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Article

Net Snowpack Accumulation and Ablation Characteristics in the Inland Temperate Rainforest of the Upper Fraser River Basin, Canada

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Abstract: This study examines the net snow accumulation and ablation characteristics and trends in the Inland Temperate Rainforest (ITR) of the Upper Fraser River Basin, British Columbia (BC), Canada. It intends to establish whether elevation and/or air temperature play(s) a dominant role in hydrological year peak snow water equivalent (SWE) and whether regional patterns emerge in the interannual variability in peak accumulation. To that end, SWE and air temperature data from seven snow pillow sites in the Upper Fraser River Basin at elevations ranging from 1118 to 1847 m above sea level are analyzed to infer snowpack characteristics and trends for hydrological years 1969–2012, with 2005–2012 being the actual period of data overlap. Average peak SWE ranges from 391.3 mm at Barkerville, BC on 16 April to 924.4 mm at Hedrick Lake, BC on 27 April. Snow cover duration lasts 206–258 days, with snow onset dates from mid-October to early November and snow off dates from late May to early July. Statistically-significant ($p \leq 0.05$) cross correlations exist between peak SWE at nearly all sites, indicating regional coherence in seasonal synoptic activity across the study area. However, the lack of relationships between peak SWE and elevation as well as air temperature parameters indicate that mesoscale to local processes lead to distinct snow accumulation and ablation patterns at each site. Four sites with the longest records exhibit no trend in peak SWE values between 1990 and 2012. Changes to snowpack regimes may pose a threat to the productivity and immense biodiversity supported by the ancient western red cedar and hemlock stands growing in the wet toe slopes of the ITR. Thus, it is imperative that continued monitoring of snowpack conditions remains a top priority in the Upper Fraser River Basin, allowing for a better understanding of ecosystem changes in a warming climate.

Keywords: snowpack; snow accumulation; snow ablation; Inland Temperate Rainforest; climate change; ecohydrology; Upper Fraser River Basin; British Columbia

1. Introduction

Snow forms an important component of the hydrological cycle and climate of high latitude and mountainous regions (e.g., [1,2]). As an example, the Upper Fraser River Basin of British Columbia (BC), Canada, is a snow-dominated system with snowmelt resulting in an annual pulse of freshwater each spring and early summer that periodically leads to flooding. This region is dominated by abundant wintertime snowfall throughout northern BC's mountainous terrain and vast forests. It is also the site of the Inland Temperate Rainforest (ITR), an ecosystem unique to BC that is characterized by its continentality and anomalously humid climate. The continentality is depicted by the same weather systems and precipitation patterns that nourish the coastal rainforests of BC but in a cooler climate regime, creating a secondary zone of high precipitation as they cross the interior mountain ranges [3]. Confined to the wettest subzones of the Interior Cedar-Hemlock zone (*i.e.*, biogeoclimatic zones ICHwk and ICHvk [4]), the ITR experiences plentiful snowmelt during late spring that is followed by ample rainfall during the height of the growing season [5,6]. The ITR shelters old-growth forests, with some trees surpassing 1000 years of age, which are largely found in valley-bottom to mid-slope positions on the windward slope of the interior mountain ranges, between latitudes 50°N and 54°N [3]. In particular, cedars and hemlocks are abundant at the toe slopes, whereas old-growth Engelmann spruce and subalpine fir (ESSF) thrive at higher elevations in the subalpine above 1500 m above sea level (a.s.l.) [3,6]. The ITR is ≈700 km inland from the coast of the Pacific Ocean and differs from BC's coastal rainforests because most of its annual precipitation falls as snow (especially at elevations above 1000 m a.s.l.). Snowpacks are moderate to deep in the ITR, with the wettest portions accumulating up to 2 m of settled snow in mid to late winter and much deeper accumulation occurring at higher elevations in the ESSF zone [3]. Deep snowpacks extend the snowmelt period over the summer months, which are thought to gradually replenish soil moisture in wet toe slope topographic positions, where ancient western redcedar and hemlock stands are found [3]. In addition, the abundant snowmelt infiltration in this region sustains groundwater supply and minimizes soil moisture deficit during dry summer periods, making stands less susceptible to fires and insect mortality [6].

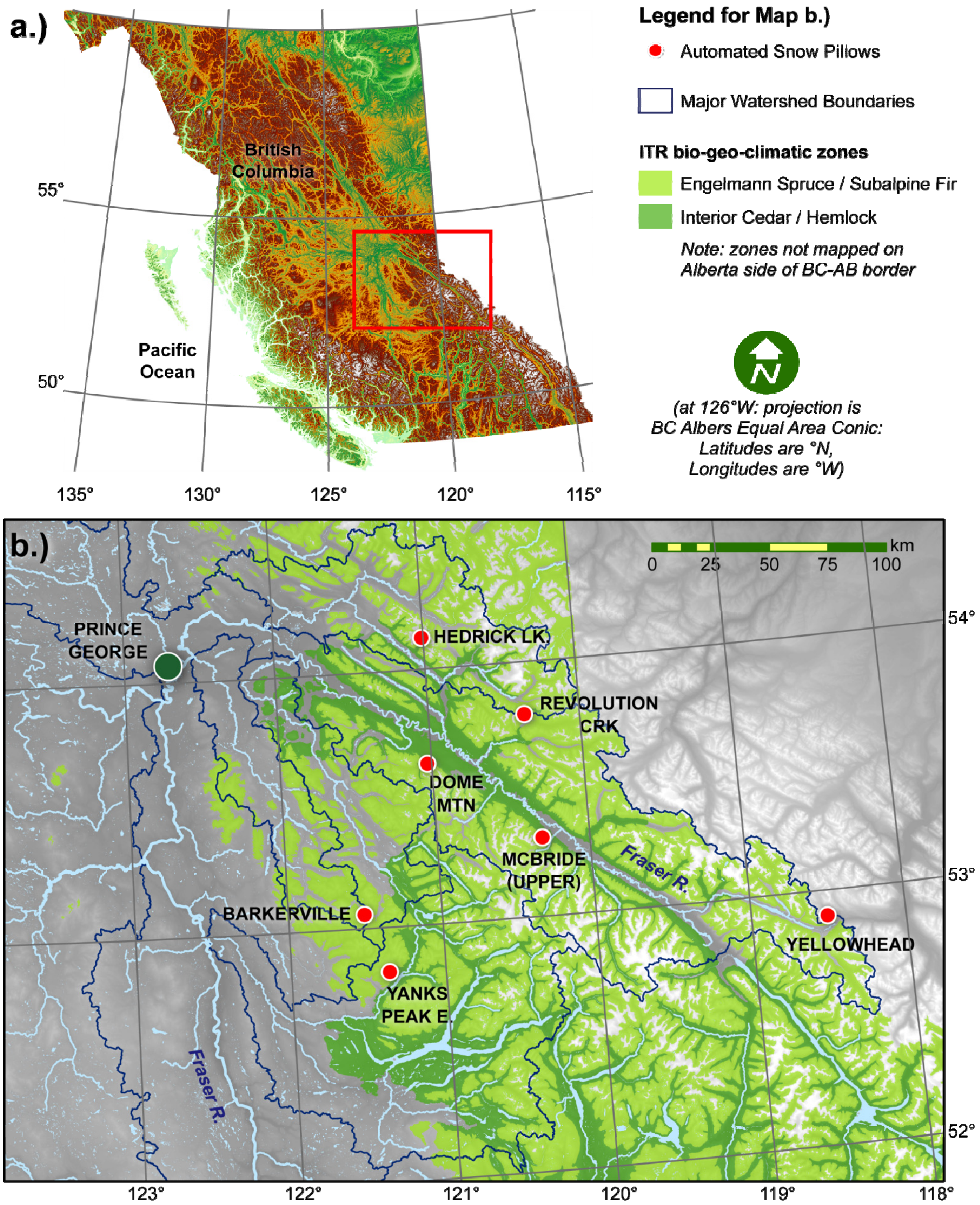
Despite the important role of snow in its hydrology, no study has explored net snowpack accumulation and ablation characteristics of the Upper Fraser River Basin. Previous research on western North American snowpack characteristics focused on the Sierra Nevada [7], Colorado Rockies [8], and the Pacific Northwest [9] including the headwaters of the Columbia River Basin in BC [10]. In western North America, climate change is one of the main drivers of declining mountain snowpacks; however, land use modifications and other factors also impact winter snowpacks [11]. A regional analysis of snowpack metrics for the interior western USA indicated widespread decreases in the duration of snow cover throughout the intermountain west, reduced maximum Snow Water Equivalent (SWE), and faster melts in the Colorado River and Rio Grande Basins in response to widespread warming [12]. Forest harvest practices, such as partial or clear cutting, can result in greater accumulation of total SWE in the winter snowpack [13,14]. Mountain pine beetle killed forests exhibit higher winter snowpack accumulation [15], more rapid snowmelt [16], and increased sublimation [17]. Following wild fires, canopy loss can increase net winter season snow sublimation and post-burn areas may experience increased winter ablation that can lead to reduced snow water inputs relative to healthy forests [18]. Danard and Murty [19] observed declining snowpack accumulations from 1966 to 1989 along with warming trends in the Fraser and Skeena River Basins of BC. Sites along the middle Fraser River were found to experience sharp declines in final day of the month snow depths between 1947 and 2003 [20]. In addition, Moore and McKendry [21] established regional snow accumulation anomalies and their relationship to atmospheric circulation patterns across BC. Snowpack evolution in the Quesnel River Basin of BC has been investigated using passive microwave remote sensing data; however, their application is limited by the region's deep snowpacks that cause the remote sensing products to become saturated [22]. A study conducted by Shrestha *et al.* [23] reported declining contributions of snow in 21st century runoff projections in the Fraser River Basin. The latter findings are in accord with those of Morrison *et al.* [24] and Kerkhoven and Gan [25] where it is suggested that the Fraser River may transition from a snowmelt- to a rainfall-dominated regime by 2100. Thus, climate change is expected to alter SWE amounts and timing of the spring melt in the Upper Fraser River Basin, which may significantly impact its ecosystems including that of the ITR.

Thus, the objective of this study is to quantify net snowpack accumulation and ablation characteristics across the Upper Fraser River Basin with a focus on the implications to the ecohydrology of its ITR. Some of the research questions motivating this effort are: (1) Does elevation and/or air temperature play(s) a dominant control on peak SWE accumulation in the study area?; (2) Is there consistent regional interannual variability in peak SWE values?; and (3) What may be the implications of observed changes in snowpack accumulation and ablation characteristics to the ecohydrology of the ITR, particularly in the context of climate change? To address these questions, daily SWE and air temperature data from seven snow pillow sites in the study area are used to explore snowpack characteristics such as annual peak accumulation and its timing, net accumulation and ablation rates, and duration with emphasis on their interannual variability. Basic information on these variables is needed to better understand the impacts of changing snow regimes on ITR dynamics and evolution. In particular, we investigate the sensitivity of peak SWE accumulation and its timing to air temperatures and then place this in the context of projected long-term regional warming and implications to old growth trees in the ITR.

2. Study Area

The Upper Fraser River Basin of BC located upstream of the Nechako River spans ≈35,000 km^2 at a mean elevation of 1413 m a.s.l., forming the headwaters of the Fraser River, the largest Canadian river flowing into the Pacific Ocean (Figure 1; [26]). The Upper Fraser River flows northwestward from the Rocky Mountains into the Rocky Mountain trench prior to veering southward near the city of Prince George, BC where its annual discharge rate is ≈26 $km^3 \cdot yr^{-1}$ [26]. The watershed's highest peak is Mount Robson at 3954 m a.s.l. whereas its lowest level is in Prince George, BC at 575 m a.s.l. [27]. Mixed deciduous (aspen, willow and birch trees) and coniferous (lodgepole pine, subalpine fir and Engelmann spruce) forests cover the basin with meadows and rocky terrain in alpine areas. In addition, ancient western redcedar stands are typically found at toe slopes of the Cariboo and Rocky Mountains [3]. These old growth stands are part of the ITR that is characterized by abundant precipitation with a large fraction arriving as snowfall.

Figure 1. (**a**) Map of British Columbia showing the study area (red rectangle) and the Fraser River Basin (blue outline) and (**b**) Location of the seven snow pillow sites in the Upper Fraser River Basin (blue line) used in the present study.



The climate of the Upper Fraser River Basin is characterized by relatively cold, snowy winters as well as warm and occasionally wet summers throughout the entire basin. The 1971–2000 mean annual air temperature at Prince George, BC is 4.0 °C and total annual precipitation is 601 mm with 36% arriving in the form of snow (Environment Canada, 2014, https://weather.gc.ca/). Air temperatures decline with altitude, but precipitation generally increases at higher elevations, with greater fractions of the precipitation falling as snow. In the ITR, elevations range from 400 to 1500 m a.s.l., mean annual air temperatures range from 2.7 to 4.5 °C, and the mean annual precipitation is 788–1240 mm [3]. Snow covers the ground for about four months in the valley bottom of the basin and exhibits a linear increase in duration with elevation, with permanent snow fields and glaciers above ≈2400 m a.s.l. [28]. Frequent snowfalls lead to the development of a significant snowpack, with maximum SWE accumulations exceeding 800 mm at elevations above 1500 m a.s.l. [22]. The onset of snowmelt in spring induces an annual pulse of freshwater in the Upper Fraser River and its vast network of tributaries [29] making snowpack monitoring by the BC River Forecast Centre most essential.

The Upper Fraser River Basin is important to the economy of western Canada with many active resource extraction industries (e.g., mining and forestry). Other important economic drivers in the basin include agriculture, recreation and tourism. The Fraser River is also one of the most productive salmon rivers in the world with up-river migrations in the millions each year [26,30]. Five species of salmon migrate and spawn in the Fraser River and its many tributaries, forming important resources for commercial and recreational fisheries as well as a source of sustenance for First Nations communities in the area. It is also a region where air temperatures have warmed by ≈1 °C since the early 1940s, possibly leading to warmer winters during which less precipitation will fall as snow that may in turn negatively impact snowpack levels [27]. There is thus an urgent need to better understand net snowpack accumulation and ablation characteristics in the ITR of the Upper Fraser River Basin.

3. Data and Methods

Data from seven BC River Forecast Centre snow pillow sites in the Upper Fraser River Basin are used in the present study (Table 1). Data used include daily SWE as well as minimum and maximum air temperature. The period of data availability varies across sites and spans a minimum of seven winters with the earliest records beginning in 1968 at Barkerville, BC. The mean daily air temperature is estimated from the average of the daily minimum and maximum air temperatures. Data are then verified for completeness and gaps of seven days or less are in-filled through linear interpolation over time. Hydrological years (defined here as 1 September to 31 August of the following year since accumulation occasionally begins in September in the study area) with longer data gaps are eliminated from the analyses. Longer periods of missing air temperature data, often during summer, are in-filled with the daily mean for the same day over the period of record. A 5-day moving average of daily SWE and air temperature is then applied to filter out short-term fluctuations in these quantities arising from synoptic activity and possible instrumental errors (e.g., [31]).

Several statistics are then compiled from the time series of 5-day moving averages of SWE and air temperature. This includes the mean and standard deviation of snow season length, snow onset and off dates, peak accumulation and timing, April 1 SWE, net accumulation and ablation rates, air temperature conditions during these periods, and melt factor (*i.e.*, degree-day factor) over the sites' respective periods of data availability. Continuous snow cover was defined to occur for SWE values ≥20 mm. Degree day factors (a.k.a. melt coefficients with units of mm·°C^{-1}·day^{-1}) are the amounts of snowmelt that occurs per positive degree day and are thus assessed by tracking the air temperatures ≥0 °C at each site when SWE decreases [32]. Cycles of the mean daily SWE and air temperature at the seven sites over the course of a hydrological year are also presented. Cross correlations between time series of peak SWE values at each site for overlapping periods are also established. In addition, correlations between peak SWE values and air temperature conditions during the accumulation and ablation periods are computed. All correlation values are considered statistically-significant when $p \leq 0.05$. Further analyses investigate the relationships between snowpack characteristics and geographical location (latitude, longitude and elevation). Trends (considered statistically-significant when $p \leq 0.05$) in peak SWE values and their timing are then established using linear regressions for four sites (Barkerville, Revolution Creek, Yanks Peak East and Yellowhead) with the longest temporal records.

Table 1. Information on the BC River Forecast Centre snow pillow stations (locations shown in Figure 1) used in this study.

Station Name	Station I.D.	Latitude (°N)	Longitude (°W)	Elevation (m a.s.l.)	Mean Annual Peak SWE (mm)	Years of Data Availability
Barkerville	1A03P	53.05	121.48	1483	391.3	1968–2012[1]
Dome Mountain	1A19P	53.62	121.02	1768	893.4	2005–2012
Hedrick Lake	1A14P	54.10	121.00	1118	924.4	1999–2012
McBride (Upper)	1A02P	53.30	120.32	1608	562.6	1971–2012[2]
Revolution Creek	1A17P	53.78	120.37	1676	882.6	1984–2012
Yanks Peak East	1C41P	52.82	121.35	1683	900.5	1996–2012
Yellowhead	1A01P	52.90	118.53	1847	582.2	1997–2012

Note: [1]Data from 29 November 1988 to 25 August 1996 are missing. [2]Data from 4 June 1986 to 7 July 2006 are missing.

Even though the snow pillow data of SWE have some limitations (see Section 5.3), it is critical to assess SWE across the Upper Fraser Basin based on seven snow pillow sites over the past few decades. Thus, it allows for the mean characteristics, interannual variability, and long-term trends over the hydrological year to be established for this region. In addition, this study on the evolution of snowpack conditions in a very remote and undersampled region provides crucial information to begin understanding the ITR's ecohydrology and dynamics in a highly variable and rapidly changing environment.

4. Results

The mean hydrological year cycle of daily SWE exhibits smooth, nearly linear trends during the accumulation period (Figure 2). Accumulation of snow begins as air temperatures transition to subfreezing conditions (Figure 3). Once formed (SWE ≥ 20 mm), the snowpack remains continuous over time and is marked in individual hydrological years by rapid increases from snowfall events (not shown). Peak accumulation averages ≈750 mm SWE at the seven sites and typically occurs in mid- to late April when there is then a rapid decline in SWE until the termination of melt. Air temperatures approach 0 °C at peak accumulation and remain above this value during snow melt.

Figure 2. Hydrological year cycle of daily mean SWE at seven snow pillow sites in the Upper Fraser River Basin over their respective periods of data availability. For the plot of average SWE, the red, orange, green and blue vertical lines denote the snow onset date, April 1, day of maximum SWE, and snow offset date, respectively whereas the cyan and magenta horizontal lines denote the snow accumulation and ablation periods, respectively.

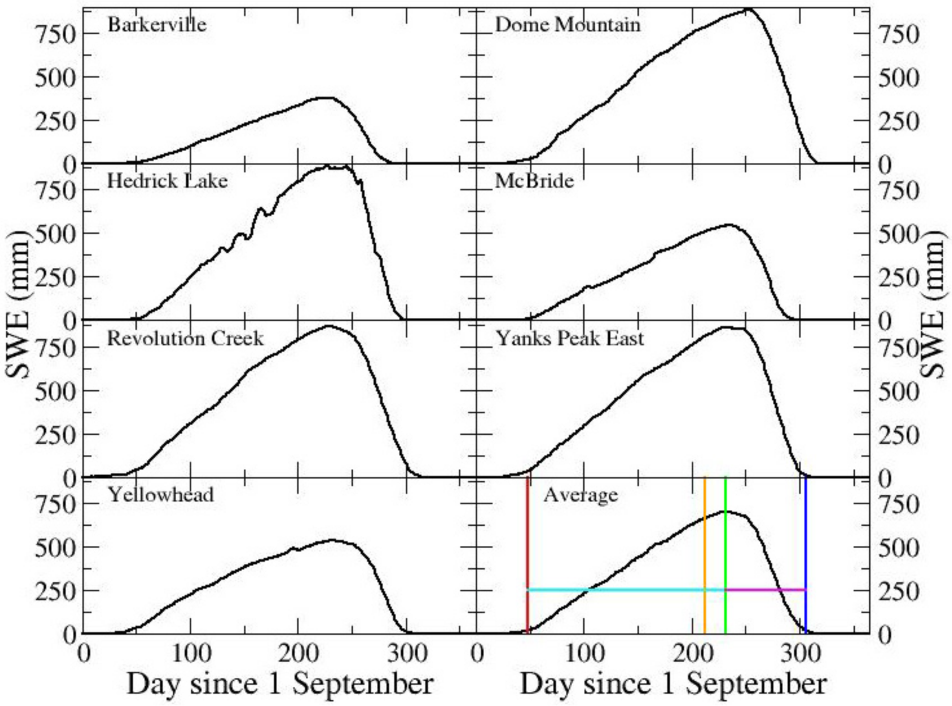


Figure 3. Hydrological year cycle of daily mean air temperature at seven snow pillow sites in the Upper Fraser River Basin over their respective periods of data availability.

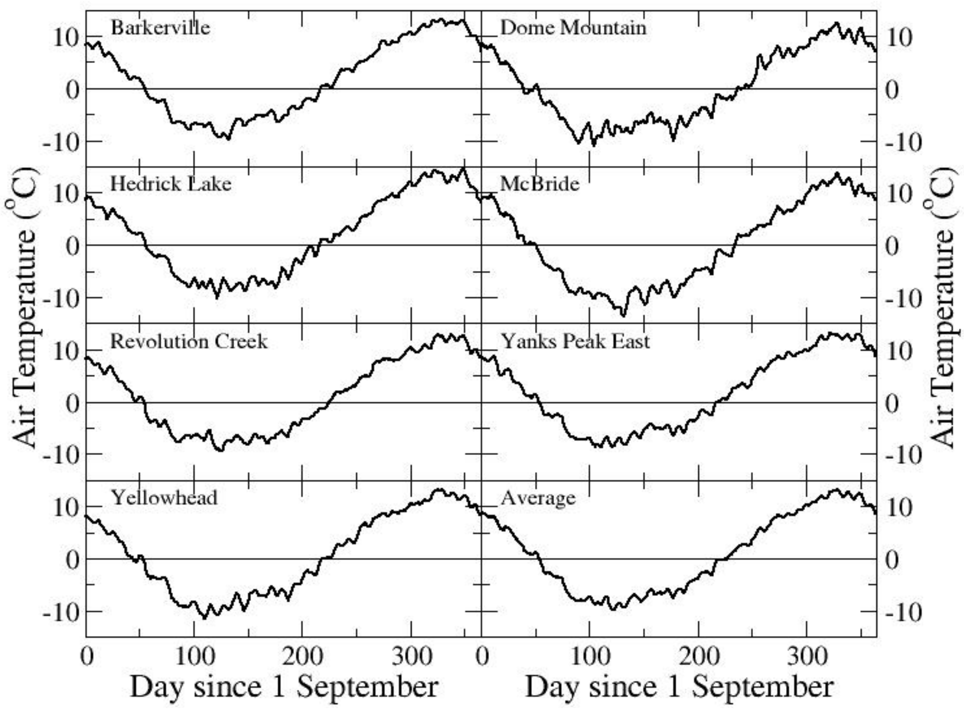


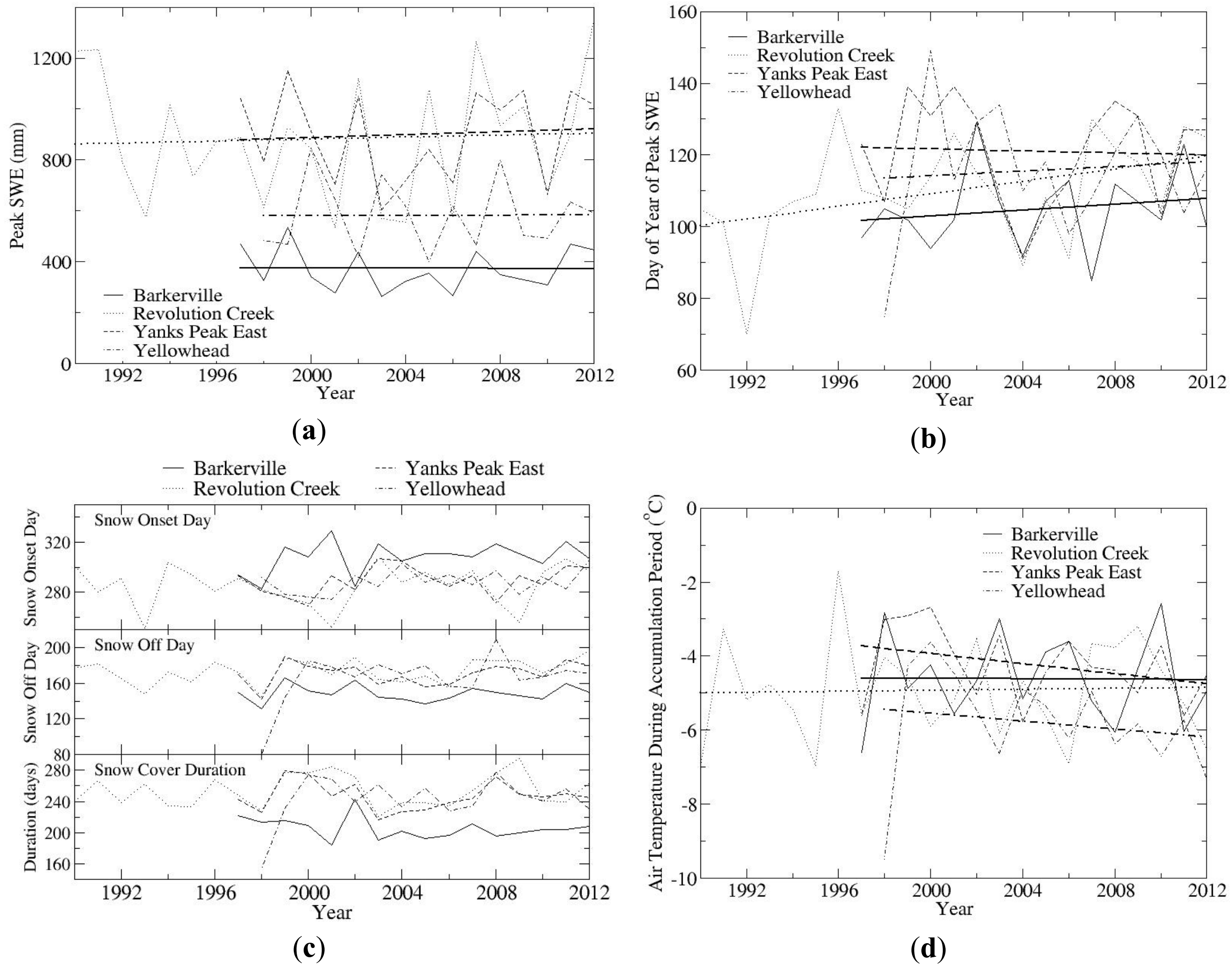
Table 2 provides overall statistics for snowpack characteristics at the seven sites of interest over their respective periods of record. Hydrological year maximum SWE varies between 391.3 mm at Barkerville and 924.4 mm at Hedrick Lake, with a large degree of interannual variability. Peak accumulation dates vary from 16 April at Barkerville to 12 May at Dome Mountain with standard deviations of about two weeks. Mean snow onset dates range from 12 October at Revolution Creek to 5 November at Barkerville, whereas mean snow off dates range from 30 May at Barkerville to 3 July at Dome Mountain. This implies the duration of snow cover extends from 206 to 258 days at the study sites. The period of net accumulation typically lasts three to four times longer than the period of net ablation. After peak accumulation, the mean net ablation rate varies between 8.8 and 21.7 mm·day^{-1} with a mean degree day factor ranging from 1.4 to 5.2 mm·day^{-1}·°C^{-1}.

Table 2. Results of the SWE analyses.

Variable Mean and [SD] (Units)	Barkerville	Dome Mtn.	Hedrick Lake	McBride (Upper)	Revolution Creek	Yanks Peak East	Yellowhead
Annual Maximum SWE (mm)	391.3 [79.7]	893.4 [219.7]	924.4 [205.8]	562.6 [122.5]	882.9 [247.1]	900.5 [172.9]	582.2 [132.6]
Annual Day of Year of Maximum SWE (mm)	105.8 (16 Apr) [10.9]	131.7 (12 May) [5.1]	117.4 (27 Apr) [12.0]	115.9 (26 Apr) [7.0]	110.0 (20 Apr) [14.2]	121.0 (1 May) [14.0]	115.7 (25 Apr) [16.6]
April 1 Annual SWE (mm)	363.9 [75.8]	782.1 [208.5]	851.7 [213.7]	509.0 [110.8]	829.8 [234.6]	808.1 [144.4]	507.7 [183.5]
Snow Onset Day	309.2 (5 Nov) [13.1]	290.6 (18 Oct) [10.0]	305.8 (2 Nov) [13.6]	302.7 (30 Oct) [11.6]	285.3 (12 Oct) [16.0]	289.3 (16 Oct) [11.0]	288.3 (15 Oct) [9.1]
Snow Offset Day	150.4 (30 May) [9.7]	183.9 (3 Jul) [9.4]	163.0 (12 Jun) [9.0]	163.0 (12 Jun) [6.8]	173.7 (23 Jun) [13.2]	170.9 (21 Jun) [15.1]	165.4 (14 Jun) [27.1]
Duration of Snow Cover (days)	206.4 [16.8]	258.4 [9.2]	222.4 [17.7]	225.5 [14.3]	253.6 [20.1]	246.8 [17.3]	242.3 [28.5]
Net Accumulation Period (days)	161.9 [19.2]	206.3 [12.3]	176.8 [18.8]	178.4 [14.3]	189.9 [22.6]	196.9 [21.1]	192.7 [21.0]
Net Ablation Period (days)	44.6 [11.9]	52.1 [11.1]	45.6 [15.4]	47.1 [9.3]	63.7 [13.1]	49.9 [9.8]	49.7 [19.0]
Ablation Rate (mm·day^{-1})	8.8 [2.5]	16.7 [2.4]	21.7 [7.2]	12.0 [3.7]	13.8 [3.6]	18.3 [4.5]	16.9 [20.6]
Degree Day Factor (mm·day^{-1}·°C^{-1})	2.2 [1.3]	2.8 [0.2]	4.0 [1.3]	5.2 [8.4]	3.5 [2.0]	2.4 [0.5]	1.4 [0.6]

The four sites with the longest time series of net snow accumulation show no recent trends in peak SWE (Figure 4a). There is less covariability between peak SWE at Yellowhead and values observed at the other sites owing in part to its more eastern and distant location in the Canadian Rockies (see Figure 1). The snow pillow site at Revolution Creek exhibits a statistically-significant trend toward a later date for the occurrence of peak accumulation (Figure 4b). Time series of snow onset dates, snow off dates, snow cover duration and mean air temperature during the accumulation period exhibit mixed trends (none however at $p < 0.05$) at the four sites with extended temporal coverage (Figure 4c,d). Peak SWE is significantly correlated with the date on which it occurs at Yanks Peak East ($r = 0.66$) and positively but not significantly correlated at other sites, partially explaining the tendency toward later occurrences of this event. A correlation analysis reveals no statistically-significant relationships between peak SWE and either latitude ($r = 0.53$), longitude ($r = 0.19$) or elevation ($r = -0.11$). The cross correlation analysis of peak SWE between sites shows a high degree of coherence (Table 3), indicating that wintertime synoptic activity affects the sites in a similar fashion.

Figure 4. Hydrological year (**a**) peak SWE time series and linear trends; (**b**) day of peak SWE time series and linear trends; (**c**) snow onset days, snow off days, and snow cover duration time series; and (**d**) annual mean daily air temperature during the accumulation period time series and linear trends at four snow pillow sites in the Upper Fraser River Basin over their respective periods of data availability, 1990–2012.



Mean annual air temperatures at the seven sites of interest are all near 0 °C despite their 729 m range in elevations (Table 4). Snow onset days occur a week or two after the day when air temperatures fall below the freezing point whereas there is a 1–2 month delay in snow off dates after air temperatures rise above the melting point. Thus, the duration of the period with air temperatures <0 °C is about a month less than the period with the presence of snow cover. Mean air temperatures during the accumulation period range from −7.4 °C at McBride (Upper) to −4.2 °C at Yanks Peak East whereas during the ablation period they range from 4.3 °C at McBride (Upper) to 7.2 °C at Yanks Peak East.

Table 3. Cross correlation coefficients for hydrological year peak SWE at seven snow pillow sites. Bold values denote statistically-significant results and *n* is the number of values used in the analyses.

Station Name	Barkerville	Dome Mountain	Hedrick Lake	McBride (Upper)	Revolution Creek	Yanks Peak East	Yellowhead
Barkerville		**0.85**	**0.92**	**0.59**	**0.71**	**0.93**	**0.75**
Dome Mountain	*n* = 6		**0.91**	**0.95**	**0.98**	**0.82**	0.73
Hedrick Lake	*n* = 12	*n* = 7		**0.82**	**0.88**	**0.90**	**0.64**
McBride (Upper)	*n* = 12	*n* = 6	*n* = 6		**0.94**	0.52	**0.84**
Revolution Creek	*n* = 15	*n* = 7	*n* = 13	*n* = 6		**0.77**	**0.73**
Yanks Peak East	*n* = 15	*n* = 7	*n* = 13	*n* = 6	*n* = 16		**0.56**
Yellowhead	*n* = 14	*n* = 7	*n* = 13	*n* = 6	*n* = 15	*n* = 15	

Table 4. Results of the air temperature analyses.

Variable Mean and [SD] (Units)	Barkerville	Dome Mtn.	Hedrick Lake	McBride (Upper)	Revolution Creek	Yanks Peak East	Yellowhead
Annual Air Temperature (°C)	1.7 [1.0]	0.1 [0.7]	1.8 [0.6]	-0.1 [1.3]	1.0 [0.9]	1.6 [0.7]	0.8 [0.7]
Date with Air Temperature <0°C	290.5 (17 Oct) [10.7]	274.9 (2 Oct) [6.0]	293.6 (21 Oct) [10.7]	282.6 (10 Oct) [6.7]	284.0 (11 Oct) [9.6]	285.4 (12 Oct) [8.4]	281.7 (9 Oct) [7.0]
Date with Air Temperature >0°C	108.2 (19 Apr) [12.5]	130.7 (11 May) [6.7]	111.5 (22 Apr) [11.2]	127.0 (7 May) [19.3]	117.4 (27 Apr) [12.9]	115.3 (25 Apr) [10.3]	115.7 (25 Apr) [10.6]
Duration of Air Temperature <0°C	182.9 [20.3]	221.0 [9.6]	183.1 [18.1]	209.6 [22.8]	198.7 [16.0]	195.1 [14.8]	199.3 [13.4]
Air Temperature During Net Accumulation Period (°C)	−5.3 [1.4]	−5.4 [1.1]	−4.9 [1.2]	−7.4 [1.8]	−4.9 [1.3]	−4.2 [0.9]	−5.8 [1.4]
Air Temperature During Net Ablation Period (°C)	4.6 [1.4]	6.6 [0.5]	6.3 [1.0]	4.3 [2.2]	5.6 [2.4]	7.2 [1.1]	6.1 [3.2]

Correlation analysis between peak SWE and air temperature variables reveals few relationships between these conditions. There are statistically-significant anti-correlations between peak SWE and air temperature during the accumulation period at Barkerville (r = −0.42) and Hedrick Lake (r = −0.68). In addition, there are statistically-significant correlations between peak SWE and air temperature during the ablation at Revolution Creek (r = 0.46) and Yanks Peak East (r = 0.80).

5. Discussion

5.1. Analysis and Synthesis of Results

The quantitative results presented in this study provide much needed information on the snowpack accumulation and ablation characteristics within the Upper Fraser River Basin of western Canada that explain in part the existence of the region's ITR. Based on the available snow pillow data, we find no relationship between hydrological year peak SWE and site elevation or most air temperature parameters for the accumulation and ablation seasons; however, a clear regional pattern emerges in the interannual variability of peak snow accumulation.

The lack of statistically-significant correlations between the observed peak SWE and either elevation, latitude or longitude shows that local effects are important in establishing maximum accumulation at a given snow pillow site. For instance, the orographic enhancement of precipitation partly depends on the local slope and topography, exposure to the dominant wind direction, and mountain circulations. Déry *et al.* [33] determined that wind redistribution of snow is also common at exposed sites in the study area. Thus, elevation is not the sole factor controlling peak accumulation observed in the ITR. Nonetheless, the spatial coherence in peak SWE clearly reveals that the dominant control on its interannual variability is synoptic-scale activity affecting the region. The frequency and intensity of storms affecting the Upper Fraser River Basin is modulated in part by phases of the Pacific Decadal Oscillation (PDO) and El Niño/Southern Oscillation (ENSO) [10,21,34]. For instance, the cool phases of the PDO and La Niña events are known to enhance cool season precipitation, and hence snowfall accumulation, in central BC. Some relatively lower cross correlation values in the annual peak SWE values between sites are nonetheless noted in Table 3. As an example, the relatively poor correlation ($r = 0.52$) between hydrological year peak SWE at Yanks Peak East and McBride (Upper) may be the result of their windward *versus* leeward locations of the Cariboo Mountains, respectively, despite their general proximity (see Figure 1). Thus, while synoptic-scale activity controls year-to-year variations in peak snow accumulation in the Upper Fraser River Basin, local scale effects remain important when point-scale observations of SWE are assessed.

There is a general lack of relationship between snowpack characteristics and air temperature at the seven snow pillow sites of interest. This may be due in part to the relatively small range (<2.0 °C) in mean annual air temperatures observed at these sites (Table 4). The mountainous landscape of this region is susceptible to frequent air temperature inversions during winter [33] that may explain this lack of sensitivity. One notable exception is the statistically-significant anticorrelations between the air temperature during the period of accumulation and peak SWE recorded at Barkerville and Hedrick Lake, the two lowest elevation sites (1483 m a.s.l. and 1118 m. a.s.l., respectively). Peak snow accumulation at lower elevation snow pillow stations may thus be more sensitive to the air temperature at the onset of the seasonal snowpack when it may fluctuate near 0 °C for extended periods of time. In contrast, higher elevation stations with longer periods of accumulation and cooler air temperatures are less sensitive to the snow onset conditions. The presence of statistically-significant correlations between air temperature during the ablation period and peak SWE observed at Revolution Creek and Yanks Peak East cannot be explained by elevation alone. Other site factors such as slope and aspect, the degree of exposure and sky view factor, among others, could potentially explain the different responses seen across sites. Detailed information on each snow pillow site is unavailable to us, thus limiting our interpretation of these results.

5.2. Comparison with Other Studies

Hsieh and Tang [10] investigated the April 1 SWE in the Upper Columbia River Basin based on snow course measurements spanning 1950–1999, with mean values ranging from 120 to 1248 mm. The adjacent Upper Fraser River Basin has a narrower range of values in April 1 SWE (364–852 mm) according to the snow pillow sites that also span a more limited range of elevations (729 m a.s.l. *vs.* 1420 m a.s.l.). In accord with the results of Bohr and Aguado [35], the April 1 SWE underestimates the hydrological year peak SWE by up to 13% as this occurs from mid-April to early May at the seven snow pillow sites. This reveals the importance of recording continuous daily SWE from snow pillow sites in estimating total snowmelt contribution to river runoff in the Upper Fraser River Basin.

Mote [9] and Mote *et al.* [11] report declining mountain snowpacks across western North America, with more prominent changes at elevations <1800 m a.s.l. The decreases in SWE coincide with rising air temperatures, suggesting climate change is driving overall snowpack trends in this region. In the Upper Fraser River Basin, there are few long-term time series of SWE from snow pillow sites; however, four locations with the better temporal coverage indicate no trend in peak accumulation or air temperatures during the accumulation period from 1990 to 2012. These results may thus reflect large-scale climatic variability associated with the PDO and other large-scale sea-surface temperature and atmospheric patterns (e.g., [10,21,34]) rather than long-term trends associated with warming in the area. Mote *et al.* [36] report that warmer, drier years are often associated with El Niño events and/or the warm phase of the PDO, which tend to be associated with below-average snowpack, streamflow, flood risk, salmon survival, forest growth, and above-average forest fire risk. It is clear that multi-decadal time series of SWE in the study area are needed to establish the precise impacts of air temperature warming on snowpack accumulation and ablation characteristics in the Upper Fraser River Basin.

5.3. Study Limitations

Despite providing some initial, essential information on snowpack characteristics within the ITR, there remain limitations to this study. There are only seven active snow pillow sites covering ≈35,000 km^2 of the Upper Fraser Basin, limiting the spatial representation of the results and within those records, only four sites have records surpassing 15 years. The relatively short period of data availability limits the analysis of long-term trends in snow accumulation and ablation patterns across the study area. In addition, the snow pillow data of SWE employed in this study are subject to collection issues and measurement errors related to snow deformation, bridging and disturbance of the heat fluxes between soil and snow (e.g., [37,38]) that add uncertainty to these results. Another point of consideration is the representativeness of the snow pillow sites, which provide data only at one point in comparison to the surrounding landscape, in addition to being affected by interception, wind fields, solar shading, and thermal radiation. In the study area, snow pillows are located at relatively high elevations (1118–1847 m a.s.l.) and are generally situated in a small clearing within a forested environment. Thus, snowpack conditions at a snow pillow site may not necessarily reflect actual amounts of nearby environments and as a result, snow pillow data in BC are often used to establish an index of relative snow accumulation. These underlying issues make it unclear how representative the chosen sites in this study are of the complex topography of the region although the snow pillow sites are deployed at mid-elevation ranges and snowpack duration varies linearly with elevation here, providing a relative and accurate index of regional-scale accumulation.

Another gap in this study is the lack of reliable precipitation data for the seven snow pillow stations. This limits the interpretation of the combined factors of air temperature and precipitation on the observed SWE characteristics in the Upper Fraser Basin. Thus, a current effort is under way to verify the accuracy of several different precipitation products for possible application to the study of snowpack characteristics in the complex topography of the Upper Fraser Basin.

5.4. Possible Implications to the ITR

The extent and duration of the snowpack in the ITR is vital to the biodiversity of the ecosystem that it sustains. An endangered mountain caribou's seasonal habitat distribution is closely linked to the ITR that they utilize as a transitional corridor from summer to winter range, migrating through the different elevational zones influenced by various snowpack conditions and forage availability. Snowpack depth and supportability are critical components in determining whether a forest is accessible to caribou and considered quality arboreal lichen winter range [39]. Movement in fresh snow has higher energetic costs that may drive caribou in high-snowfall areas to lower elevations, where low-snowfall areas are located and more forage sources are more readily available [3]. Thus, when the snowpack accumulates rapidly in the early winter it allows the caribou to move to their preferred winter range more easily [40]. Once the deep snowpacks accumulate in late winter at the higher elevations, caribou migrate to the ESSF zone where they can reach the lichens found higher up the trees [3]. Changes in the onset of accumulation have the potential to degrade or delay the winter forage for the caribou herds, which may negatively impact their winter survival and potentially decrease their population sizes. Caribou survival can also be impacted by shallow snowpacks that decrease access to arboreal lichens in late winter at higher elevations, improve habitat conditions for other ungulates (e.g., moose), and increase risk of predation by wolves [3].

Water availability is an important issue in the ITR as the gradual release of water into the soil during spring melt sustains soil moisture throughout the drier summer months. Water availability will prove to be a limiting factor if spring melt occurs earlier and summer temperatures increase. Trujillo *et al.* [41] identified an elevational range between 1800 m a.s.l. and 2100 m a.s.l. of the Sierra Nevada, United States, as being most sensitive to changes in water availability. Given the elevation of the forests in the Upper Fraser Basin and the latitude at which they are situated, it can be inferred that the same sensitivity could be encountered. This has implications on the survivability of the western redcedar at these sites, as physiological adaptation to changes in precipitation and soil moisture has not been observed in the interior species [42]. In addition, the regeneration of tree seedlings in high altitude forests of western North America is mediated by the snow cover duration, which affects the length of the growing season on high-precipitation sites and the soil moisture supply on low-precipitation sites [43]. Thus, predicted snowpack decreases due to expected warming winter temperatures may impact tree regeneration in the ITR. However, Peterson [43] reports that tree growth in high-snowfall environments (under a maritime climate and near the treeline) is generally limited more by precipitation than by temperature, with growth being negatively correlated with snowpack depth. Holtmeier and Broll [44] also showed that treelines controlled mainly by orographic influences are not very susceptible to the effects of warming climates. Since the climate of the ITR is similar to that of coastal rainforests it can be inferred that future tree growth may not be greatly impacted by warming temperatures that are expected to increase winter precipitation in the form of rainfall and decrease snowpack depths.

In the coming years as global warming trends continue, climatic variation may have a negative impact on the composition and function of the old-growth forests in the ITR. In particular, snow cover changes may have direct and indirect effects on the forest dynamics of this region. A similar scenario is taking place along the north coast of BC and southern coastal Alaska, where reduced snowpack has led to root exposure damage during freeze-thaw events and is thought to be causing the widespread mortality of yellow-cedars [45]. In the Upper Fraser River Basin, winter snowpacks are predicted to decline by over 30% by the year 2080 [6]. Such a decrease in the snowpack of the ITR may decrease soil moisture and groundwater repletion, making the ancient western redcedar and hemlock stands more susceptible to fires and insect mortality in the dry summer months. These vegetation changes may also result in less carbon storage within the ITR during the 21st century [46].

5.5. Climate Change and Prospects for the Future

There have recently been declines in the duration of the Northern Hemisphere seasonal snow cover, especially during spring in western North America [47–49]. Brown and Mote [50] and Räisänen [51] further show that future changes in snow accumulation regimes will vary across the Northern Hemisphere, with the Canadian Archipelago and northern Siberia projected to experience greater snow accumulations in the future. Changing snow accumulations in BC also depend on elevation, with low elevation snowpacks being more sensitive to warming than those at high elevations [50]. In fact, some analyses and climate model projections suggest snowfall, and hence peak accumulation, may initially increase at elevations >1500 m a.s.l. of the mountainous terrain of western Canada in the next few decades [52]. The regional modeling study by Kim *et al.* [53] also reports snowfall increases in the Sierra Nevada at elevations above 2500 m a.s.l. in a doubled CO_2 climate. Global Climate Model (GCM) projections for the ITR sourced from Climate BC data [54] suggest this region will become warmer and wetter during the 21st century, with more pronounced warming in winter. Despite projected increases in wintertime precipitation, the amount of snowfall is expected ultimately to decline across the ITR, possibly leading to reduced snow accumulations. Thus, it is particularly important to continue monitoring net snowpack accumulation and ablation characteristics to better understand the long-term impacts of climate change in the ITR, further contributing to atmospheric CO_2 levels and global warming [46].

6. Conclusions

This study examined snow accumulation and ablation characteristics using SWE and air temperature data at seven snow pillow sites in the ITR of the Upper Fraser River Basin, BC. The analyses revealed peak SWE values ranging from 391.3 mm at Barkerville and 924.4 mm at Hedrick

Lake. Peak accumulation occurs from mid-April to early May, thus allowing April 1 SWE values to underestimate annual peak SWE by up to 13%. Snow cover duration lasts between 206 days at Barkerville and 258 days at Dome Mountain, with snow onset dates from mid-October to early November and snow off dates from late May to early July. The interannual variability in peak snowpack accumulation exhibits strong regional coherence, indicating that wintertime synoptic activity affects the study area similarly. Mean daily air temperatures approach −5 °C during the accumulation period and 5 °C during the ablation period. There are few statistically-significant relationships between peak SWE values and air temperature variables at the seven sites of interest. This indicates there is, at the present time, insufficient information to determine whether warmer winters lead to less snow accumulation in the ITR of the Upper Fraser River Basin. It thus remains unclear whether projected warmer air temperatures and more abundant precipitation in the 21st century will lead to more or less snow accumulation in this region. Nonetheless, the basic information on the amounts and timing of snow accumulation and ablation, although based on limited data, provide crucial information for ecohydrologists investigating the dynamics of fluctuating hydrological processes and their impacts on the ITR, an ecosystem of global importance. Future work will therefore explore 20th century snow accumulation and ablation characteristics using a macroscale land surface model and then investigate 21st century projections of peak SWE in the Upper Fraser River Basin.

The lack of a correlation between peak SWE and elevation as well as most air temperature parameters found in this study may be due to local effects on snow accumulation patterns. This suggests that complex interactions between the topography and air temperature lead to distinct, local snow accumulation and ablation patterns. While the interannual variability in peak SWE is controlled by large-scale, synoptic storm activity in the Upper Fraser River Basin, mesocale to local scale processes appear to have a dominant influence on the snowpack characteristics at the seven snow pillow sites used in the present study. Orographic enhancement of precipitation, channeling of winds in complex topography, blowing snow, and solar radiation loading may all contribute to different snowpack characteristics despite proximal locations. Thus, care must be used when interpreting snowpack evolution based solely on point data. As regional climate models attain higher spatial resolutions, their land surface schemes need to begin incorporating some of the small-scale processes that affect snowpack accumulation and ablation. To that end, an effort is currently under way to incorporate subgrid-scale snow processes in the Canadian Regional Climate Model, version 5 (e.g., [55]). This work will provide further insights on the range of factors and processes affecting snowpack characteristics in the complex topography of western Canada.

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Author contributions

Stephen J. Déry conceived the project, carried out the snow and air temperature analyses, interpreted the results and wrote the initial manuscript. Heidi K. Knudsvig, Marco A. Hernández-Henríquez and Darwyn S. Coxson provided guidance on the analyses, reviewed and commented on the initial manuscript and with S.J.D. contributed additional text to it.

Conflict of Interest

The authors declare no conflict of interest.

References

1. Barnett, T.P.; Adam, J.C.; Lettenmaier, D.P. Potential impacts of a warming climate on water availability in snow-dominated regions. *Nature* **2005**, *438*, 303–309.
2. Déry, S.J.; Sheffield, J.; Wood, E.F. Connectivity between Eurasian snow cover extent and Canadian snow water equivalent and river discharge. *J. Geophys. Res.* **2005**, doi:10.1029/2005JD006173.
3. Stevenson, S.K.; Armleder, H.M.; Arsenault, A.; Coxson, D.; Delong, S.C.; Jull, M. *British Columbia's Inland Rainforest*; UBC Press: Vancouver, BC, Canada, 2011; p. 432.
4. Ketcheson, M.V.; Braumandal, T.F.; Meidinger, D.; Utzig, G.; Demarchi, D.A.; Wikeem, B.M. Interior Cedar–Hemlock Zone. In *Ecosystems of British Columbia*. Meidinger, D., Pojar, J., Eds.; BC Ministry of Forests, Research Branch: Victoria, Canada, 1991; Special Report Series No. 6; pp. 167–181. Available online: www.for.gov.bc.ca/hfd/pubs/docs/Srs/Srs06/chap11.pdf (accessed on 29 November 2013).
5. Arsenault, A.; Goward, T. Ecological characteristics of inland rainforests. *Ecoforestry* **2000**, *15*, 20–23.
6. Coxson, D.S.; Goward, T.; Connell, D.J. Analysis of ancient western redcedar stands in the upper Fraser watershed, and scenarios for protection. *J. Ecosyst. Manag.* **2013**, *13*, 1–20.
7. Rice, R.; Bales, R.C.; Painter, T.H.; Dozier, J. Snow water equivalent along elevation gradients in the Merced and Tuolumne River basins of the Sierra Nevada. *Water Resour. Res.* **2011**, doi:10.1029/2010WR009278.
8. Meromy, L.; Molotch, N.P.; Link, T.E.; Fassnacht, S.R.; Rice, R. Subgrid variability of snow water equivalent at operational snow stations in the western USA. *Hydrol. Process.* **2013**, *27*, 2383–2400.
9. Mote, P.W. Trends in snow water equivalent in the Pacific Northwest and their climatic causes. *Geophys. Res. Lett.* **2003**, doi:10.1029/ 2003GL017258.

10. Hsieh, W.W.; Tang, B. Interannual variability of accumulated snow in the Columbia basin, British Columbia. *Water Resour. Res.* **2001**, *37*, 1753–1759.
11. Mote, P.W.; Hamlet, A.F.; Clark, M.P; Lettenmaier, D.P. Declining mountain snowpack in western North America. *Bull. Am. Meteorol. Soc.* **2005**, *86*, 39–49.
12. Harpold, A.; Brooks, P.; Rajagopal, S.; Heidbuchel, I.; Jardine, A.; Stielstra, C. Changes in snowpack accumulation and ablation in the intermountain west. *Water Resour. Res.* **2012**, doi:10.1029/2012WR011949.
13. Troendle, C.A.; King, R.M. The effect of partial and clearcutting on streamflow at Deadhorse Creek, Colorado. *J. Hydrol.* **1987**, *90*, 145–157.
14. Golding, D.L.; Swanson, R.H. Snow distribution patterns in clearings and adjacent forest. *Water Resour. Res.* **1986**, *22*, 1931–1940.
15. Boon, S. Snow accumulation following forest disturbance. *Ecohydrology* **2012**, *5*, 279–285.
16. Pugh, E.; Small, E. The impact of pine beetle infestation on snow accumulation and melt in the headwaters of the Colorado River. *Ecohydrology* **2012**, *5*, 467–477.
17. Biederman, J.A.; Brooks, P.D.; Harpold, A.A.; Gochis, D.J.; Gutmann, E.; Reed, D.E.; Pendall, E.; Ewers, B.E. Multiscale observations of snow accumulation and peak snowpack following widespread, insect-induced lodgepole pine mortality. *Ecohydrology* **2012**, doi:10.1002/eco.1342.
18. Harpold, A.A.; Biederman, J.A.; Condon, K.; Merino, M.; Korgaonkar, Y.; Nan, T.; Sloat, L.L.; Ross, M.; Brooks, P.D. Changes in snow accumulation and ablation following the Las Conchas Forest Fire, New Mexico, USA. *Ecohydrology* **2013**, doi: 10.1002/eco.1363.
19. Danard, M.; Murty, T.S. On recent climate trends in selected salmon-hatching areas of British Columbia. *J. Clim.* **1994**, *7*, 1803–1808.
20. Burford, J.E.; Déry, S.J.; Holmes, R.D. Some aspects of the hydroclimatology of the Quesnel River Basin. *Hydrol. Process.* **2009**, *23*, 1529–1536.
21. Moore, R.D.; McKendry, I.G. Spring snowpack anomaly patterns and winter climatic variability, British Columbia, Canada. *Water Resour. Res.* **1996**, *32*, 623–632.
22. Tong, J.; Déry, S.J.; Jackson, P.L.; Derksen, C. Testing snow water equivalent retrieval algorithms for passive microwave remote sensing in an alpine watershed of western Canada. *Can. J. Remote Sens.* **2010**, *36*, S74–S86.
23. Shrestha, R.R.; Schnorbus, M.A.; Werner, A.T.; Berland, A.J. Modelling spatial and temporal variability of hydrologic impacts of climate change in the Fraser River basin, British Columbia, Canada. *Hydrol. Process.* **2012**, *26*, 1840–1860.
24. Morrison, J.; Quick, M.; Foreman, M.G.G. Climate change in the Fraser River watershed: Flow and temperature projections. *J. Hydrol.* **2002**, *263*, 30–44.
25. Kerkhoven, E.; Gan, T. Differences and sensitivities in potential hydrologic impact of climate change to regional-scale Athabasca and Fraser River basins of the leeward and windward sides of the Canadian Rocky Mountains respectively. *Clim. Chang.* **2011**, *106*, 583–607.
26. Déry, S.J.; Hernández-Henríquez, M.A.; Owens, P.N.; Parkes, M.W.; Petticrew, E.L. A century of hydrological variability and trends in the Fraser River Basin. *Environ. Res. Lett.* **2012**, doi:10.1088/1748-9326/7/2/024019.

27. Picketts, I.P.; Werner, A.T.; Murdock, T.Q.; Curry, J.; Déry, S.J.; Dyer, D. Planning for climate change adaptation: lessons learned from a community-based workshop. *Environ. Sci. Policy* **2012**, *17*, 82–93.
28. Tong, J.; Déry, S.J.; Jackson, P.L. Topographic control of snow distribution in an alpine watershed of western Canada inferred from spatially-filtered MODIS snow products. *Hydrol. Earth Syst. Sci.* **2009**, *13*, 319–326.
29. Tong, J.; Déry, S.J.; Jackson, P.L. Interrelationships between MODIS/Terra remotely sensed snow cover and the hydrometeorology of the Quesnel River Basin, British Columbia, Canada. *Hydrol. Earth Syst. Sci.* **2009**, *13*, 1439–1452.
30. Eliason, E.J.; Clark, T.D.; Hague, M.J.; Hanson, L.M.; Gallagher, Z.S.; Jeffries, K.M.; Gale, M.K.; Patterson, D.A.; Hinch, S.G.; Farrell, A.P. Differences in thermal tolerance among sockeye salmon populations. *Science* **2011**, *332*, 109–112.
31. Déry, S.J.; Stahl, K.; Moore, R.D.; Whitfield, P.H.; Menounos, B.; Burford, J.E. Detection of runoff timing changes in pluvial, nival and glacial rivers of western Canada. *Water Resour. Res.* **2009**, doi:10.1029/ 2008WR006975.
32. DeWalle, D.R.; Rango, A. *Principles of Snow Hydrology*; Cambridge University Press: Cambridge, UK, 2008; p. 410.
33. Déry, S.J.; Clifton, A.; MacLeod, S.; Beedle, M.J. Blowing snow fluxes in the Cariboo Mountains of British Columbia, Canada. *Arct. Ant. Alpine Res*. **2010**, *42*, 188–197.
34. Bao, Z.; Kelly, R.; Wu, R. Variability of regional snow cover in spring over western Canada and its relationship to temperature and circulation anomalies. *Int. J. Climatol.* **2010**, *31*, 1280–1294.
35. Bohr, G.S.; Aguado, E. Use of April 1 SWE measurements as estimates of peak seasonal snowpack and total cold-season precipitation. *Water Resour. Res.* **2001**, *37*, 51–60.
36. Mote, P.W.; Parson, E.A.; Hamlet, A.F.; Keeton, W.S.; Lettenmaier, D.; Mantua, N.; Miles, E.L.; Peterson, D.W.; Slaughter, R.; Snover, A.K. Preparing for climatic change: the water, salmon, and forests of the Pacific Northwest. *Clim. Chang.* **2003**, *61*, 45–88.
37. Goodison, B.E.; Ferguson, H.L.; McKay, G.A. Measurements and Data Analysis. In *The Handbook of Snow: Principles, Processes, Management & Use*; Gray D.M., Male, D.H., Eds.; Pergamon Press: Toronto, ON, Canada, 1981; pp. 191–274.
38. Johnson, J.B.; Schaefer, G.L. The influence of thermal, hydrologic, and snow deformation mechanisms on snow water equivalent pressure sensor accuracy. *Hydrol. Process.* **2002**, *16*, 3529–3542.
39. Antifeau, T. The Significance of Snow and Arboreal Lichen in the Winter Ecology of Mountain Caribou (*Rangifer tarandus caribou*) in the North Thompson Watershed of British Columbia. Master’s Thesis, University of British Columbia, Vancouver, BC, Canada, 1987.
40. Terry, E.; Mclellan, B.; Watts, G. Winter habitat ecology of mountain caribou in relation to forest management. *J. Appl. Ecol.* **2000**, *37*, 589–602.
41. Trujillo, E.; Molotch, N.P.; Goulden, M.L.; Kelly, A.E.; Bales, R.C. Elevation-dependent influence of snow accumulation on forest greening. *Nat. Geosci.* **2012**, *5*, 705–709.
42. Fan, S.; Grossnickle, S.; Russell, J. Morphological and physiological variation in western redcedar (*Thuja plicata*) populations under contrasting soil water conditions. *Trees* **2008**, *22*, 671–683.

43. Peterson, D.L. Climate, Limiting Factors and Environmental Change in High-Altitude Forests of Western North America. In *The Impacts of Climate Variability on Forests*, Beniston, M., Innes, J.L., Eds.; Springer: Berlin/Heidelberg, Germany, 1998; Lecture Notes in Earth Sciences Volume 74, pp. 191–208.
44. Holtmeier, F.K.; Brol, G. Sensitivity and response of northern hemisphere altitudinal and polar treelines to environmental change at landscape and local scales. *Global Ecol. Biogeogr.* **2005**, *14*, 395–410.
45. Daniels, L.D.; Maertens, T.B.; Stan, A.B.; McCloskey, S.P.; Cochrane, J.D.; Gray, R.W. Direct and indirect impacts of climate change on forests: three case studies from British Columbia. *Can. J. Plant Pathol.* **2011**, *33*, 108–116.
46. Matsuzaki E.; Sanborn, P; Fredeen, A.L.; Shaw, C.H.; Hawkins, C. Carbon stocks in managed and unmanaged old-growth western redcedar and western hemlock stands of Canada's inland temperate rainforests. *Forest Ecol. Mgmt.* **2013**, *297*, 108–119.
47. Déry, S.J.; Brown, R.D. Recent Northern Hemisphere snow cover extent trends and implications for the snow-albedo feedback. *Geophys. Res. Lett.* **2007**, doi: 10.1029/2007GL031474.
48. Choi, G.; Robinson, D.A.; Kang, S. Changing Northern Hemisphere snow seasons. *J. Clim.* **2010**, *23*, 5305–5310.
49. Derksen, C.; Brown, R. Spring snow cover extent reductions in the 2008–2012 period exceeding climate model projections. *Geophys. Res. Lett.* **2012**, doi: 10.1029/2012GL053387.
50. Brown, R.; Mote, P. The response of northern hemisphere snow cover to a changing climate. *J. Clim.* **2009**, *21*, 6404–6424.
51. Räisänen, J. Warmer climate: More or less snow? *Clim. Dynam.* **2008**, *30*, 307–319.
52. Pollack, E.W.; Bush, A.B.G. Climate change in western North America caused by a CO_2 rise: A coupled atmosphere-ocean model simulation. *Atmos. Ocean* **2012**, *50*, 70–85.
53. Kim, J.; Kim, T.K.; Arritt, R.W.; Miller, N.L. Impacts of increased atmospheric CO2 on the hydroclimate of the Western United States. *J. Clim.* **2002**, *15*, 1926–1942.
54. Wang, T.; Hamman, A.; Spittlehouse, D.L.; Aitken, S.N. Development of scale-free climate data for western Canada for use in resource management. *Int. J. Climatol.* **2006**, *26*, 383–397.
55. Separovic, L.; Alexandru, A.; Laprise, R.; Martynov, A.; Sushama, L.; Winger, K.; Tete, K.; Valin, M. Present climate and climate change over North America as simulated by the fifth-generation Canadian regional climate model. *Clim. Dyn*. **2013**, *41*, 3167–3201.