vegetation - a complicating factor that is not parameterized in SNTHERM (e.g., Neumann and Marsh, 1998). At the *Cold Up* open site; the regression line explained more of the variance for melt than for vapour fluxes ($r^2 = 0.44$ and 0.23), respectively (Figure 3c, d). The two circled values in Figure 3d correspond to the last two days of the study period, when the snow had partially melted out around the pails, increasing the local advective energy transfers. The upper circled value in Figure 3c corresponds to JD 99, when the measured melt was much higher than the estimate. Notably, such intense melt was produced by large-scale advection of a warm air mass and strong southerly winds (see below). The lower, earlier (JD 71) circled value resulted from high day-time melt rates followed by cold night-time temperatures that caused water to freeze in the drain holes of the lysimeter, thereby artificially decreasing the amount of “measured” melt. Removal of the above outliers resulted in closer agreement between the methods for melt ($r^2 = 0.71$; Figure 3e). The relationship between measured and modelled melt differed between sites. This could be an artifact of lower diurnal variation in the snow-surface energy balance at *Cold Mid* due to the surrounding trees moderating the radiation balance.

Ablation season melt and vapour losses were summed for each method, and compared to the measured SWE for both instrumented sites. At *Cold Up (Cold Mid)*, the modelled SNTHERM and gravimetric changes accounted for 92% (100%) and 88% (106%) of the surveyed changes in SWE. Figure 4 shows the full-season changes in snow mass based on daily values for both techniques progressively subtracted from an initial SWE value for each site. The ablation curves closely mirror each other, except for the forested sites of *Cold Low* and *Cold Mid*, where the snowpacks had much lower average densities than the open sites (286 vs. 337 kg m^{-3}). The greater disparity at these sites may be due to the difficulties in obtaining spatially representative samples under forest canopies (e.g., Winkler and Moore, 2006; Ellis and Pomeroy, 2007). The highest level of similarity among the three sets of values was found for the open site at *Cold Up,* although even the forested site showed good consistency between the snow course and gravimetric measurements. This latter site, however, also showed pronounced divergence among results after the snow cover became patchy (~JD 101); specifically, gravimetric values were noticeably lower than those of the areal SWE. This highlights the difficulty of accurately up-scaling point gravimetric measurements of snow ablation to the meso-scale, particularly under conditions where local-scale advection prevails.

Although both modelled and measured results were highly comparable, the gravimetric sublimation values were occasionally overestimated. This was due to incomplete melt-water drainage over the entire study period, and local-scale advection during periods of patchy snow cover.

3.5.3 Spatial Variation of Melt and Vapour Fluxes

Based on the above results, the SNTHERM values were used to define vapour fluxes and melt for the *Cold Up* and *Cold Mid* open sites, and the gravimetric measurements for all other sites. Overall, melt rates averaged 6.5 (max. 40.5) mm d^{-1} at the open sites of all elevations, but only 4.0 (max. 27.4) mm d^{-1} under forest canopy. Similarly, sublimation was also greatest in open sites averaging 0.4 (max. 0.83) mm d^{-1} and approximately half, 0.2 (max. 1.06) mm d^{-1}, at the forest sites, the maximums both occurring at *Cold Up*. Total seasonal sublimation for open and forest sites were: *Cold Up* (8 and 17 mm), *Cold Mid* (1.7 mm, and 0.4 mm condensation under the canopy) and *Cold Low* (2.6 and 6.8 mm). This accounts for a 1-4% loss of peak SWE in the open sites, and 4-12% at the forested sites (Table 5). These values are within the ranges reported by other authors (summarized in Table 6) and similar in magnitude to that for the Austrian Alps (Kaser, 1981), the foothills of the Alberta Rockies (Bernier and Swanson, 1993), and the boreal forest and taiga of Siberia (Zhang *et al.*, 2004; 2008). Total SWE losses to sublimation would likely be higher over the full winter season, as this process was consistent prior to the initiation of snow melt, and most condensation events noted occurred after the onset of ablation. Overall, sublimation was by far the dominant vapour flux over the ablation period, with condensation events infrequent and of lower magnitude. Evaporation of free water from the surface of melting snow grains did occur with low frequencies and magnitude.

To determine whether values of vapour flux and melt differed significantly between open and forested areas at the same elevation and between areas with similar forest cover at different elevations, Mann-Whitney U tests ($\alpha = 0.05$) were used. The results indicate that *Cold Up* was the only site showing a significant difference in vapour fluxes between open area and forest cover, and that these losses increased significantly with elevation for both open and forested sites. Melt rates also increased significantly with elevation, along with net radiation and wind speed, but did not differ significantly between forest and open sites at the same elevation. Other studies have also found that orographic effects play an important role, as wind speed and solar radiation vary with slope angle, elevation and aspect, leading to relatively higher sublimation and melt rates on steep south facing slopes, and at higher elevations (e.g., Pomeroy *et al*., 2003; Zhang *et al*., 2004).

Figure 5 shows the average daily melt and vapour flux values for all sites in Coldstream Basin. In general, loss of SWE to melt was greater and more variable at the open sites at high-elevations, while sublimation showed greater spatial variability. The forest at the *Cold Mid* site experienced an average net gain of water vapour through condensation over the course of the ablation period, again, likely an artifact of the terrain and microclimate at this site, as it was situated in a slight depression, was substantially colder than the open site (by an average of 2-3˚C) and sheltered from the wind. The higher sublimation values at the *Cold Up* forest site are likely a result of its open canopy. Similar relationships have been reported for sublimation in clearings of varying sizes (Golding, 1978) and forest density (Suzuki *et al*., 1999).

There is general agreement that higher sublimation and melt rates in open locations are driven by greater exchanges of latent and sensible heat, and higher net radiation. Because forest canopy effectively reduces wind speed, turbulent fluxes are negatively correlated with stand density (e.g., Doty and Johnson, 1969; Zhang *et al.*, 2004). However, canopy effects on net radiation also play an important role. For example, a study of variations in snowmelt energy balance with forest density found that while the above held true for sensible heat fluxes, latent heat exchanges increased slightly (Suzuki *et al.*, 1999).

As noted above, the highest sublimation rates generally occurred in open sites and at higher elevations. The percentage of maximum seasonal SWE lost to sublimation is given for each site in Table 5. At the *Cold Low* and *Cold Up* sites, the percentage of SWE lost to this process was significantly higher under canopy than in the open (12% vs. 3% and 11% vs. 4% respectively). The open *Cold Mid* site lost the smallest portion of its SWE to sublimation (1%). While sublimation from the snow surface was an important process in the Coldstream Basin, its magnitude was outweighed by greater SWE values in the openings compared to under forest canopies. This disparity increased with elevation, as there was 63% more SWE at *Cold Up* in the open, but only 7 mm more was lost to sublimation. However, this relationship is subject to change in the future with the increasing size and number of pine beetle salvage clear-cuts, as the balance between higher accumulation, longer fetch lengths and higher wind speeds will increase blowing snow sublimation losses in mid-winter.

3.5.4 Snow Energy Balance – Open Sites

The study period was characterised by a sublimation-dominated vapour flux regime at both the instrumented sites, with some diurnal variation. Before JD 64, diurnal (12 hour) fluxes of net radiation and latent heat were largely negative, and sensible heat transfers were positive (Figure 6a, b). The magnitude of upward latent heat fluxes was much greater at *Cold Up* (15.4 W m^{-2}) than at *Cold Mid* (3.2 W m^{-2}), a result of lower atmospheric vapour pressure and higher wind speeds at *Cold Up*. After JD 64, the ratio of net radiation to sensible heat directed at the snow surface increased with time, and latent heat transfers showed greater temporal variation, particularly at *Cold Mid*. During periods of high sensible heat transfer to the snowpack, latent heat transfers were reduced in magnitude and remained negative, with the exception of JD 71 at *Cold Up*, where free melt-water at the snow surface evaporated. The final days of the ablation period were characterised by relatively higher net radiation and lower sensible heat fluxes; small latent heat transfers and high snowmelt rates. These conditions became more pronounced as ablation progressed and snow cover became patchy, reflected incoming solar radiation decreased, and longwave radiation re-emitted from bare soil increased. Overall, sensible heat fluxes accounted for the majority of snow melt, but as day lengths increased throughout the ablation season, the relative proportion of the energy balance composed of radiative inputs increased. When highly stable conditions prevailed in the surface boundary layer and sensible heat transfers dominated (avg. 95 W m^{-2}) relative to small (or negative) net radiation fluxes, the latent heat flux was positive and condensation occurred (Figure 6a, b). The latter condition was more common at the *Cold Mid* site than at the *Cold Up* site; night-time condensation events occurred 48 and 29% of the time, respectively. Overall, the net vapour flux at the snow surface was negative, as sublimation events were more frequent and of higher magnitude than condensation events. This relationship between the relative magnitudes of the important energy fluxes has been noted in previous studies of snowmelt energy balance (Granger and Male, 1978; Suzuki *et al*., 2006).

3.5.5 Classification of Melt and Vapour Fluxes

To identify the meteorological conditions that resulted in high magnitude melt and sublimation events, these fluxes were summed on a daily time step and grouped according to magnitude (Figure 7). The measurements taken after areal snow cover dropped below 80% were excluded, as this was the threshold where local scale advection began to significantly alter the snow energy balance. Melt values were strongly positively skewed towards the lower magnitude categories, except for the *Cold Low* site which approached a normal distribution. The most common daily melt rate following initiation of snowpack ablation was 0-5 mm d^{-1} in both the open and forested sites (Figure 7). The exceptions were the *Cold Low* and *Cold Up* sites, where the most common melt rates were 5-10 mm d^{-1}, and 0 mm d^{-1}, respectively. The degree of skewness at each site is due to elevational influences, as warmer temperatures occurred earlier at the lower sites, thereby reducing the relative number of non-melt days at the lower sites during the study period.

Sublimation events at the two low elevation sites were generally of low magnitude (-0.25 – 0 mm d^{-1}), while the next larger vapour flux class (-0.50 -0.25 mm d^{-1}) dominated at the two high-elevation sites. At the lower sites, the distribution is slightly negatively skewed towards low sublimation rates and occasional condensation events, while sublimation losses prevailed at the upper elevations of the basin, particularly as the period of record increased (i.e. *Cold Up*; Figure 7).

3.5.6 Elevational Gradients of Climatic Variables during High Magnitude Events

The highest magnitude sublimation (>0.25 mm d^{-1}, n=6) and melt (>10 mm d^{-1}, n=9) events were defined as days that showed consistently high SWE losses occurring throughout the entire basin. Elevational gradients were plotted to explore the noted increases in melt and sublimation rates with increasing elevation, and to highlight the different atmospheric boundary layer (ABL) conditions associated with these snowpack mass-balance changes. The values from *Cold Ranch* are included for reference, although snow cover was not present at this site for most of the study period. The ABL conditions associated with high melt rates differ markedly from those associated with high rates of sublimation (Figure 8). During large melt events, air temperature and vapour pressure at *Cold Mid* (*Cold Up*) averaged 6.7˚C (5.6˚C) and 0.69 kPa (0.62 kPa). Air temperature during large sublimation events at *Cold Mid* (*Cold Up*) averaged 1.3˚C (-1.0˚C), while vapour pressure averaged 0.42 kPa (0.36 kPa), respectively. Air temperature gradients differed between event types, decreasing an average of -0.26 ˚C 100 m^{-1} during melt events and -0.47 ˚C 100 m^{-1} during sublimation events. On average, vapour pressure followed the same gradient for both event types, decreasing -0.02 kPa 100 m^{-1} over the entire elevational range of the basin. Overall, melt events in the Coldstream Basin were associated with warm and relatively high moisture content air masses. Sublimation events were characterized by much colder and drier air masses, and temperature decreased more rapidly with increases in elevation during these events than during large melt events.

Figure 9 shows the resultant wind vectors at three elevations in the basin during large melt events. The lowest (*Cold Ranch*) and highest (*Cold Up*) sites are both easterly, while *Cold Mid* is southwesterly, a marked difference that highlights the topographical influences on wind direction within the basin. Wind speeds show a similar pattern, with the highest and lowest sites experiencing higher wind speeds than the mid-elevation site. Cold Mid is located on an eastern aspect, but is in the lee of a much higher ridgeline running north-south; thus, the wind direction here might be influenced by an eddy effect from the nearby ridge. Figure 10 shows a similar pattern for sublimation at the lower two sites, but the resultant vector at *Cold Up* indicates that these events were characterized by southerly winds instead of the easterlies associated with melt in the basin.

While these gradients and values provide an overview of the basin-scale atmospheric boundary layer conditions that drive high magnitude melt and vapour loss events, they provide little information about the associated larger scale synoptic conditions. Future research should attempt to identify the linkages between site-scale snow surface processes, conditions in the regional atmospheric boundary layer, and the driving synoptic patterns. In addition, because this study is limited to a single year of data, future work should attempt to place the 2007 study period in as long-term historical context as possible. Of primary interest is whether these high magnitude events are increasing in magnitude and/or frequency concurrent with the rapid changes in the climate and forest cover of the Okanagan Basin.

3.6 Conclusions

Snow energy and mass-balance processes were quantified along an elevational and forest cover gradient in the Coldstream Basin, British Columbia during the late-winter and spring of 2007. Within the basin, peak annual SWE increased linearly with elevation and was typically reduced by half under forest canopy relative to the paired open site at the same elevation. Snow density increased (decreased) with elevation at the forest (open) sites. All measures of snow mass balance provided similar results until the areal snow cover fraction dropped below 80%, at which point small scale advection began to dominate the snow energy balance. Over the 45 day ablation season, melt averaged 6.5 mm d^{-1} in the open sites and 4.0 mm d^{-1} under forest canopies, while sublimation averaged 0.4 mm d^{-1} in the open sites and 0.2 mm d^{-1} at the forest sites. This accounted for a 1-4% loss of peak SWE at the open sites, and 4-12% at the forested sites. Sublimation and melt rates increased significantly with elevation, but these processes showed no statistically significant difference between the open and forested sites at the same elevation. Condensation events were common later in the study period, but of a lower magnitude and frequency than sublimation events. At the higher elevations, greater snow accumulation in the open outweighed the higher rates of sublimation; however, this balance can be expected to shift as land cover changes in the future.

During the early portion of the ablation season, high snowmelt rates were driven by large sensible heat fluxes and relatively lower net radiation. This balance shifted as the ablation season progressed and higher solar angles increased the proportion of shortwave radiation directed at the snow surface at the higher elevation sites. Sensible heat was found to be the predominant driver of snow melt for this study period, far outweighing the influence of incoming radiation or latent heat transfers to the snowpack. High magnitude melt events were characterized by warm, moist air masses and high easterly winds, while large vapour loss events were driven by colder, drier conditions and higher than average southerly winds.

The results of this study are in broad agreement with those conducted elsewhere, and lay the foundation for future snow energy and mass-balance research in the Okanagan Basin. Further work is needed to discern whether the relative differences in melt and sublimation rates associated with elevation gradients and forest cover hold true in the drier and less vegetated southern portions of the Basin. Of particular interest in this Basin, indeed in the whole of western Canada, is the shift towards larger open areas in forests due to the ongoing mountain pine-beetle salvage cutting. This will likely result in pronounced changes to the snow energy balance, as increased fetch lengths result in higher magnitudes of the turbulent fluxes, leading to increased sublimation losses. Since this paper presents the findings of a single study season, a natural next step would be the initiation of a long-term study of several high-elevation sites within the Okanagan Basin, and to define a methodology for placing the study period described here in a historical context. The relationships presented here between snow energy and mass balance, and the atmospheric boundary layer conditions could be expanded with the use of nearby radiosonde records to provide a useful proxy record of the historical climatic drivers of snow mass and energy-balance processes.

This work provides a clearer understanding of the climatic and physiographic variables that affect the timing and volume of the spring snowmelt and subsequent freshet that is vital to the replenishment of the Okanagan Basins surface and groundwater resources. The wide range in magnitude of events recorded will also allow more accurate predictions of the potential effects of changing climate and land cover on this already highly stressed resource, and whether current high-magnitude, low-frequency sublimation and melt events can be expected to increase in both magnitude and frequency in the future.

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Figure 10: Wind roses for (a) *Cold Ranch*, (b) *Cold Mid* and (c) *Cold Up* during high magnitude vapour loss events (>0.75 mm d^{-1}).

Table 1

Site	Latitude (North)	Longitude (West)	Elevation (m)	Record Period (JD)	Variables Measured	Forest Cover
Coldstream Ranch	50˚ 13'	119˚ 11'	482	46-107	Wind speed & dir., Temp., RH, ↓ & ↑ LW and SW radiation	N/A
Cold Low - Open	50˚ 15.852'	119˚ 05.262'	651	69-85	Gr. & SC	N/A, east aspect
Cold Low - Forest	50˚ 15.859'	119˚ 05.272'	658	69-89	Gr. & SC	2nd growth Hm & Wrc, no understory, moderate crown closure
Cold Mid - Open	50˚ 16.806'	119˚ 06.029'	980	69-86	Wind speed & dir., air temp., RH, snow temp. profile, net radiation, Gr., SC	Es saplings, 2-3 m spacing
Cold Mid - Forest	50˚ 16.851'	119˚ 06.038'	970	69-92	Gr. & SC	2nd growth Hm & Wrc, no understory, full crown closure
Cold Bonzai - Open	50˚ 18.270'	119˚ 05.036'	1185	87-104	Gr. & SC	Es saplings, 4 m spacing
Cold Bonzai - Forest	50˚ 18.218'	119˚ 05.022'	1186	87-95	Gr. & SC	Mature Hm & Wrc, Ma
Cold Up - Open		119˚ 05.915'	1450	44-108	Wind speed & dir., air temp., RH, snow temp. profile, net radiation, Gr., SC	Seedling Es, some Ma
Cold Up - Forest	50˚ 19.225'	119˚ 05.965'	1404	69-108	Gr. & SC	Mature Hm, Wrc, Pl, Ma mod. crown closure

Table 2

	Open Sites		*Forested Sites*		*Forest:Open*	
	SWE (mm)	Density (%)	SWE (mm)	Density (%)	% SWE	% Density
Cold Low	96	36.5	55	28.3	58	78
Cold Mid	118	32.2	59	29.0	50	90
Cold Bonzai	172	32.8	86	31.4	50	96
Cold Up	273	28.8	178	29.1	63	101
Average	165	32.6	95	29.5	55	90

Table 3

Instrument	Error	Melt RMSE	Melt % orig. Avg.	Vapour Flux RMSE	Vapour Flux % orig. Avg.
HMP45C Temperature	+/-0.2C at 20C, +/- 0.3C at 0C	0.08	31.4	0.0021	-15.3
HMP45C Humidity	+/-2% RH (0-90%), +/-3% RH (90-100%) - at 20˚C	0.062	31.7	0.0023	-24.8
107B Temperature Probe	+/-0.3˚C	0.062	22.1	0.001	8.3
CNR1 Net Radiometer	+/-2.0˚C, max. 25 W m^{-2} directional error	Snowpack melted by JD 73			
TV-110-L320 Wind Speed	+/-0.25 m/s < 5.0 m/s	0.132	37.3	0.0037	12.7
TD-106-5D Wind Direction	+/- 3˚				
Cumulative error		0.193	56.8	0.006	45.2
Mt. Rose Snow Tube	+/- 10%				
MXX-2001 balance	+/-0.1g	0.0027	0.038	0.0027	0.018
Lysimeter pails (HDPE)	+/- 0.037g (water absorption)	0.00098	0.014	0.00098	0.007
Cumulative error		0.0037	0.052	0.0037	0.025

Table 4

	Air Temperature (˚C)			*Relative Humidity (%)*			*Net Radiation (W m^{-2})*			*Wind Speed (m s^{-1})*		
Day	Cold Ranch	Cold Mid	Cold Up	Cold Ranch	Cold Mid	Cold Up	Cold Ranch	Cold Mid	Cold Up	Cold Ranch	Cold Mid	Cold Up
Average	6.20	3.83	1.10	66.25	64.32	68.38	-30.30	67.78	25.02	1.90	1.20	2.19
Maximum	16.68	14.64	13.44	100.00	96.42	100.87	7.01	209.31	255.98	4.56	2.05	4.68
Minimum	-1.98	-3.92	-7.14	37.92	38.88	36.87	-72.24	-33.05	-92.81	0.60	0.37	0.80
Std. Dev.	4.43	4.07	4.14	21.22	18.57	19.05	22.81	73.70	73.99	0.73	0.37	0.83
Night												
Average	2.54	0.54	-1.06	81.63	79.85	78.46	123.26	-28.84	-26.23	2.29	1.61	1.90
Maximum	10.87	9.31	9.23	100.00	99.56	102.01	287.19	1.55	1.75	4.26	3.21	5.34
Minimum	-6.72	-8.46	-9.24	46.75	51.70	40.17	2.39	-69.28	-71.74	0.19	0.19	0.40
Std. Dev.	3.50	3.39	3.93	14.59	14.19	16.66	80.78	18.41	18.21	0.95	0.60	0.98
Cold Ranch	--	**0.96**	**0.88**	--	**0.91**	**0.85**	--	**-0.51**	**-0.47**	--	**0.52**	**0.09**
Cold Mid	--	--	**0.95**	--	--	**0.93**	--	--	**0.78**	--	--	**0.04**
Cold Up	--	--	--	--	--	--	--	--	--	--	--	--

Table 5

	Melt			Sublimation			% of Max.
	Max.	**Min.**	**Avg.**	**Max.**	**Min.**	**Avg.**	**SWE**
	Open Sites						
Cold Low	16.30	0.21	6.26	0.37	-0.75	-0.16	2.68
Cold Mid	40.45	0.00	3.94	0.48	-0.45	-0.03	1.43
Cold Bonzai	42.71	0.02	9.98	0.67	-1.18	-0.41	4.28
Cold Up	34.95	0.00	5.22	0.41	-0.83	-0.14	3.91
	Forested Sites						
Cold Low	13.26	0.01	5.99	0.41	-0.39	-0.10	12.41
Cold Mid	10.11	0.01	3.28	0.75	-0.44	0.02	-0.75
Cold Bonzai	7.81	1.04	2.79	-0.20	-0.89	-0.41	4.29
Cold Up	27.35	0.00	3.91	0.18	-1.06	-0.45	10.89

Table 6

Study	Methods	Site Type	Sublimation/Evaporation
Bernier and Swanson, 1993	B.A., Gr	Forest and Open	0.25 - 1.07 mm d^{-1}
Doty and Johnston, 1969	Gr	Open	0.15 (Jan) - 1.56 (April) mm d^{-1}
Fassnacht, 2004	B.A.	various sites in USA	0.23 - 0.67 mm d^{-1}
Fox *et al*., 2008	SNTHERM	Glacier	Negligible
Golding, 1978	B.A.	Sub-alpine forest	1.2 mm d^{-1} (1975), 2.0 mm d^{-1} (1976)
Gustafsson *et al*., 2001	SNTHERM	Open fields	8% of total SWE losses
Hood *et al.,* 1999	A.P.	Alpine	0.9 - 1.8 mm d^{-1}
Kaitera and Teräsvirta, 1972	B.A.	Boreal forest	0.35 (sub-canopy) - 0.45 (open) mm d^{-1}
Kaser, 1982	Gr	Alpine	Mean 0.25 mm d^{-1}, max 2.0 mm d^{-1}
Koivusalo and Heikinheimo, 1999	SNTHERM, UEB	Boreal forest	Negligible
Liston and Sturm, 2004	Review	Tundra/arctic	35-50% of annual arctic precipitation
Lundberg and Koivusalo, 2003	SC	Boreal forest	11 - 30% intercepted snow
Marks and Dozier, 1992	B.A.	Alpine	Mean 2 mm d^{-1}
Martinelli, 1960	Gr	Alpine	+/- 0.67 mm d^{-1}
Meiman and Grant, 1974		Alpine, forest, open	45 - 60% snow season precipitation
Molotch *et al.,* 2007	E.C.	Sub-alpine forest	0.41 - 0.71 mm d^{-1}
Nakai *et al*., 1999a	E.C.	Boreal forest canopy	1.2 mm d^{-1} (snow covered)
Nakai *et al*., 1999b	E.C.	Boreal forest canopy	0.6 mm d^{-1} (n = 60)
Parviainen and Pomeroy, 2000	B.A., E.C.	Boreal forest	0.16 - 0.72 mm d^{-1}
Pomeroy *et al.,* 1997	SC, model	Forest/tundra	37 mm/winter (28% total SWE)
Pomeroy *et al*., 1998	B.A.	Boreal forest	0.41 - 1.88 mm d^{-1}
Pomeroy and Essery, 1999	E.C.	Prairie	1.8 mm d^{-1}
Rylov, 1969	Gr	Open (semi-arid)	0.08 (Jan) - 0.6 (April) mm d^{-1}
Schmidt *et al.,* 1998	B.A., Gr	Sub-alpine forest	0.61 (south), 0.43 (north) mm d^{-1}
Storck *et al*., 2002	Gr, SC	Sub-alpine forest	100 mm/winter, <1 mm d^{-1}
Suzuki *et al*., 1999	B.A., Gr	Larch forest	0.55 - 0.96 mm d^{-1}
Suzuki *et al.,* 2006	B.A., SNTHERM	Taiga, larch forest	1.0 (forest), 2.0 (open) mm d^{-1}
Zhang *et al*., 2003	B.A., Gr	Taiga	0.2 - 1.0 mm d^{-1}
Zhang *et al.,* 2004	B.A., Gr	Taiga, larch forest	0.22 - 0.32 mm d-1
Zhang *et al.,* 2008	A.P., E.C.	Flat plain, boreal forest	0 - 1.2 mm d^{-1} (~20% total SWE)
			4.8 (forest), 5.9 (open) mm d^{-1}
		Aspen stand	0.12 (Jan) - 1.05 (April) mm d^{-1}
		Conifer stand	0.33 (Feb) - 0.48 (April) mm d^{-1}

Figure 1

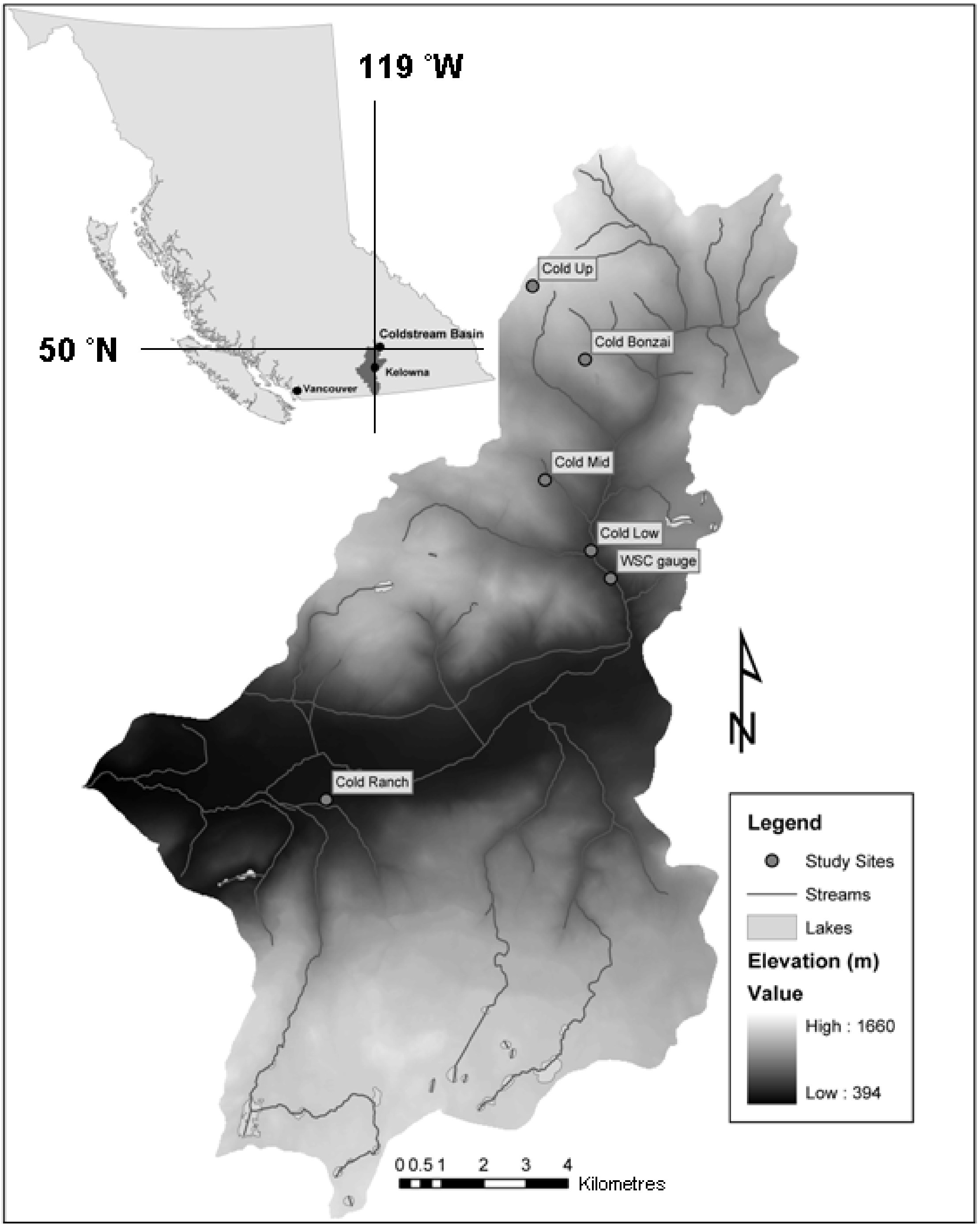

119 °W
50 °N
Coldstream Basin
Kelowna
Vancouver
Cold Up
Cold Bonzai
Cold Mid
Cold Low
WSC gauge
Cold Ranch
N
Legend
Study Sites
Streams
Lakes
Elevation (m)
Value
High : 1660
Low : 394
0 0.5 1 2 3 4
Kilometres

Figure 2

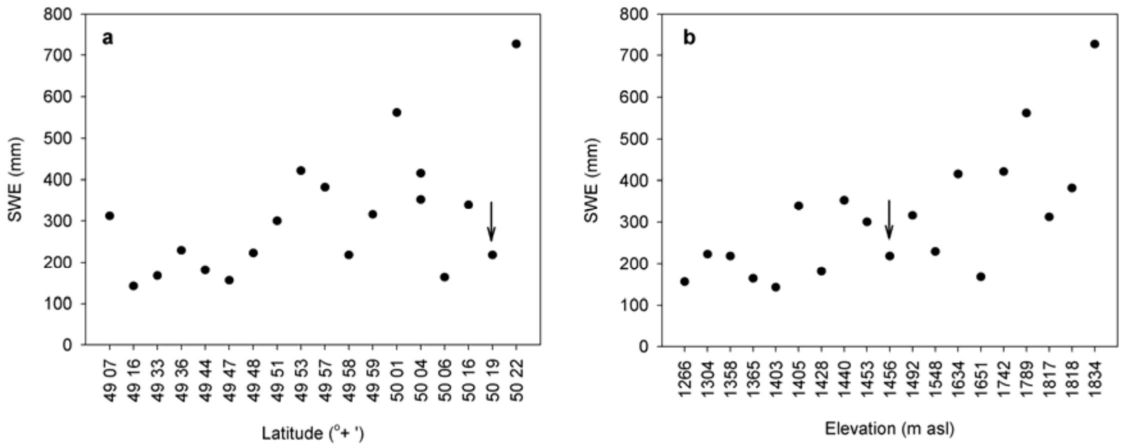
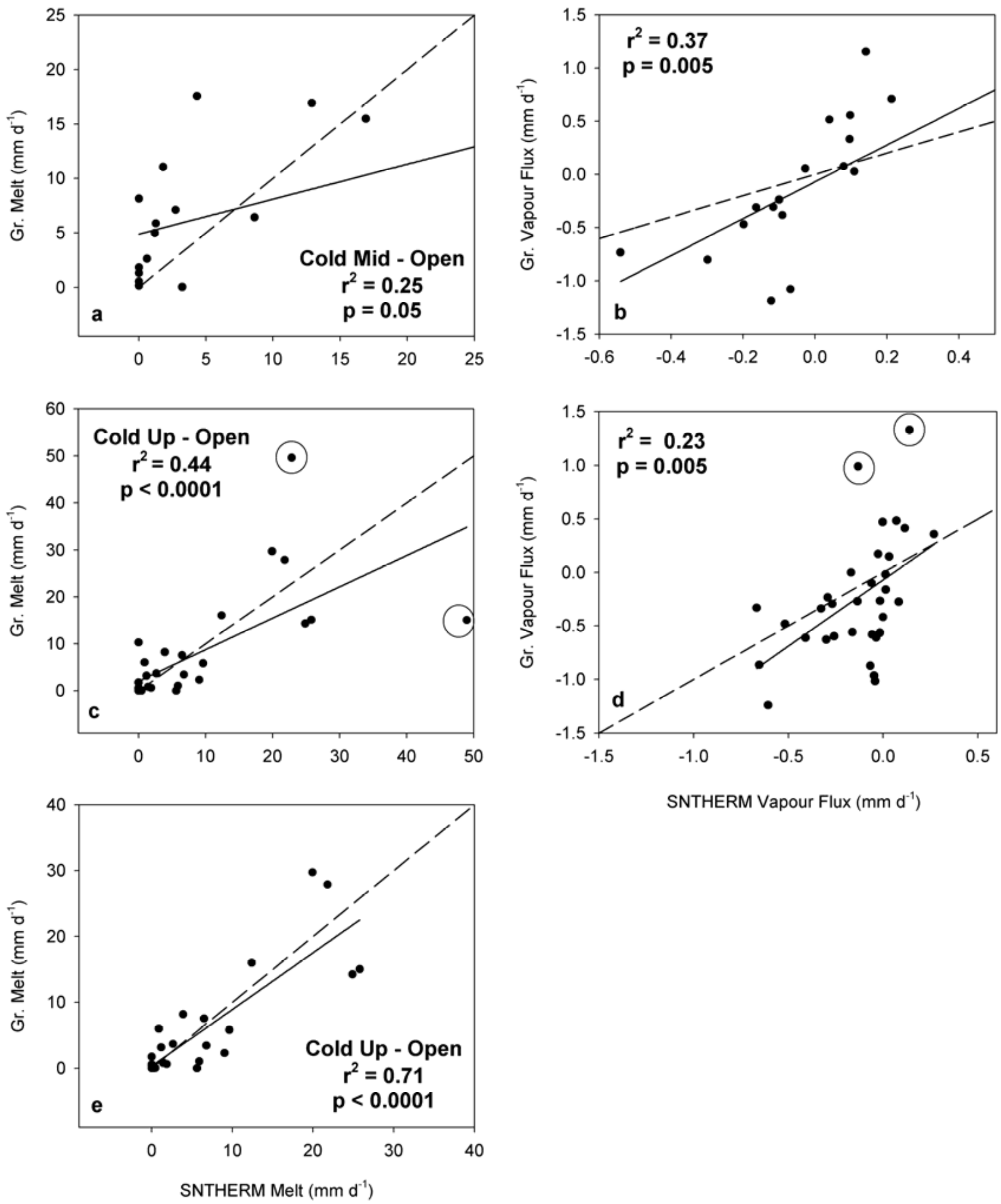

a
b
SWE (mm)
800
700
600
500
400
300
200
100
0
49 07
49 16
49 33
49 36
49 44
49 47
49 48
49 51
49 53
49 57
49 58
49 59
50 01
50 04
50 06
50 16
50 19
50 22
Latitude (°+ ')
1266
1304
1358
1365
1403
1405
1428
1440
1453
1456
1492
1548
1634
1651
1742
1789
1817
1818
1834
Elevation (m asl)

Figure 3



Cold Mid - Open
$r^2 = 0.25$
p = 0.05
a
Gr. Melt (mm d^{-1})
$r^2 = 0.37$
p = 0.005
b
Gr. Vapour Flux (mm d^{-1})
Cold Up - Open
$r^2 = 0.44$
p < 0.0001
c
Gr. Melt (mm d^{-1})
$r^2 = 0.23$
p = 0.005
d
Gr. Vapour Flux (mm d^{-1})
SNTHERM Vapour Flux (mm d^{-1})
Cold Up - Open
$r^2 = 0.71$
p < 0.0001
e
Gr. Melt (mm d^{-1})
SNTHERM Melt (mm d^{-1})

Figure 4

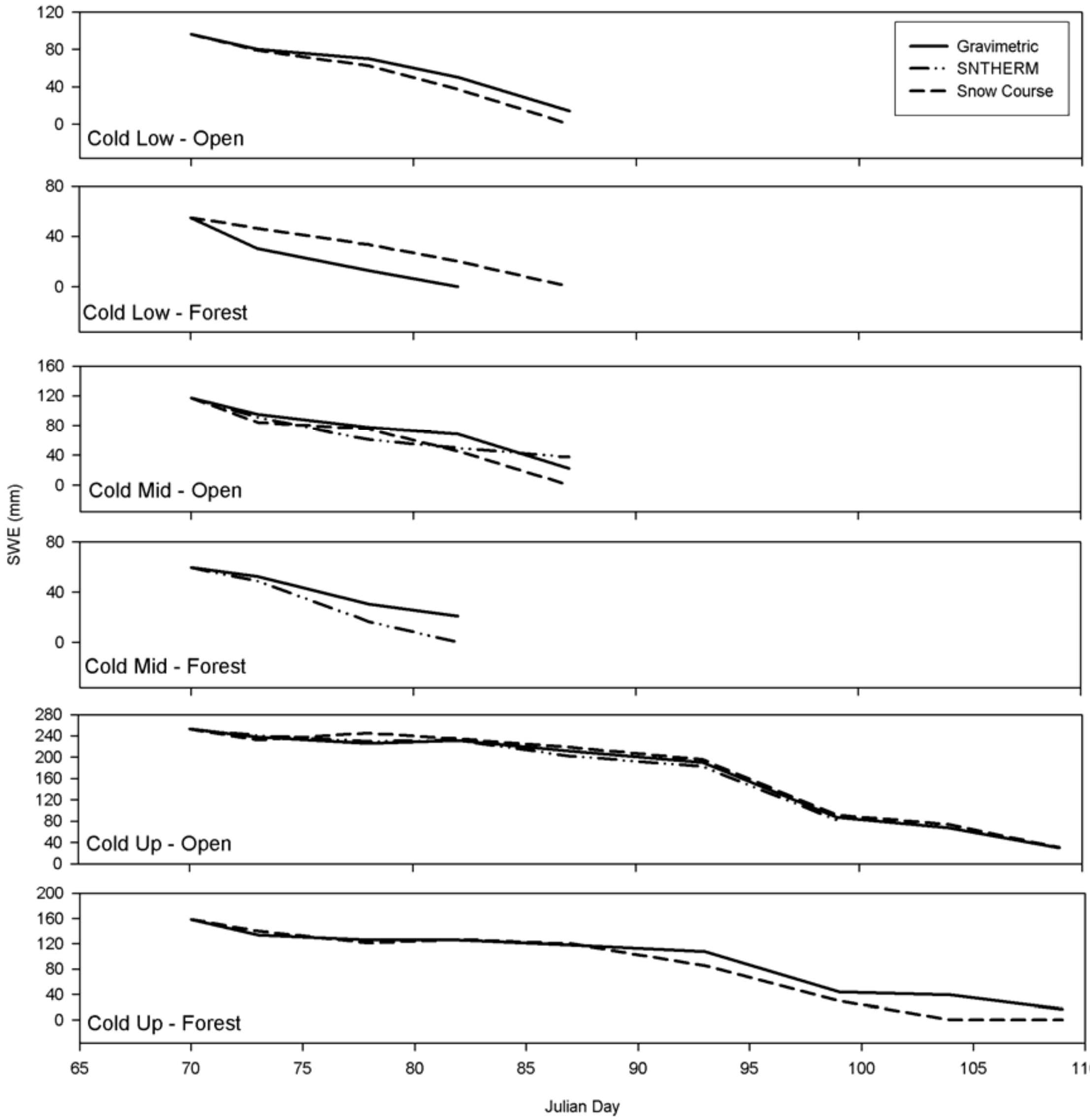

Gravimetric
SNTHERM
Snow Course
Cold Low - Open
Cold Low - Forest
Cold Mid - Open
Cold Mid - Forest
Cold Up - Open
Cold Up - Forest
SWE (mm)
Julian Day

Figure 5

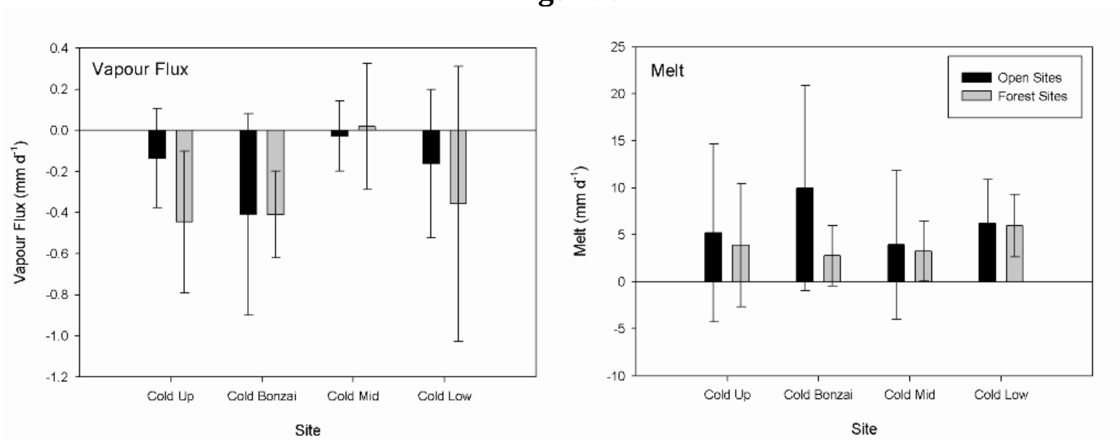

Vapour Flux
Vapour Flux (mm d-1)
0.4
0.2
0.0
-0.2
-0.4
-0.6
-0.8
-1.0
-1.2
Cold Up
Cold Bonzai
Cold Mid
Cold Low
Site
Melt
Melt (mm d-1)
25
20
15
10
5
0
-5
-10
Open Sites
Forest Sites
Cold Up
Cold Bonzai
Cold Mid
Cold Low
Site

Figure 6a

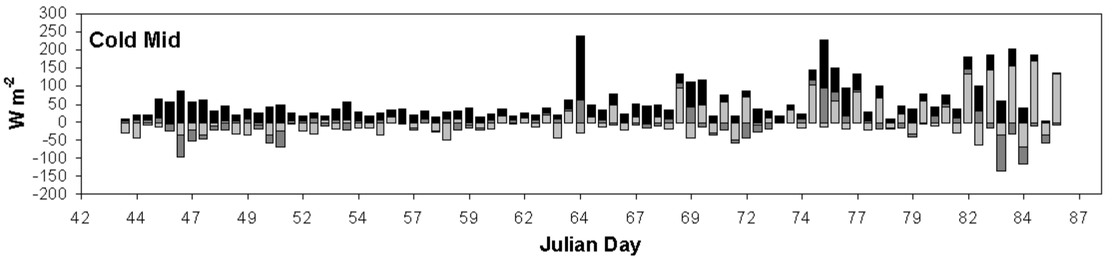


Figure 6b

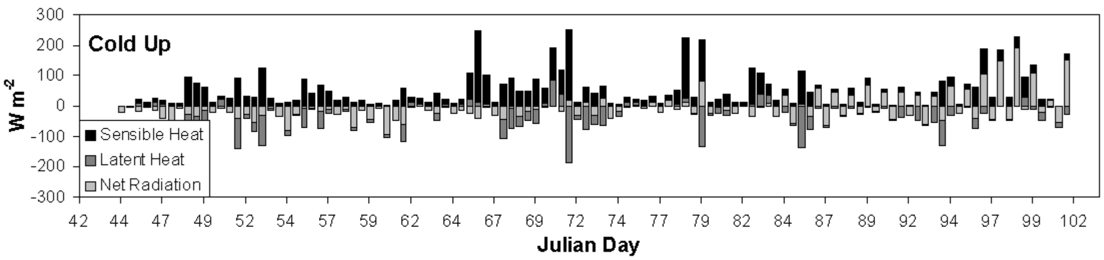


Figure 7

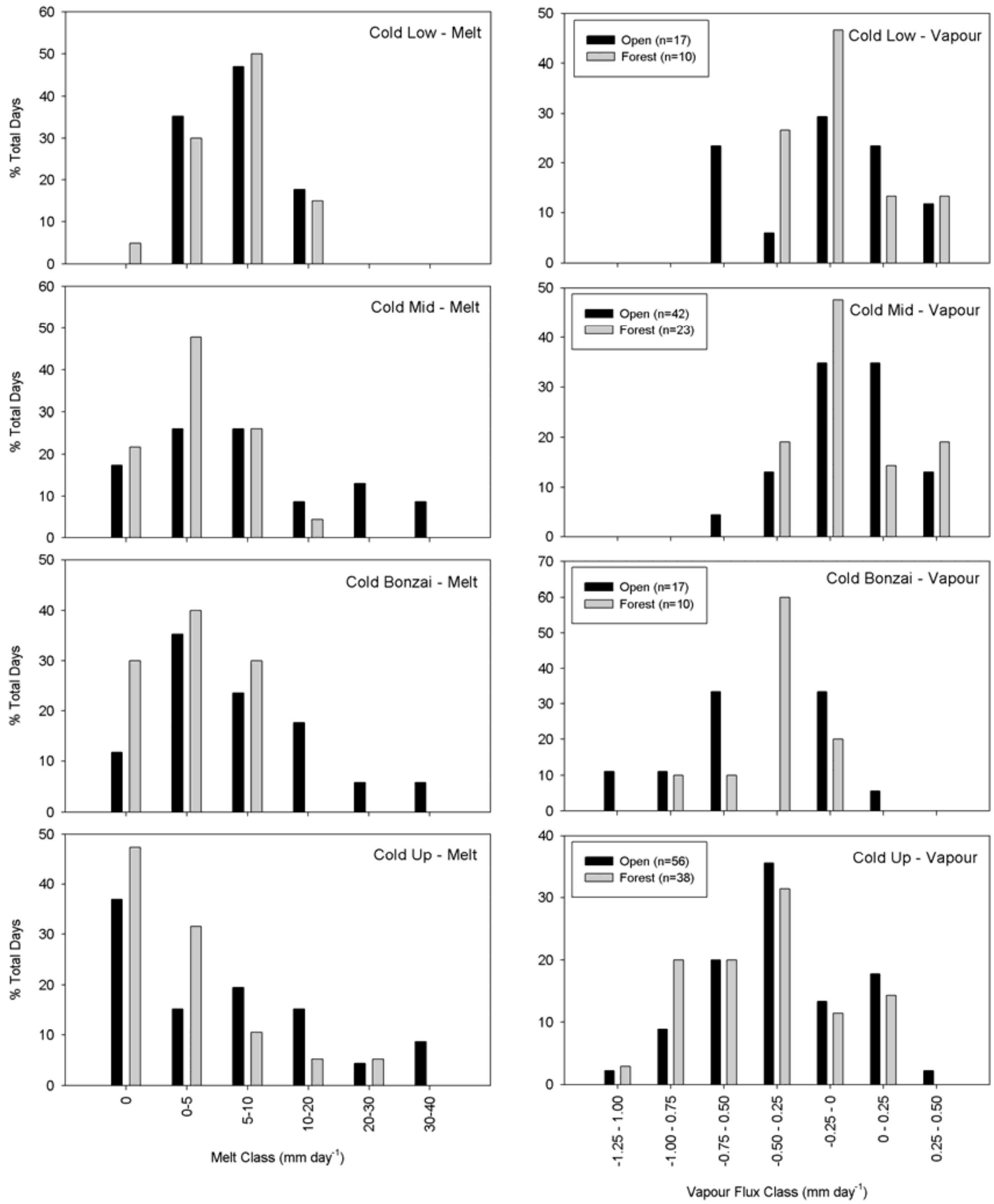

Cold Low - Melt
Cold Low - Vapour
Open (n=17)
Forest (n=10)
Cold Mid - Melt
Cold Mid - Vapour
Open (n=42)
Forest (n=23)
Cold Bonzai - Melt
Cold Bonzai - Vapour
Open (n=17)
Forest (n=10)
Cold Up - Melt
Cold Up - Vapour
Open (n=56)
Forest (n=38)
% Total Days
0
0-5
5-10
10-20
20-30
30-40
Melt Class (mm day^{-1})
-1.25 - 1.00
-1.00 - 0.75
-0.75 - 0.50
-0.50 - 0.25
-0.25 - 0
0 - 0.25
0.25 - 0.50
Vapour Flux Class (mm day^{-1})

Figure 8

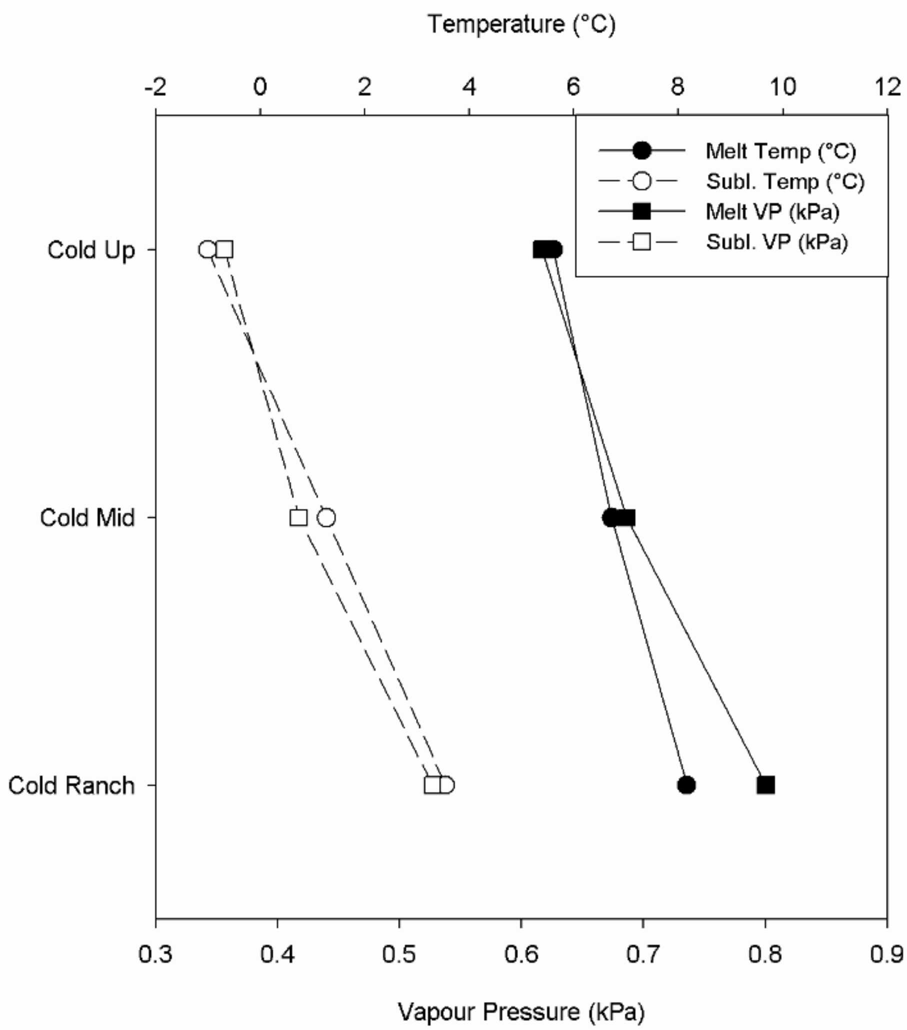
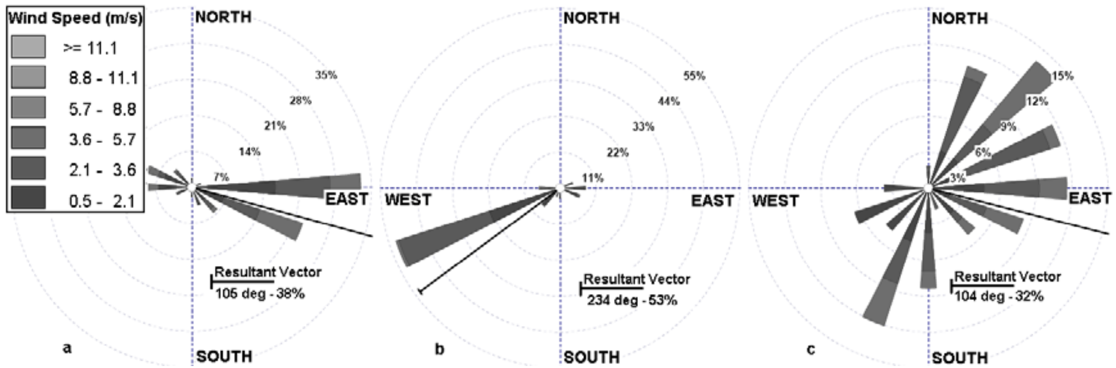
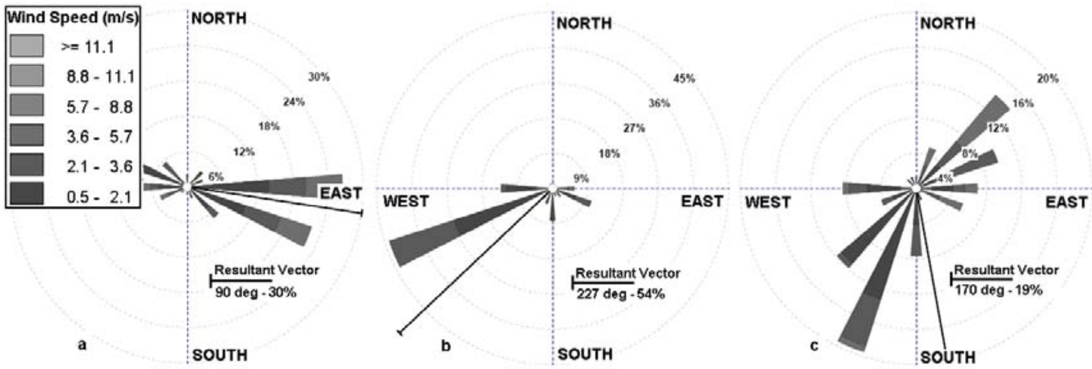

Temperature (°C)
-2
0
2
4
6
8
10
12
Melt Temp (°C)
Subl. Temp (°C)
Melt VP (kPa)
Subl. VP (kPa)
Cold Up
Cold Mid
Cold Ranch
0.3
0.4
0.5
0.6
0.7
0.8
0.9
Vapour Pressure (kPa)

Figure 9



Wind Speed (m/s)
>= 11.1
8.8 - 11.1
5.7 - 8.8
3.6 - 5.7
2.1 - 3.6
0.5 - 2.1
NORTH
35%
28%
21%
14%
7%
EAST
Resultant Vector
105 deg - 38%
a
SOUTH
NORTH
55%
44%
33%
22%
11%
WEST
EAST
Resultant Vector
234 deg - 53%
b
SOUTH
NORTH
15%
12%
9%
6%
3%
WEST
EAST
Resultant Vector
104 deg - 32%
c
SOUTH

Figure 10


Wind Speed (m/s)
>= 11.1
8.8 - 11.1
5.7 - 8.8
3.6 - 5.7
2.1 - 3.6
0.5 - 2.1
NORTH
30%
24%
18%
12%
6%
EAST
Resultant Vector
90 deg - 30%
a
SOUTH
NORTH
45%
36%
27%
18%
9%
WEST
EAST
Resultant Vector
227 deg - 54%
b
SOUTH
NORTH
20%
16%
12%
8%
4%
WEST
EAST
Resultant Vector
170 deg - 19%
c
SOUTH

CHAPTER 4: BOUNDARY-LAYER CONDITIONS AND TELECONNECTIONS ASSOCIATED WITH EXTREME SNOW ABLATION IN A SEMI-ARID BASIN, WESTERN CANADA

Abstract

High-elevation snowpacks provide critical inputs to the hydrological system through the annual snowmelt process. It is of particularly high importance in semi-arid basins with low summer precipitation, such as the Okanagan Basin in western Canada, where most of the summer streamflow is derived from these snowpacks. However, little is currently known about the surface energy balance that controls snow ablation, or the links between these processes and conditions in the atmospheric boundary layer. To better understand these linkages, and the potential effects of a changing climate, surface ablation processes in a high-elevation sub-basin, Coldstream Creek, were statistically linked to the controlling atmospheric boundary layer conditions. These relationships were then used to create a proxy record of historical melt and vapour flux events for the 1972-2007 period coinciding with that from the local radiosonde record. Trends in the variables of interest were explored using time-series analysis. Relationships with the dominant teleconnections that influence climate in western Canada were also examined to provide further insight into the large-scale controls on snowmelt processes. Significant trends were found toward earlier dates of snowmelt initiation and freshet initiation in the Coldstream Basin. Significant positive trends were found in average spring melt, annual deviation of estimated melt values from the long-term mean, and temperature at the 850 mb height. Ablation season climatic and hydrological variables were highly positively correlated with the winter and spring Multivariate El Niño Southern Oscillation index, with the positive phase (i.e., El Niño) associated with higher than average melt rates.

Negative (positive) phases of the winter Pacific North American pattern and spring Pacific Decadal Oscillation were strongly associated with high (low) magnitude sublimation events. This combination of strong correlations and significant trends over time indicates that given future temperature increases, both the magnitude and variability of melt and sublimation events are likely to increase, particularly during favourable phases of the above teleconnections.

Keywords: Energy balance; Snow hydrology; Atmospheric boundary layer; Trends; Teleconnections; Okanagan Basin.

4.1 Introduction

The ongoing and predicted climatic changes resulting from human industrial activity have resulted in significant changes in all natural systems including the hydrological cycle, which appears to be intensifying as a result (Huntington, 2006; Lemke *et al*., 2007). Evidence of a changing climate is often first noticed in high-elevation and high-latitude areas, where even slight changes in temperature can have large impacts on hydrologic systems. In particular, increasing frequency and magnitude of flood and drought events, decreasing water storage in Northern Hemisphere snowpacks, and a shift towards earlier and more intense snowmelt are projected (e.g., Adam *et al*., 2009). One direct result of earlier snowmelt initiation is a shift towards earlier dates of maximum spring runoff. In Canada, annual mean streamflow has decreased significantly in the southern regions, with the strongest declines in August and September, whereas March and April discharge has exhibited significant increases, particularly in British Columbia (Zhang *et al.,* 2001). Two separate studies report that shifts towards earlier peak flows and a spring pulse onset of 10-30 days are common in basins less than 2500 m in elevation (Regonda *et al*., 2005; Stewart *et al*., 2005).

Due to the importance of seasonal snowpacks in the functioning of many terrestrial and climatic systems, much effort has been directed at predicting the effects that anthropogenic climate change may have on snowpack variability (Räisänen, 2008; Vavrus, 2008; Bavay *et al*., 2009). A consistent conclusion regarding 20th century North American snowcover and ablation is that variability has increased, ablation is occurring earlier and in some cases more rapidly, with a concurrent shift towards decreased snow cover duration at lower elevations (Dyer and Mote, 2006). These changes have been directly linked to an increase in spring temperatures, and highlight an ongoing research need for hydro-climatic trend detection at high elevations (McCabe and Clark, 2005; de Jong *et al*., 2009; Stewart 2009). In particular, water resources in semi-arid regions with high-elevation snowpacks are highly susceptible to changes in snowmelt timing and volume as this process maintains stream flows throughout the dry summer months. Given these regions disproportionate reliance on vulnerable high-elevation snowpacks, it is critical that snowmelt processes and links to the meteorological drivers of these processes are better understood.

Research literature and basin meteorological data concerning the upper atmospheric conditions driving snow energy and mass exchanges at higher elevations in the Okanagan are currently non-existent. Also, despite the critical importance of the high elevation snowpack to the Okanagan Basins water balance; very little long term high-elevation data exists with which to evaluate climate-change effects.

A study conducted of four sites along an elevational transect in a high-elevation basin in the north Okanagan during the ablation season of 2007 found that 1-4% of the basins peak snow water equivalent (SWE) was lost to sublimation (Jackson and Prowse, 2009). The largest sublimation events (>0.25 mm d^{-1}) were associated with low atmospheric vapour pressure, temperatures below 0˚C, and higher than average wind speeds, while melt rates were driven almost entirely by sensible heat fluxes and exceeded 30 mm d^{-1} during large-scale advection events. The magnitude and variability of both types of snow ablation increased significantly with elevation. Large melt events were characterized by warm, moist air masses and high easterly winds, while sublimation was driven by colder, drier conditions and higher than average southerly winds. The linkages made between surface and atmospheric processes were cursory, and given that a warming atmosphere is likely to alter the frequency and magnitude of energy being transferred from the atmospheric boundary layer (ABL) to the surface of these critical high-elevation snowpacks, the current linkages between ABL conditions and surface level snowmelt processes require improved understanding. Furthermore, in order to place these linkages and the potential changes in a historical context, recent trends in the relevant hydrologic and climatic variables, as well as correlations to the teleconnection indices that play a dominant role in the climate of western North America need to be elucidated.

Hence, the objectives of this study are as follows: assess how well atmospheric boundary layer variables are correlated with ablation-season snow-surface energy exchanges in a representative sub-basin of a large semi-arid basin in western Canada; determine if these relationships are useful predictors of snowmelt and sublimation events during past ablation seasons; analyze time-series of hydro-meteorological variables for trends in magnitude and frequency; and, assess the relationships with large-scale teleconnection patterns.

4.2 Background

4.2.1 Snow Surface - Atmosphere Linkages in Snowmelt Regimes

The historical and predicted shifts in snow-melt timing are driven by large-scale atmosphere processes operating at the macro scale both temporally and spatially, but they are expressed over time by changes in the daily energy balance of the snowpack at the micro and meso-scale. While there are numerous worldwide examples in the scientific literature of site or basin-specific studies of snow surface energy balance, relatively few focus on the linkages between the ground level and atmospheric boundary layer (ABL; 0-3000 m asl) meteorological variables that act as the link between macro-scale climatic conditions and those at the site-scale.

In general, the energy balance of a melting snowpack is driven by radiation and the turbulent exchanges of sensible and latent heat. An important assumption in site-specific energy balance studies is that snow surface turbulence is locally generated at the micro-scale (i.e., roughness length). However, this assumption is violated in high relief topography and, even in areas of homogenous land cover, the ground-level energy fluxes are driven by air mass characteristics (i.e., temperature, water vapour and wind profiles; Granger and Male, 1978; Helgason and Pomeroy, 2005). Radiation fluxes are highly dependent on micro-scale variability in forest cover and topography (Fierz *et al.*, 2003), and meso-scale climatic conditions. Of the latter, cloud cover is the most important, which is a direct result of the proportion of available moisture in the ABL (Grundstein and Leathers, 1998). Recent studies of linkages between snow energy balance and upper-atmosphere conditions in North America have largely focused on synoptic map pattern analyses (e.g., Yarnal, 1984; Romolo *et al*., 2006a), and defining atmospheric water vapour lapse rates for basin scale snow-melt modelling exercises (Garen and Marks, 2005), but have never been made in a semi-arid, high-elevation hydroclimatic regime such as the Okanagan Basin.

4.2.2 Trends in Snow Melt and Meteorological Drivers

Numerous studies of western North American watersheds with alpine snow have noted earlier dates of maximum spring freshet and reduced mid-elevation snow water equivalent (Stewart *et al*., 2004; Lemke *et al*., 2007). For example, in the Northern Hemisphere, a decline in March and April snow covered area of 7% +/-3.5% (1922 – 2005) is significantly correlated to an increase in spring temperatures. This decrease is especially prominent in the spring and especially since the late 1970s (Lemke *et al*., 2007). Maximum high-elevation snow water equivalent (SWE), dates of spring freshet onset and center of hydrograph mass in western North America have shifted earlier in the year by one to four weeks since 1950, and are linked to positive 700 mb geopotential height anomalies (Stewart *et al*., 2005). These trends are exacerbated by greater declines at lower elevations where spring temperatures are closer to the 0˚C threshold (Stewart, 2009.)

Radiosonde data from the Northern Hemisphere have recently been analysed for trends in ABL variables, which have a direct influence on snow energy balance. For instance, Ross and Elliot (2001) reported significant positive trends in the surface to 500 mb precipitable water, 850 mb specific humidity, dewpoint and temperature from 1973-1995 over the Northern Hemisphere. Water vapour increases were found to be larger, more uniform and more significant over North America than Eurasia, and are attributable to the late 1970s climate shift to a warm phase of the Pacific Decadal Oscillation (PDO) that primarily affected North America. Trenberth *et al*. (2005) extended this analysis by using the NCEP and ERA-40 re-analysis data-sets. They report that the variability in atmospheric water vapour from 1988-2001 was dominated by the El Niño Southern Oscillation (ENSO), particularly by the high-magnitude 1997-98 event.

4.2.3 Ablation Season Teleconnection Linkages

Teleconnections are a global phenomenon, influential over varying spatial and time scales. They have been linked to snowpack variability in Europe (Bednorz, 2004; Scherrer and Appenzeller, 2006), Siberia (Iijima *et al*., 2007), and Eurasia in its entirety (Clark *et al*., 1999; Bamzai and Shukla, 1999).The three main large-scale teleconnection patterns that have been identified as exerting a strong influence on annual snowpack fluctuations in western North America are the ENSO (Rasmussen and Carpenter, 1982), the PDO (Trenberth and Hurrell, 1994), and the Pacific North American Pattern (PNA) (Wallace and Gutzler, 1981). Time series of the derived hydroclimatic variables were correlated with the Multivariate ENSO Index (MEI), the PDO (Mantua *et al*., 1997) and PNA indices to highlight possible teleconnection influences. The MEI is considered to be a more robust index of the ENSO than the traditional Southern Oscillation Index (SOI) , and is based on sea-level pressure, zonal and meridional surface winds, sea surface temperature, surface air temperature, and cloud cover fraction (Wolter and Timlin, 1993; Wolter, 1998).

In south-eastern British Columbia, positive (negative) phases of the PNA are associated with high (low) spring temperatures. A positive PNA exerts the strongest influence on snow cover variability in this region, and is associated with reduced snow cover in western Canada. The ENSO produces the same effect but with a much weaker correlation (Brown and Goodison, 1996; Moore and McKendry, 1996). A particularly strong signal has also been reported to emerge as elevation increases (Hsieh and Tang, 2001). Negative ENSO phases (La Niña) are significantly related to late melt initiation dates, indicating that this phase of the ENSO pattern not only results in deeper snowpacks in western North America, but greater duration due to cooler than average spring temperatures (Romolo *et al*., 2006b). In general, it can be stated that temperature is higher (lower) in British Columbia and precipitation lower (higher) during El Niño (La Niña) years (Stahl *et al.,* 2006). These teleconnections are also linked, with magnified effects when two or more are in phase, and reduced climatic signals when they are out of phase (Barton and Ramirez, 2004; Stahl *et al*., 2006).

4.3 Study Site Description

Covering an area of 8046 km^2, the Okanagan Basin extends approximately 185 km from north to south and comprises a portion of the northern Columbia River Basin (Figure 1). The region is characterized by a dry continental climate, with precipitation averaging 250-300 mm in the valley and >1000 mm at the higher elevations, 85% of which has been estimated to be lost to evapotranspiration (Cohen and Kulkarni, 2001). Summer rainfall is largely driven by local-scale convection, with winter precipitation influenced primarily by synoptic weather systems originating in the Pacific Ocean. Average annual temperature decreases and precipitation increases along a south to north transect within the basin, with the southern portion containing Canada's only true desert ecosystem (Cohen and Kulkarni, 2001). Distribution of snow water equivalent (SWE) in the Basin follows the same spatial pattern.

In British Columbia, and the Okanagan Basin in particular, the melting high-elevation snowpack provides critical water inputs to lower elevations during the summer months when precipitation is lowest and agricultural and municipal demand is highest. The valley bottoms are semi-arid, with sparse forest cover in the south. Precipitation and forest cover increase with elevation in the basin, and the wetter high-elevation areas provide the majority of the critical freshwater recharge to the semi-arid low-lying areas. The Okanagan is currently subject to increasing stress on its water resources, as irrigation, intensive agriculture and a rapidly expanding population surpass previous predictions made by water resource authorities (Cohen and Neale, 2006). In addition, extensive forest cover changes resulting from the mountain pine-beetle epidemic will have a pronounced effect on the hydrological response of the previously forested areas. In particular, snow accumulation and ablation patterns will change as openings in the forest become larger and more interconnected, increasing snow accumulation, melt rates and fetch lengths (Carroll *et al.*, 2006; Boon, 2009).

The Okanagan is already undergoing substantial hydro-climatic change. Over the past three decades, spring and fall temperatures have increased significantly while annual precipitation has decreased. Streamflow changes show a significant difference between the western and eastern portions of the basin, with winter flows increasing in the former due to increases in early winter rains (Whitfield and Cannon, 2000). These changes in the basins hydrology coincide with a shift towards a 2 days/decade (1901 – 2007) earlier occurrence of the spring 0˚C isotherm at the Vernon Coldstream Ranch station in the northern portion of the Okanagan Basin (Bonsal and Prowse, 2003; Figure 2). Climate projections generated by Global Circulation Models suggest that as winter temperature and precipitation increase over the next century less precipitation will fall as snow and the snowmelt season will occur 4-6 weeks earlier, resulting in “considerable reductions” in annual and spring flow volumes (Merritt *et al.,* 2006).

The sub-basin selected for study (Coldstream Basin), was considered to have the best combined hydrometric and meteorological records available within the Okanagan Basin. Of particular importance was the high-elevation access and additional meteorological data available from the Silver Star Ski Resort situated above the upper elevations of the catchment, and the high-resolution radiosonde record from the Kelowna and Vernon Airports (Table 1).

Located approximately 20 km east of Vernon, Coldstream Basin covers 58.50 km^2 and intersects with the larger Coldstream Valley drainage, which is 205 km^2 in total area and flows into Kalamalka Lake (Water Survey of Canada, 2009; Figure 1). Climate normals for the period 1971-2000 from the Vernon Coldstream Ranch station show the mean temperature of the coldest (January) and warmest (July) months to be -5.0˚C and 19.1˚C respectively, and the average total annual precipitation is 357 mm, with 128 mm of that falling as snow (Environment Canada, 2009). At higher elevations, the precipitation regime is dominated by winter snowfall, with the snow course at Silver Star Resort recording the highest mean April 1 SWE value in the Okanagan Basin. Over the last four decades, the Coldstream Creek basin has experienced an earlier onset of maximum spring discharge (5 days/decade) and, lower fall and winter flows (Whitfield, 2001).

The four sites (each consisting of open and forest covered sub-plots) studied in Jackson and Prowse (2009) were distributed along an elevational transect in the Coldstream Basin, at 650 m (*Cold Low*), 980 m (*Cold Mid*), 1143 m (*Cold Bonzai*), and 1450 m (*Cold Up*).

4.4 Methodology

To assess how well ABL variables are correlated with ablation season snow surface energy exchanges, data from a single high-elevation site in the Coldstream Basin (Jackson and Prowse, 2009) were correlated with radiosonde data for the 2007 ablation season, and regression equations developed to explain variances in the micro-scale energy balance using the radiosonde data. To determine the utility of these equations as predictors of historical snow ablation events, a proxy record of melt and sublimation events was derived from the complete regional radiosonde record (1972-2007). Potential changes in the historical magnitude and frequency of these events were explored by analyzing this proxy record, along with the meteorological variables that influence melt and sublimation and other hydroclimatic variables for the presence of significant trends. Finally, these variables were correlated with average winter and spring teleconnection indices to determine whether relationships existed with the PDO, MEI and PNA (described below).

4.4.1 Atmospheric Boundary Layer Data

The radiosonde data used in this study are a combination of two records from sites approximately 50 km apart; the Vernon Airport (1972-1994) and Kelowna Airport (1994-2007; Table 1; Figure 1). While the original purpose of radiosonde data was operational, comprehensive analyses of the data quality have also been conducted to determine their utility for climate change research (Elliot and Gaffen, 1991; Garand *et al*., 1992). Data for this study were extracted from the Integrated Global Radiosonde Archive (Durre *et al*., 2006; Durre and Yin, 2007). Previous work suggested that most of an upward trend in water vapour in the Northern Hemisphere occurred after 1973; therefore the analysis presented here is constrained to the local radiosonde record period, and not extended using the NCAR Reanalysis dataset as in Trenberth *et al*. (2005). Since the focus of this study lies in the lower troposphere (<700 mb height, or ~3 km asl), the temperature corrections recommended by Luers and Eskridge (1998) were unnecessary.

Climate records that span long periods are subject to artificial "change points" in the time-series resulting from station moves and changes to instrumentation and the surrounding physical environment. Potential artificial change points in the radiosonde time-series were identified from the station histories following Gaffen *et al*. (2000) to ensure that physical change points resulting from step changes in upper atmosphere conditions were retained. An annual correlation of temperature and relative humidity from Coldstream Ranch was used to determine if the 1994 relocation of the radiosonde station significantly changed the relationship between the upper-air and ground-level climate variables (Gaffen *et al*., 2000). A correlation of 0.95 +/- 0.01 between the two radiosonde data sets and the Coldstream Ranch data indicates that the data were representative of ABL conditions over the Coldstream Basin regardless of its release location.

Time-series analysis can be adversely affected by periods of missing data. Over the entire record, only 2.9% of the radiosonde data were missing, and of the 8 days with more than 24 hours of missing data, only 2 fell within an ablation season. To provide a complete time-series for the historical change analysis, missing data in the radiosonde record were estimated using average lapse rates between the Coldstream Ranch data and the radiosonde data. Thus, the likelihood of estimated data affecting the conclusions of this study is extremely low.

4.4.2 Snow Surface - Atmosphere Linkages

To clearly define the atmospheric conditions resulting in the surface ablation processes, melt and vapour fluxes from the 2007 study period were grouped along with the corresponding air temperature, net radiation, vapour pressure, wind speed and estimated turbulent fluxes of latent and sensible heat. A Mann-Whitney U test determined whether the melt and vapour flux classification had identified groups of meteorological variables that differed significantly in magnitude. These groups were compared to the ABL conditions to assess which meso-scale variables most influenced the temporal variation in melt and sublimation at *Cold Up*, the highest study-site in the basin (Jackson and Prowse, 2009). This site was chosen for several reasons: it is the most open and highest of the four sites within the basin and therefore most likely to be representative of other high-elevation areas in the Okanagan Basin; the snow lysimeter, snow course and modelled estimates were in closest correspondence here; the site (1450 m) lies in the mid-range of the 850mb geopotential height for the study period (1283-1584 m); and the upper atmosphere conditions were most highly correlated with the meteorological variables recorded here (0.93, temperature; 0.82, vapour pressure and 0.43, wind speed).

Geopotential height, temperature, vapour pressure, potential temperature, zonal and meridional wind fields and wind speed for the standard near-surface measurement levels were extracted from the IGRA archive. All values for the 850 mb and 700 mb heights were grouped into diurnal time series, with the UTC 1200 and 2400 soundings corresponding to 0400 and 1600 PST. These soundings were taken to represent the average conditions for the preceding 12 hour period, and all ground level meteorological variables were averaged over the same interval to represent diurnal variations at *Cold Up*. The 700 mb wind values were used under the assumption that the height of the surroundings peaks would introduce localized topographical effects in the 850 mb wind field. This conclusion is supported by wind roses plotted for the three instrumented sites, which show a high degree of variation in speed and direction, indicating that topographic forcing substantially alters the wind field within the basin (Jackson and Prowse, 2009).

To assess which ABL variables were most highly correlated with the vapour flux and melt totals at *Cold Up* site, a multiple stepwise regression with forward selection was used. All variables were retained at the 95% significance level, and autocorrelation was tested using the Durbin-Watson statistic. As the correlation between climatic variables is a concern when using multiple regression techniques, vapour pressure was used in place of relative humidity in an attempt to remove some of the dependence of atmospheric water vapour concentrations on temperature.

4.4.3 Definition of Ablation Season

In order to identify trends in the historical frequency of high magnitude melt and sublimation events in the Coldstream Basin, the length of each ablation season at *Cold Up* had to be estimated. This was done by determining the Julian day (JD) dates of the initiation of ablation and the onset of patchy snow cover (which marks the end of measurable snow surface energy exchanges unaffected by local advection from snow-free areas) at the site in the 2007 field season (Jackson and Prowse, 2009). Because data taken from a single year study are subject to the unique snowpack and meteorological conditions of that year, the study of 2007 was placed in a historical context. At the nearby Silver Star Resort, the snow course recorded an April 1st 2007 SWE that was 98% of the historical average (1959-2007). Therefore the study year can be considered representative of average historical conditions.

The snow surface energy exchanges shifted towards an ablative regime at *Cold Up* when the -2˚C isotherm was crossed, corresponding on average to the 2˚C isotherm at Coldstream Ranch (Jackson and Prowse, 2009). A running 7-day mean of temperature at Coldstream Ranch defined the day that this isotherm was last crossed, and therefore the 'beginning' of the ablation season (Regonda *et al*., 2005).

Inter-annual variability of the snowpack was accounted for using SWE data from a nearby automatic snow pillow within the Okanagan Basin (Mission Creek) to benchmark the end of each ablation season in over the study period (Table 1). In 2007, the ablation season snow surface energy exchanges became dominated by local scale advection from snow free areas on JD 101, marking the 'end' of the measurable (by conventional methods) snowpack ablation. This corresponded to a SWE value at Mission Creek that was 93% of the 2007 maximum – the delay in loss of snowcover here is attributable to its higher elevation. Therefore, the day of each year that showed a SWE value 93% of maximum (on the ablation side of the curve) at Mission Creek was assumed to correspond to the 'end' of ablation conditions at *Cold Up*. This 'window' was then used to define the snowmelt season for each year at *Cold Up*, and the results were checked against the snow course data from Silver Star. Since this method of defining the ablation season is based partially on air temperature, the March-April-May (MAM) period was also selected for trend analysis to ensure that any trends or correlations resulting from the use of a temperature defined ablation season did not affect the identified relationships.

4.4.4 Trends in Snowmelt and Meteorological Drivers

The study period (1972-2007) is marked by pronounced large-scale climatic changes, including rising temperatures globally, the well documented phase shift of the PDO in 1976/77, and increased frequency and magnitude of the ENSO (Lemke *et al*., 2007). To determine if similar changes occurred in the study basin, the defined ablation season initiation and completion dates, proxy measures of ablation season melt and sublimation events, April 1st SWE values, and streamflow timing and volume were tested for significant trends. The April 1 SWE values are from the nearby Silver Star snow course and span the period 1959-2007. The onset of the spring freshet was determined from the daily discharge record (1968-2007) at the Coldstream Creek gauge using the method outlined in Cayan *et al.* (2001).

An annual time series of days with higher than average melt and sublimation rates was estimated using the regression equations described above. The study period averages were calculated for the estimated sublimation and melt values, and all days that experienced events of greater magnitude than 1 standard deviation above the mean were tallied for each year. To test for the possibility that snowmelt events are becoming more intense and of longer duration under a changing climate, high magnitude melt events of two and two or greater days in length were also documented. A list of all variables tested for trends is in Table 2.

Trends were tested for using three methods, a simple linear regression, and two versions of the Mann-Kendall test – a non-parametric test for trend that is commonly used in climate and hydrologic time series analysis (Mann, 1945; Kendall, 1975). The first used the MAKESENS program, which employs the Sen's nonparametric method to test for linear trends (Salmi *et al*., 2002). The potential for the result being affected by serial-correlation in the data was addressed by employing a statistical package (Bronaugh and Werner, 2008), which first pre-whitens the data to remove serial autocorrelation, then applies the Mann-Kendall test with Sen's slope analysis. The pre-whitening method presented by Zhang *et al*. (2001) was used as it performs better than the method proposed by Yue *et al*. (2002) when the time-series being tested has low serial correlation.

4.4.5 Ablation Season Teleconnection Linkages

This analysis was conducted on a monthly basis, as well as by the standard winter – December-January-February (DJF); and spring periods – March-April-May (MAM). The winter period was included to allow potential climatic lag effects to be taken into account, and the latter best represents the ablation season in the Okanagan Basin. The climatic shift attributed by other authors (e.g., Stahl *et al*., 2006; Sobolowski and Frei, 2007) to the PDO phase change in 1976-77 occurred just five years after the beginning of the radiosonde record. Therefore, the period prior to the phase shift is insufficient to conclusively determine whether this resulted in a statistically significant change-point in the time series.

Due to the well documented influence of ENSO (Woo and Thorne, 2003; Regonda *et al*., 2005), PDO (Hamlet *et al*, 2005; Stewart *et al*., 2005) and PNA on hydroclimatic trends, the effects of these teleconnections were taken into account by employing a two stage trend detection process. First, the ablation season variables showing significant trends following the first Mann-Kendall analysis were regressed against all significantly correlated teleconnection indices using multiple stepwise regression. Second, the residuals of the regression equations that explained a significant proportion of the annual variation with the ENSO, PDO and PNA were tested again for trends. This allowed the trends that are likely due to changes in the phase of the correlated teleconnections to be statistically separated from those that are a result of other processes (i.e., anthropogenic climate change). However, it should be noted that this attempt at separating the influence of teleconnections from those of anthropogenic climate change, while statistically sound, is not necessarily reflective of the real world relationships. It still uncertain whether these teleconnections are impacted by anthropogenic climate change, and to what degree (Corti *et al*., 1999).

4.5 Results and Discussion

4.5.1 Snow Surface - Atmosphere Linkages

Strong and significant relationships were found between ablation-season snow-surface energy exchanges and the driving meteorological conditions in the ABL over the Coldstream Basin in 2007. The ABL conditions that drive melt and sublimation events differ significantly between these two ablation processes. For both melt and vapour fluxes, the average ABL conditions are significantly different between magnitude classes, with the exception of small sublimation events (>-0.25 mm d^{-1}) (Figures 3-5). The difference between small and large condensation and small sublimation events results from increased meridional winds at the 700 mb geopotential height (~ 3000 m), which lies above the highest peaks in the Okanagan, and can thus be considered representative of free stream wind speed and direction in the ABL. The significant differences between classes show that the grouping of sublimation and melt is a reliable method for identifying meteorological conditions associated with low frequency, high magnitude events.

Seventy percent of the variance in sensible heat fluxes is predicted by temperature at the 850 mb height and zonal winds at the 700 mb height. Sensible heat is the primary driver of melt in all classes, with net radiation inputs playing a substantially smaller role. However, the ABL conditions describe only 47% of the variance in melt without ground-level net radiation included as a predictor, and 57% when this variable is included, reinforcing the importance of the site specific radiation balance to snowmelt in this basin (Jackson and Prowse, 2009). Figure 3 shows the average air temperature, wind speed, radiation and turbulent fluxes for the days in each melt magnitude class. In general, magnitude and variation are similar for temperatures measured at both ground level and the 850 mb height. Wind speeds are significantly higher and have greater variation at the 850 mb height than at *Cold Up*. The high fluxes of latent heat for both the 'no melt' and high melt (30-40 mm d^{-1}) classes are associated with high sublimation rates for the former and high condensation rates for the latter (Jackson and Prowse, 2009; Figure 3).

Fifty-four percent of the variance in diurnal sublimation is explained by vapour pressure and temperature at the 850 mb height. There is slight positive autocorrelation in the residuals at subsequent time steps, which is common in the ABL, as the current state will retain some 'memory' of the preceding conditions. The differences in meteorological variations at *Cold Up* and in the ABL between vapour flux magnitude classes are shown in Figure 4. Temperatures in both data sets are below 0˚C during most sublimation events and average 2-3˚C during condensation events. Vapour pressure is negatively correlated to the sublimation rate, increasing to 0.8 kPa during the highest magnitude condensation events (0.25-0.50 mm d^{-1}), and is higher at the ground level than at the 850 mb height, a result of vapour exchange between the snow surface and the boundary layer. Wind speed is variable among classes, with 850 mb height winds higher than those at *Cold Up*, with the exception of the highest magnitude sublimation event class. No significant relationship exists between wind speed at *Cold Up* and sublimation, although wind speeds at the 700 mb height do appear to be higher for the vapour flux transfers at the extreme ends of the range encountered during the study period.

The plotted energy flux classes clearly show latent heat transfers with greater magnitude than either net radiation or sensible heat fluxes for the highest magnitude sublimation class, with the latter two components of the snow energy balance showing much greater variation about the negative values (Figure 4). Condensation events are characterized by high sensible heat transfers and negligible radiative fluxes, and in the case of the larger events, sensible heat fluxes average 95 W m^{-2}.

The obvious deviation from the above relationships is the 0.5-0.75 mm d^{-1} sublimation event class. Vapour pressure, temperature and sensible heat fluxes in this class are more consistent with condensation events than sublimation events. Wind speeds at *Cold Up* don't differ significantly from the other event classes, unlike the winds at the 850 mb height. This indicates that this event class is more likely due to evaporation of free water in the snowpack during periods of large scale warm air advection, although only one of the three days in this class corresponds to a high magnitude melt event.

Profiles for the highest magnitude melt and sublimation events from the 2007 ablation season are shown in Figure 5. Vapour pressure and temperature at the 850 mb height average 0.27 kPa and -4˚C during large sublimation events, and 0.63 kPa and 6˚C during high magnitude melt events, with both variables showing significant differences between classes ($\alpha = 0.05$). The difference between event types is further explained by wind roses of the 700 mb height wind speed and direction over the study period (Figure 6). Large melt events are characterised by south-westerly winds, while sublimation events typically occur under westerly winds. This differs from the ground level values measured during the 2007 ablation season, when melt events at *Cold Up* were dominated by easterly winds, and sublimation events by southerly winds (Jackson and Prowse, 2009). It is uncertain how much of this difference is due to the topographical modification of surface winds. These differences highlight the difficulties encountered when comparing records of greatly differing lengths, and underscore the need for long-term datasets in sensitive high-elevation areas.

4.5.2 Trends in Snow Melt and Meteorological Drivers

The results of the trend analysis are summarized in Tables 3-5 and Figure 7. No significant trend was noted in the number of days with high magnitude sublimation or melt events therefore, given the available record, the frequency of ABL conditions conducive to these high magnitude events is not increasing (Table 4, 5). The latter is not surprising, given that temperature and water vapour in the ABL are increasing, but with a weak but positive trend reported for Canada in an analysis of Northern Hemisphere radiosonde records (Ross and Elliot, 2001). Note that while the regression equations that this proxy record and subsequent time-series analyses are based on are statistically sound, they are based on a single seasons data. This further reinforces the need for better long term high-elevation climate data sets.

Significant trends were found towards earlier dates of freshet and snowmelt initiation; comparable to results reported by similar studies (Zhang *et al*., 2001; Stewart *et al*., 2005; Table 3). The date of freshet initiation in the Coldstream Basin is significantly negatively correlated with positive phases of the DJF MEI, and MAM PDO, MEI and PNA. The initial trend was not retained after testing the residuals from the multiple regression analysis, raising the possibility that this trend results from the positive phase of the PDO that spans the majority of the period analysed.

Correlations of snowmelt initiation and freshet onset with all teleconnection indices are significant, uniformly negative, and relatively high (-0.4 to -0.6). The dates of freshet initiation and snowmelt initiation were correlated with both the DJF and MAM indices (Table 3). Positive phases of the PDO, MEI and PNA are characterised by warmer than average spring temperatures, that result in a significantly earlier onset of snowmelt and the freshet. Conversely, negative phases of these teleconnections are associated with significantly later onset of snowmelt and the freshet. Similar associations have been reported in the Peace River Basin (Romolo *et al*. (2006a; 2006b).

While maximum discharge shows no significant trend in the Coldstream Basin, other authors note that relationships exist between hydrologic indicators and teleconnection indices in central British Columbia (Woo and Thorne, 2003). No significant trend was found for April 1st SWE values at Silver Star, which is inconsistent with trends reported for the western United States (Hamlet *et al*., 2005; Regonda *et al*., 2005). However, significant negative correlations exist for April 1st SWE and the DJF PDO and PNA, indicating that warm phases of both teleconnections are linked to reduced spring snowpacks in the northern Okanagan Basin, and vice-versa. This snow course is situated at an elevation of sufficient height that it is currently buffered from the effects of rising spring temperatures at lower elevations in the Okanagan Basin, an important factor in other basins in western North America (Stewart *et al*., 2005).

The variables defined by the MAM spring period that show significant trends over the 1972-2007 period are the estimated average MAM melt, average annual melt anomalies, and the average MAM 850 mb potential temperature (Table 4). Significant trends were retained for the average MAM melt and 850 mb potential temperature after accounting for the influence of the teleconnection indices, but not for the annual melt anomalies, indicating that the magnitude of annual deviation from the long term mean for the latter is driven, at least in part, by the DJF and MAM MEI (Table 6). No significant trends were found for any of the time series of melt or sublimation variables defined by the ablation window method.

4.5.3 Ablation Season Teleconnection Linkages

In general, snowmelt and sublimation variables derived from the radiosonde record for the Coldstream Basin were strongly correlated to the winter and spring teleconnections tested in this study. All melt and sublimation variables tested show significant correlations with the DJF and MAM PDO indices (0.34 – 0.50; Tables 4 and

5). In particular, the number and percentage of days with high magnitude sublimation events, as well as the annual average and maximum sublimation in the ablation window season are negatively correlated with the positive DJF and MAM PDO indices, and the DJF PNA. This indicates that negative phases of the PDO and PNA are likely to result in increased ablation season sublimation rates, and vice-versa. The same relationships apply for ABL vapour pressure values, although the correlations with teleconnection indices are lower, and the DJF PNA appears to have a significant influence on vapour pressure at the 850 mb height during the snowmelt season.

Of all the melt variables tested, only the percent of ablation season with high magnitude melt shows a significant positive correlation with the MAM PDO (Table 5). The presence of trends in the average MAM melt and 850 mb potential temperature is particularly interesting, as these variables possess high correlations with the MAM MEI and PNA (Table 6). Both are essentially measures of air temperature, meaning that spring temperatures in the northern Okanagan Basin have been increasing since 1972, statistically independent of the influence of any of the three teleconnections included in this analysis.

In general, the snowmelt related variables are highly correlated with the DJF and MAM MEI indices, with the MAM PDO and PNA also influencing the average and maximum melt rates in a given year (0.33 to 0.61). The time series of sublimation variables show consistent correlations with the DJF PNA and MAM PDO, although the sign and magnitude varies. The average MAM sublimation, average sublimation anomaly and maximum sublimation are positively correlated with these indices, indicating that a negative phase of the PNA is associated with higher magnitude sublimation events, and vice-versa. Additionally, the frequency of high magnitude sublimation events increases with a negative phase of either the DJF PNA or the MAM PDO, and the variation in the daily sublimation rates decreases during positive phases of the PNA. One explanation for these relationships is that the lower vapour pressure in the ABL during negative phases of the PNA increases the vapour pressure gradient above the snow surface, increasing sublimation losses at the snow surface (Grundstein and Leathers, 1998).

The constructive and destructive phasing between these teleconnections and the resulting changes to snow energy and mass balance have been explored and the conclusions are consistent across studies in western North America. An El Niño (La Niña) occurring during a positive (negative) phase of the PDO is associated with increased (decreased) runoff and temperature in the Columbia River basin (Barton and Ramirez, 2004). In a similar study, positive PDO and PNA phases occurring in tandem were associated with warm, dry winter anomalies in mainland BC, with negative phases possessing the opposite association (Stahl *et al.,* 2006).

All potential temperature and vapour pressure variables for the MAM at the 850 mb height show positive correlations with the DJF and MAM PDO, MEI and PNA indices (0.33 to 0.6). Generally, positive phases of these three teleconnection indices are associated with higher temperatures and vapour pressure in the ABL in the northern Okanagan Basin.

Evidence exists for an increase in tropospheric water vapour at higher latitudes that spans various phases of the relevant teleconnection indices (Huntington, 2006; Ross and Elliot, 2001). However, the radiosonde record in the Okanagan does not show a similar trend. In theory, higher vapour pressures would have the effect of reducing sublimation losses, but would enhance snowmelt occurring during large scale advection of warm, moist air masses via the transfer of latent heat to the snow surface. The increase in atmospheric and surface water vapour noted by other authors is likely due to increasing evapotranspiration as a result of a warming climate, resulting in the paradox discussed by Brutsaert and Parlange (1998). However, due to the Okanagan Basins location in the rain shadow of the Coast Mountains, its climate is much drier than the rest of British Columbia and is classified as a semi-arid. Therefore, it is not surprising that an increased trend in ABL vapour pressure is absent, as there is much less available surface water for evapotranspiration.

If current trends continue in the Okanagan Basin, the onset of the snowmelt period and the spring freshet will continue to shift earlier in the year. The increasing ABL temperatures are likely to increase the variability and magnitude of melt events in the north Okanagan Basin during the spring period, particularly during positive phases of the ENSO and PDO. The probability of sublimation increasing in magnitude is more difficult to predict as there are no significant trends. However, the strong correlations with the DJF PNA and MAM PDO suggest that during negative phases of these two teleconnections, sublimation events are likely to increase in both magnitude and frequency relative to the long term mean.

4.6 Conclusions

Conditions in the ABL were significantly linked to snow surface energy and mass balance measured at a high-elevation basin in the northern Okanagan during the 2007 snowmelt season. Temperature and vapour pressure at the 850 mb geopotential height are significant predictors of the measured variation in near surface melt and sublimation rates. Sensible heat was found to be the primary driver of snowmelt events, with the relative influence of radiation inputs increasing in importance as the ablation season progressed. Conditions in the ABL differ significantly between the two types of ablation event, and between event magnitude classes. No significant difference was found in wind speeds between magnitude classes of melt and sublimation rates.

Significant trends were noted towards earlier dates of snowmelt and freshet initiation. Average melt for the March-April-May period, annual melt anomalies, and the 850 mb average potential temperature also exhibit positive trends over the 1972-2007 period.

This work has quantified the linkages between micro-scale snow surface processes and the associated conditions in the ABL, and clarified the connections to the dominant teleconnection patterns influencing climate in British Columbia. Significant positive correlations were found between snowmelt event magnitude and frequency and the DJF and MAM MEI, MAM PDO and PNA indices. Sublimation event magnitude and frequency is correlated with the DJF PNA and MAM PDO, with negative phases associated with high magnitudes and frequencies of sublimation. However, the meso-scale synoptic conditions that link the conditions in the ABL associated with high magnitude melt and sublimation events and the macro-scale teleconnections are still undefined. Future work should explore the links between the synoptic conditions associated with high magnitude events and the phase and strength of the various teleconnections that drive snowmelt and sublimation in the Okanagan Basin.

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Figure 6: 700 mb height wind roses for the historical record of all high magnitude (a) melt and (b) sublimation events derived from the radiosonde record (1972-2007). The resultant vector indicates the average wind speed direction for the record period.

Figure 7: Julian dates of (a) snowmelt initiation and (b) onset of spring freshet in Coldstream Creek. Trends significant at $\alpha=0.01$. Also shown are average (c) melt magnitude (mm d^{-1}), (d) melt anomalies (mm d^{-1}) and (e) 850 mb height temperature (˚C), from the radiosonde record (1972-2007). Trends significant at $\alpha=0.05$.

Table 1

Station	Site ID	Type	Record Period	Elevation (m)	Variables
Coldstream Creek	08NM142	Hydrometric	1967 - 2007	620	Discharge
Vernon CS Ranch	1128581	Climate	1991 - 2007	482	WBT, DBT, RH, WS, WD, P
Kelowna A	1123970	Climate	1971-2008	430	WBT, DBT, RH, WS, WD, P
Vernon	1128551	Climate	1971-1995	556	WBT, DBT, RH, WS, WD, P
Coldstream Up	N/A	Climate	2007	1456	T, RH, WS, WD,NR, STP, SD, SWE
Kelowna Airport	71203	Radiosonde	1994 - 2007	456	T, RH, VP, GPH, WS, WD, P
Vernon Airport	71115	Radiosonde	1972 - 1994	556	T, RH, VP, GPH, WS, WD, P
Mission Creek	2F05P	Snow Pillow	1969-2007	1780	T, SD, SWE, PC
Silver Star Mtn.	2F10	Snow Course	1959 - 2007	1840	SWE, SD

Table 2

Defined by ablation window and MAM period	**SWE and Discharge**
Values derived from regression estimates (melt and sublimation)	date of freshet start
annual average (estimated and anomaly)	annual max. discharge
annual standard deviation (estimated and anomaly)	date of annual max. discharge
annual maximum (estimated and anomaly)	total annual discharge
annual minimum (estimated and anomaly)	average annual discharge
# high magnitude events (> 1 std. dev. above historical average	annual std. dev. of discharge
% ablation season with high magnitude events	April 1 snow water equivalent
# melt events > 2 days in length	
# melt events => 2 days in length	
length of longest melt event	
Values from radiosonde record	**Ablation window variables**
850mb maximum potential temperature	date of snowmelt initiation
850mb minimum potential temperature	date of snowmelt completion
850mb average potential temperature	snowmelt season length
850mb maximum vapour pressure	
850mb minimum vapour pressure	
850mb average vapour pressure	

Table 3

	Trend Analysis			Teleconnection Correlations					
				DJF Indices			MAM Indices		
SWE and Discharge	Slope	Sig. Level	n	PDO	MEI	PNA	PDO	MEI	PNA
date of freshet start	**-0.51**[1]	**0.03**	39	**-0.38**	**-0.56**	**-0.47**	**-0.60**	**-0.64**	-0.30
annual max. Q	-0.01	0.51	39	-0.11	-0.21	-0.30	-0.02	-0.04	-0.09
date of annual max. Q (JD)	-0.31	0.20	39	-0.09	**-0.35**	**-0.35**	-0.24	-0.30	-0.25
annual min. Q	0.00	0.93	39	0.11	-0.20	0.05	0.26	0.00	0.03
date of annual min. Q (JD)	0.76	0.44	39	0.10	0.16	0.19	0.07	0.15	0.10
total annual Q	-28869.21	0.58	39	-0.23	-0.30	-0.28	-0.06	-0.10	-0.01
average annual Q	0.00	0.58	39	-0.23	-0.30	-0.28	-0.06	-0.10	-0.01
April 1 SWE	1.95	0.60	48	**-0.4**	-0.28	**-0.41**	-0.33	-0.19	-0.11
Defined by Ablation Window									
date of snowmelt initiation	**-0.12**[1]	**0.01**	107	**-0.37**	**-0.37**	**-0.45**	**-0.51**	**-0.39**	**-0.43**
date of snowmelt completion	0.00	1.00	36	-0.20	**-0.43**	-0.12	-0.23	**-0.44**	-0.20
snowmelt season length	0.58	0.86	36	-0.01	-0.10	0.18	0.10	-0.15	0.20

[1] no trend remains after testing residuals from multiple regression analysis for trends
[2] trend significant at 0.1 following residual analysis

Table 4

	Trend Analysis			Teleconnection Correlations					
				DJF Indices			MAM Indices		
March-April-May (MAM) averages	Slope	Sig. Level	n	PDO	MEI	PNA	PDO	MEI	PNA
average melt	**0.02[3]**	**0.10**	36	0.31	**0.61**	0.29	**0.50**	**0.60**	**0.46**
# high magnitude melt events	0.16	0.23	36	0.26	**0.49**	0.17	0.30	**0.48**	0.31
average melt anomalies	**0.024[1]**	**0.10**	36	0.19	**0.54**	0.15	**0.33**	**0.55**	0.29
standard deviation of melt	0.01	0.33	36	0.16	0.29	0.10	0.10	**0.34**	0.23
maximum melt	0.05	0.26	36	0.40	**0.44**	0.33	0.32	**0.43**	**0.45**
# high magnitude melt events > 2 days	0.00	0.12	36	0.32	0.28	0.24	0.18	0.27	0.17
# high magnitude melt events => 2 days	0.00	0.44	36	0.07	0.25	0.13	-0.02	0.16	0.01
length of longest melt event	0.04	0.31	36	0.25	**0.54**	0.18	0.31	**0.51**	0.21
average sublimation	0.00	0.47	36	0.40	0.14	**0.37**	**0.43**	0.10	0.29
std. dev. sublimation	0.00	0.75	36	-0.30	-0.13	**-0.47**	-0.20	0.06	-0.07
maximum sublimation	0.00	0.74	36	0.07	-0.19	-0.21	0.09	-0.09	-0.03
minimum sublimation	0.00	0.43	36	0.34	0.13	**0.38**	0.28	-0.04	0.24
# high magnitude sublimation events	-0.11	0.34	36	-0.50	-0.19	**-0.54**	**-0.53**	-0.13	-0.26
% ablation season w/ high mag. sublimation	0.00	0.47	36	0.40	0.14	**0.37**	**0.43**	0.10	0.29
Values from radiosonde record									
850mb maximum air temperature	0.04	0.30	36	**0.37**	**0.45**	0.28	0.33	**0.40**	**0.44**
850mb minimum air temperature	0.01	0.89	36	0.32	0.23	0.17	0.29	0.21	**0.40**
850mb average air temperature	**0.03[3]**	**0.10**	36	**0.34**	**0.60**	0.30	**0.52**	**0.57**	**0.50**
850mb maximum vapour pressure	0.02	0.48	36	**0.36**	0.10	0.06	0.32	0.29	0.01
850mb minimum vapour pressure	0.01	0.71	36	0.32	0.14	**0.34**	**0.34**	0.07	**0.33**
850mb average vapour pressure	0.00	0.98	36	**0.48**	**0.44**	**0.34**	**0.59**	**0.40**	**0.45**

Table 5

	Trend Analysis			Teleconnection Correlations DJF Indices			Teleconnection Correlations MAM Indices		
Derived values	Slope	α	n	PDO	MEI	PNA	PDO	MEI	PNA
average melt	0.00	1.00	36	0.03	0.05	0.08	-0.06	-0.11	-0.16
std. dev. melt	0.00	0.99	36	0.01	-0.16	0.14	-0.23	-0.26	-0.25
maximum melt	0.01	0.81	36	0.30	-0.12	0.31	0.05	-0.20	-0.05
minimum melt	0.00	0.65	36	-0.01	0.02	0.01	0.04	0.01	0.02
% ablation season days w/ melt	0.03	0.76	36	0.13	0.24	0.23	**0.36**	0.25	0.32
average melt anomalies	0.00	1.00	36	0.03	0.05	0.08	-0.06	-0.11	-0.16
std. dev. melt anomalies	0.00	0.99	36	0.01	-0.16	0.14	-0.23	-0.26	-0.25
maximum melt anomaly	0.01	0.81	36	0.30	-0.12	0.31	0.05	-0.20	-0.05
minimum melt anomaly	0.00	0.65	36	-0.01	0.02	0.01	0.04	0.01	0.02
total melt anomalies	0.05	0.95	36	0.03	0.06	0.07	-0.05	-0.10	-0.23
# high magnitude melt events	0.00	1.00	36	-0.21	-0.07	0.14	-0.14	-0.20	-0.13
% ablation season w/ high mag. melt	-0.04	0.67	36	-0.21	-0.06	0.05	0.36	-0.21	-0.22
# high magnitude melt events > 2 days	0.00	0.68	36	-0.14	0.04	0.22	-0.04	-0.05	-0.07
# high magnitude melt events => 2 days	-0.03	0.26	36	-0.19	-0.10	0.09	-0.25	-0.31	-0.19
length longest melt period in ablation season	0.00	0.97	36	-0.06	0.19	0.10	0.02	0.02	-0.02
average sublimation	0.00	0.13	36	**0.38**	0.08	**0.42**	**0.40**	0.02	0.22
std. dev. sublimation	0.00	0.31	36	-0.10	0.02	-0.24	-0.09	0.13	0.05
maximum sublimation	0.00	0.91	36	0.02	-0.12	-0.02	0.16	-0.05	0.18
minimum sublimation	0.00	1.00	36	0.30	0.17	**0.39**	0.29	0.00	0.16
% ablation season days w/ sublimation	-0.08	0.51	36	-0.09	-0.17	-0.11	-0.20	-0.25	-0.25
average sublimation anomalies	0.00	0.12	36	**0.43**	0.10	**0.49**	**0.47**	0.02	0.24
std. dev. sublimation anomalies	0.00	0.69	36	0.14	0.09	0.07	0.19	0.11	0.13
maximum sublimation anomaly	0.00	0.41	36	**0.34**	0.06	**0.42**	**0.47**	-0.02	0.22
minimum sublimation anomaly	0.00	1.00	36	0.30	0.17	**0.39**	0.29	0.00	0.16
total sublimation anomalies	0.03	0.15	36	**0.41**	0.07	**0.46**	**0.46**	0.03	0.20
# high magnitude sublimation events	-0.06	0.45	36	**-0.44**	-0.07	**-0.46**	**-0.43**	-0.04	-0.07
% ablation season w/ high mag. sublimation	-0.20	0.28	36	**-0.39**	-0.05	**-0.49**	**-0.39**	0.02	-0.14
Values from radiosonde record									
850mb maximum air temperature	0.00	0.98	36	0.25	-0.10	0.32	0.00	-0.24	-0.11
850mb minimum air temperature	0.01	0.54	36	0.12	0.14	0.10	0.11	0.22	0.15
850mb average air temperature	-0.01	0.78	36	0.04	0.07	0.13	0.05	-0.06	-0.04
850mb maximum vapour pressure	0.00	0.53	36	0.10	-0.15	0.08	0.12	-0.12	0.11
850mb minimum vapour pressure	0.00	0.88	36	0.19	0.10	0.32	0.26	-0.04	0.18
850mb average vapour pressure	0.01	0.12	36	**0.37**	0.08	**0.42**	**0.40**	0.01	0.21

Table 6

Variable	Adjusted r^2	Beta MEI	PDO	PNA
MAM average melt	0.42	0.49	---	0.27
MAM # high magnitude melt events	0.20	0.48	---	---
MAM melt anomalies	0.28	0.55	---	---
MAM length of longest melt event	0.23	0.51	---	---
MAM # high magnitude sublimation events	0.28	0.25	-0.67	---
MAM maximum 850 mb potential temperature	0.21	0.26	---	0.34
MAM average 850 mb potential temperature	0.38	0.44	---	0.33
MAM average vapour pressure	0.34	---	0.50	0.18
date of freshet start	0.29	---	-0.35	---
date of snowmelt initiation	0.28	---	-0.41	-0.22
date of snowmelt completion	0.41	---	-0.53	-0.20

Figure 1

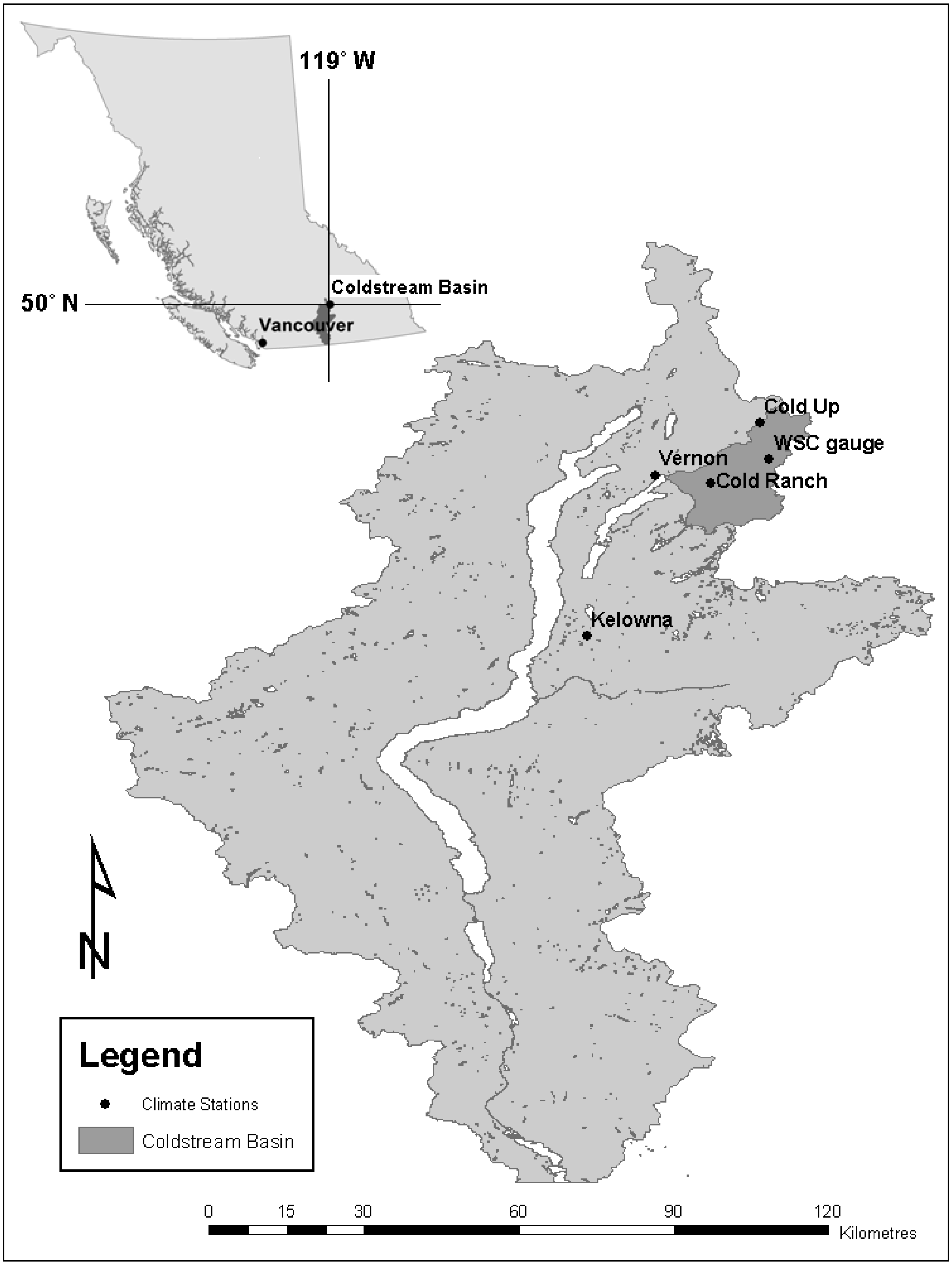

119° W
50° N
Coldstream Basin
Vancouver
Cold Up
WSC gauge
Vernon
Cold Ranch
Kelowna
N
Legend
Climate Stations
Coldstream Basin
0
15
30
60
90
120
Kilometres

Figure 2

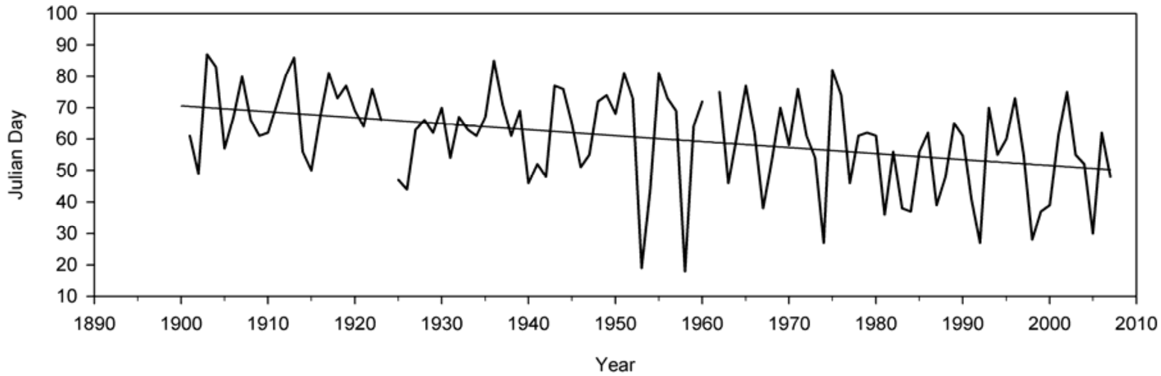
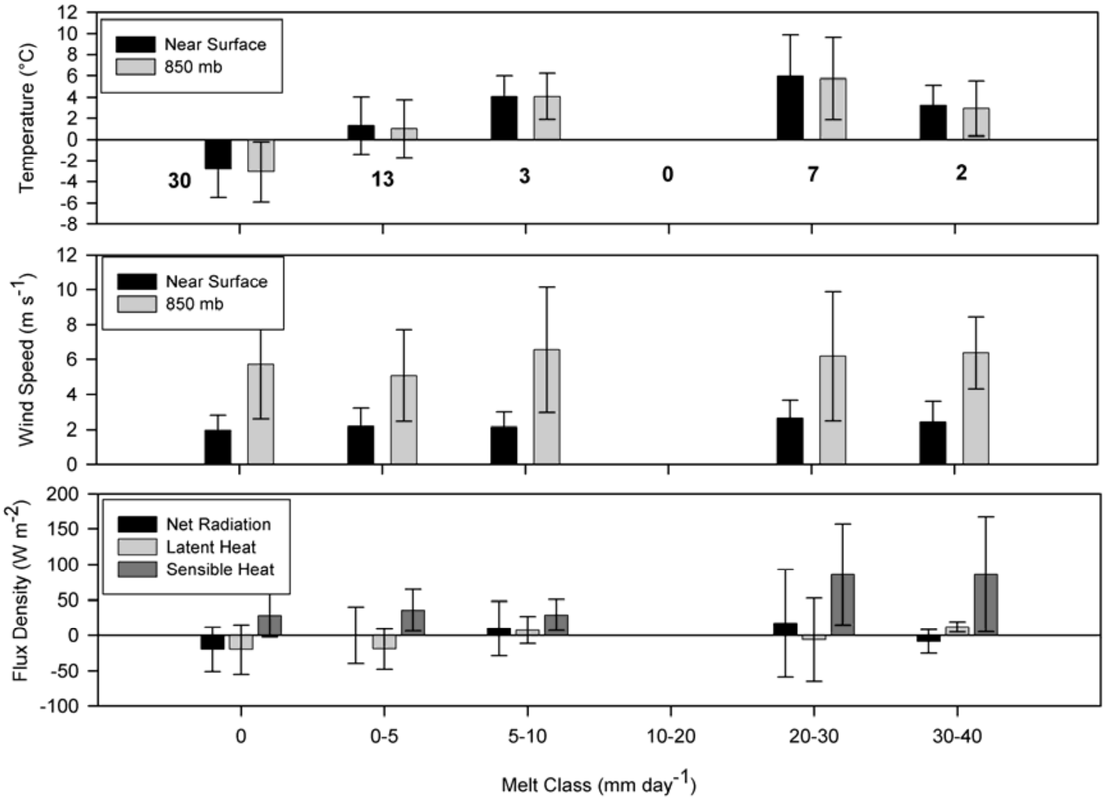

Julian Day
100
90
80
70
60
50
40
30
20
10
1890
1900
1910
1920
1930
1940
1950
1960
1970
1980
1990
2000
2010
Year

Figure 3



Temperature (°C)
Near Surface
850 mb
30
13
3
0
7
2
Wind Speed (m s-1)
Near Surface
850 mb
Flux Density (W m-2)
Net Radiation
Latent Heat
Sensible Heat
0
0-5
5-10
10-20
20-30
30-40
Melt Class (mm day-1)

Figure 4

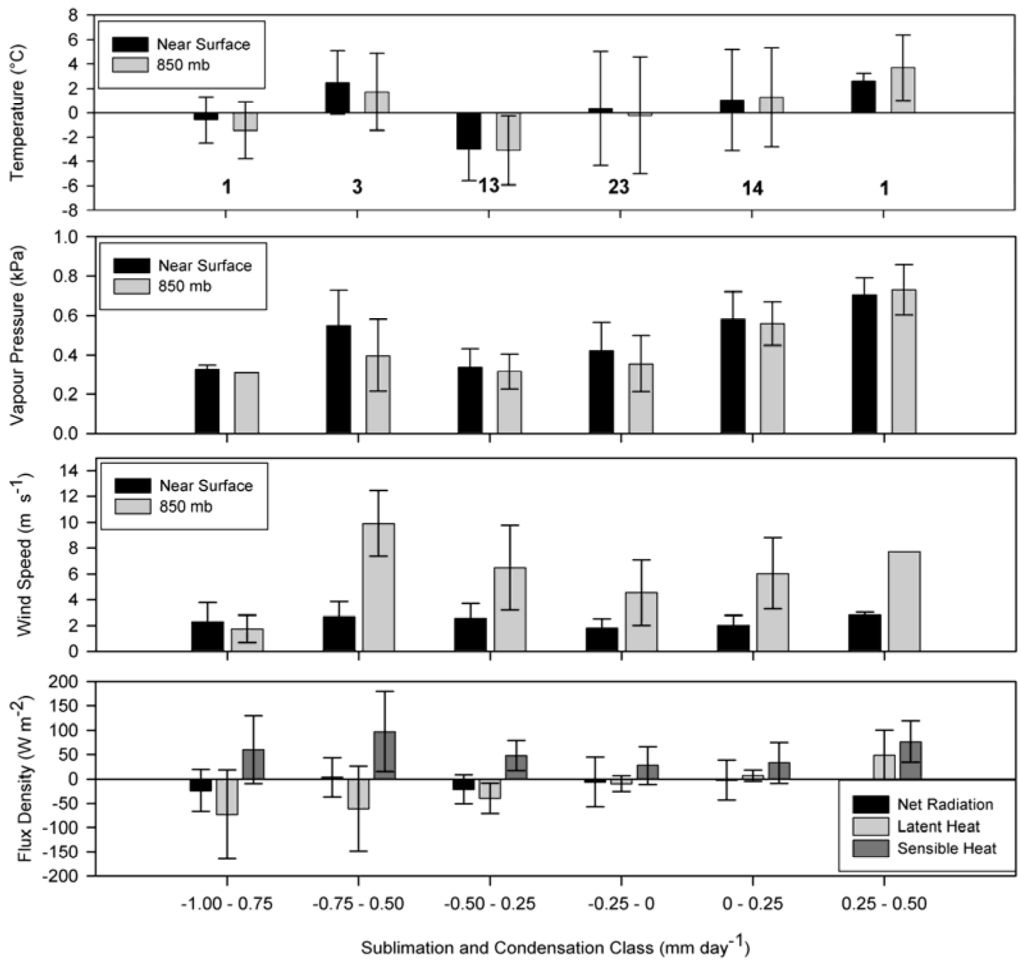
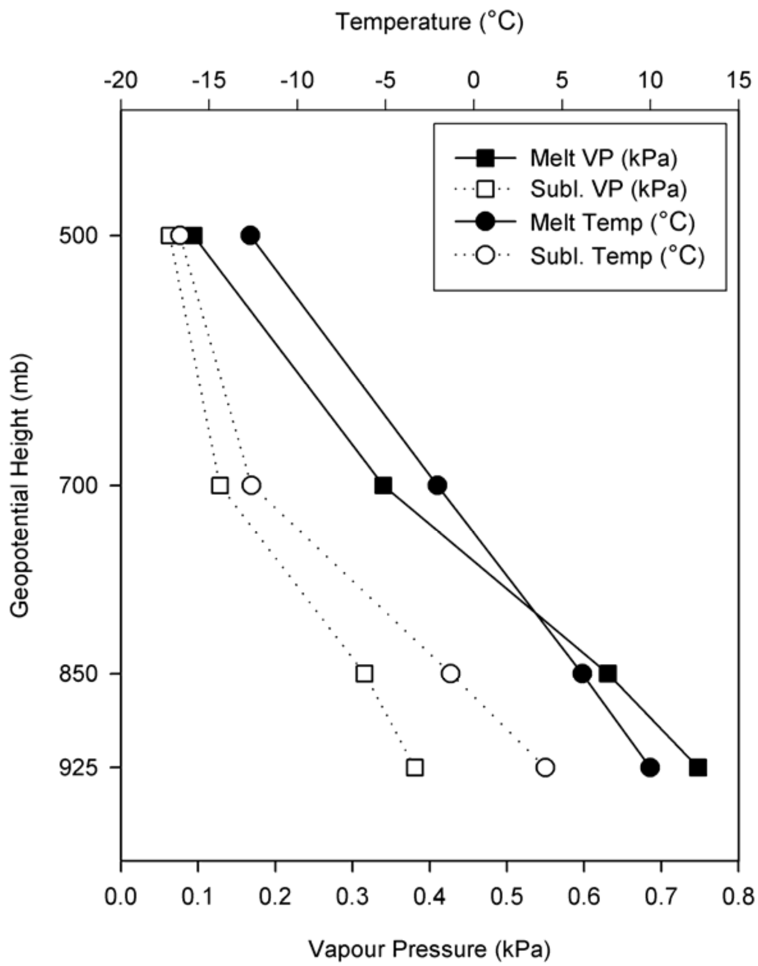

Temperature (°C)
Near Surface
850 mb
1
3
13
23
14
1
Vapour Pressure (kPa)
Near Surface
850 mb
Wind Speed (m s-1)
Near Surface
850 mb
Flux Density (W m-2)
Net Radiation
Latent Heat
Sensible Heat
-1.00 - 0.75
-0.75 - 0.50
-0.50 - 0.25
-0.25 - 0
0 - 0.25
0.25 - 0.50
Sublimation and Condensation Class (mm day-1)

Figure 5



Temperature (°C)
-20
-15
-10
-5
0
5
10
15
Melt VP (kPa)
Subl. VP (kPa)
Melt Temp (°C)
Subl. Temp (°C)
Geopotential Height (mb)
500
700
850
925
0.0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
Vapour Pressure (kPa)

Figure 6

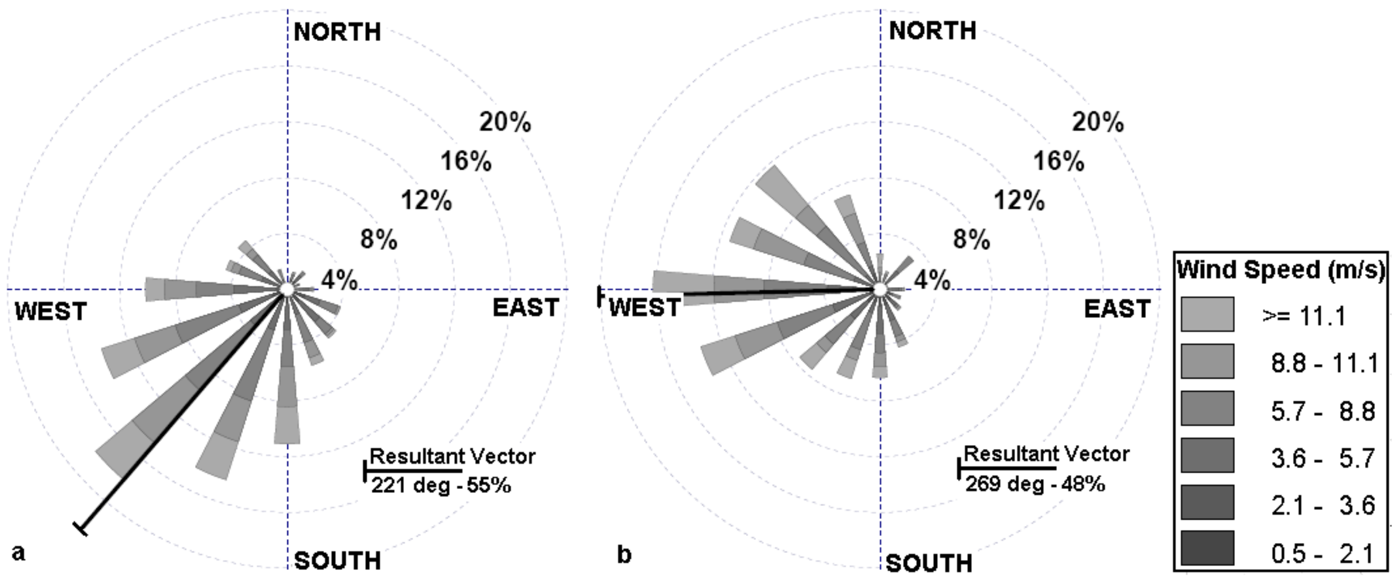
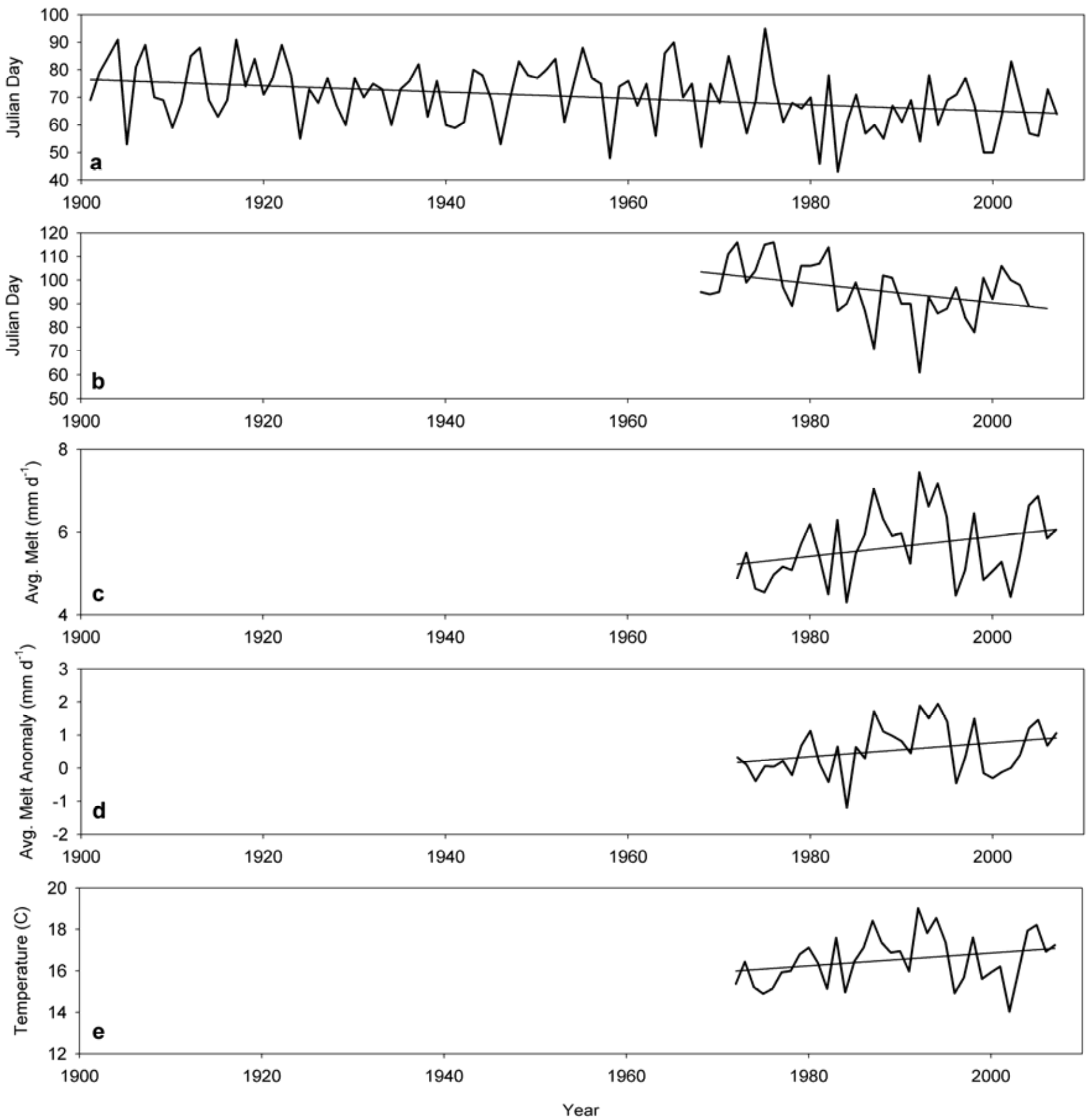

NORTH
20%
16%
12%
8%
4%
WEST
EAST
SOUTH
Resultant Vector
221 deg - 55%
a
NORTH
20%
16%
12%
8%
4%
WEST
EAST
SOUTH
Resultant Vector
269 deg - 48%
b
Wind Speed (m/s)
>= 11.1
8.8 - 11.1
5.7 - 8.8
3.6 - 5.7
2.1 - 3.6
0.5 - 2.1

Figure 7


100
90
80
70
60
50
40
Julian Day
a
120
110
100
90
80
70
60
50
Julian Day
b
8
6
4
Avg. Melt (mm d-1)
c
3
2
1
0
-1
-2
Avg. Melt Anomaly (mm d-1)
d
20
18
16
14
12
Temperature (C)
e
1900
1920
1940
1960
1980
2000
Year

CHAPTER 5: CONCLUSION

This thesis presents the first study of ablation season snow-energy balance processes in the Okanagan Basin. Variation in melt and sublimation rates were quantified along an elevational and forest-cover gradient, and statistically linked to the concurrent conditions in the atmospheric boundary layer. The first journal style manuscript presented in Chapter 3 describes the variation in snow energy balance processes along an elevational and forest cover gradient within the Coldstream Basin during the late-winter and spring of 2007.

Within the basin, peak annual SWE increased linearly with elevation and was typically reduced by half under forest canopy relative to the paired open site at the same elevation in 2007. Snow density increased (decreased) with elevation at the forest (open) sites. Over the 45 day ablation season, melt averaged 6.5 mm d^{-1} in the open sites and 4.0 mm d^{-1} under forest canopies, while sublimation averaged 0.4 mm d^{-1} in the open sites and 0.2 mm d^{-1} at the forest sites. This accounted for a 1-4% loss of peak SWE at the open sites, and 4-12% at the forested sites. Sublimation and melt rates increased significantly with elevation, but these processes showed no significant difference between the open and forested sites at the same elevation. At the higher elevations, greater snow accumulation in the open outweighed the higher rates of sublimation; however, this balance can be expected to shift as land cover changes in the future as a result of the ongoing mountain pine beetle epidemic.

During the early portion of the ablation season, high snowmelt rates were driven by large sensible heat fluxes and relatively lower net radiation. This balance shifted as the ablation season progressed and higher solar angles increased the proportion of shortwave radiation directed at the snow surface. Sensible heat was found to be the predominant driver of snow melt for this study period, far outweighing the influence of incoming radiation or latent heat transfers to the snowpack. High magnitude melt events were characterized by warm, moist air masses and strong easterly winds, while large vapour loss events were driven by colder, drier conditions and stronger than average southerly winds.

The second journal style manuscript outlines the statistical linkages between the near-surface snow energy balance and the conditions in the atmospheric boundary layer. It presents the results obtained from the analysis of a proxy record created using these linkages and the nearby radiosonde record, including the presence of significant trends in ablation season measures and correlations with the regionally dominant teleconnection indices.

An analysis of the proxy record found significant trends in the dates of snowmelt initiation and date of the freshet initiation. Average melt for the March-April-May period, annual melt anomalies and the 850 mb average potential temperature also exhibited positive trends. The links made between near-surface snow energy balance and local conditions in the atmospheric boundary layer were extended outward to the dominant teleconnection patterns influencing climate in British Columbia. Significant positive correlations were found between snowmelt event magnitude and frequency of high magnitude events and the DJF and MAM MEI, MAM PDO and PNA indices. Sublimation event magnitude and frequency of high magnitude events is correlated with the DJF PNA and MAM PDO, with negative phases associated with high magnitudes and frequencies of sublimation.

However, the meso-scale synoptic conditions that link the conditions in the ABL associated with high magnitude melt and sublimation events and the macro-scale teleconnections are still undefined. Future work should explore the links between the synoptic conditions associated with high magnitude events and the phase and strength of the various teleconnections that drive snowmelt and sublimation in the Okanagan Basin.

The results of the work presented in this thesis serve to significantly advance the understanding of ablation season snow-surface energy balance in the Okanagan Basin and the role that the synoptic climate plays in these energy and mass exchanges at the site scale. This knowledge is particularly crucial, as the Okanagan is a semi-arid basin, highly reliant on the spring snowmelt to supply the majority of its usable water, which is under intense pressure from both development and ongoing climate change. The findings outlined here will allow water managers to better understand the climatic conditions that are linked to higher sublimation and melt rates, and to predict seasonal water supply more accurately during future periods of rapidly changing land cover and climate. While the use of SNTHERM to calculate site-specific energy and mass balance on snow cover may be unrealistic at an operational level, the work presented here does have the potential to be applied in water-supply forecasting efforts. With very little work, the daily soundings of the ABL could be used in a regression model to provide an index of the melt and sublimation occurring throughout the Okanagan. This would allow water managers to more accurately predict reservoir recharge rates, volume of SWE lost to sublimation, and even flood levels in the local rivers.

While this work represents an important step towards a more robust quantification of the hydrological cycle in the Okanagan Basin, some unknowns remain. In particular, the shift towards larger open areas in forests due to the ongoing mountain pine-beetle salvage cutting will likely result in pronounced changes to the snow energy balance; as increased fetch lengths result in higher magnitudes of the turbulent fluxes - leading to increased sublimation losses. Additionally, the meso-scale synoptic conditions that link the conditions in the ABL associated with high magnitude melt and sublimation events and the macro-scale teleconnections are still undefined. Future work should explore the links between the synoptic conditions associated with high magnitude events and the phase and strength of the various teleconnections that drive snowmelt and sublimation in the Okanagan Basin.

To address these knowledge gaps, a long-term, high-elevation climate station network covering the range of climate and land-cover conditions in the Okanagan Basin is necessary. Additionally, greater spatial coverage of snow measurement sites is needed. The network, in combination with the foundational knowledge presented here, would enable regional water managers to better predict and plan for future variability in the timing and volume of the spring snowmelt and subsequent freshet that is vital to the replenishment of the Okanagan Basins surface water resources.