INSIGHTS INTO MOUNTAIN WETLAND RESILIENCE TO CLIMATE CHANGE: AN EVALUATION OF THE HYDROLOGICAL PROCESSES CONTRIBUTING TO THE HYDRODYNAMICS OF ALPINE WETLANDS IN THE CANADIAN ROCKY MOUNTAINS

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ABSTRACT

Hydrological conditions play an important role in provisioning the exceptionally valuable ecosystem services and functions of wetlands. In alpine areas, wetland functions and services are expected to be very sensitive to climate-mediated changes in hydrology. However, few field studies of alpine wetland hydrology currently exist, thus limiting understanding of how wetlands will respond to warming and drying, and how their ecosystem services and functions will change. This study examines key processes contributing to the hydrological stability of alpine wetlands in Banff National Park, AB, Canada. During the two-year study, snowmelt timing differed by over three weeks, allowing for the examination of water table patterns under comparatively wet and dry conditions. Contrary to expectations, water table positions were relatively stable in each study year, particularly in the peat-bearing soils. Hydrophysical and hydrochemical data together provide evidence that the observed stability is in part due to groundwater contributions, which made up as much as 53% of the water budget in one peatland. Soil conditions also appear to play a role in stabilizing water table regimes. The results suggest that alpine wetlands, and peatlands in particular, may be more resilient to changes in climate than currently thought. Mineral wetlands, comparatively, may have limited adaptive capacity.

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It takes a village. The same sentiment can be expressed for obtaining a graduate degree. In this respect, I'd like to thank the many people that helped me through this process, and in particular those that helped with data collection: Hongye Wu, Dr. Daniel Karran, Carlie Elliott, Richard Helmle, May Guan, Angus Duncan, and Dr. Jonathan Conway. I'd also like to thank Dr. Masaki Hayashi and Dr. Laurence Bentley (University of Calgary) and their lab groups for their soil and geophysical expertise, and in particular Alexandra Zheltikova.

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Finally, I want to thank my mom. Without her as an example of determination, I'm not sure I would have gotten through this process. Thank you, mom, for all you have done and continue to do for me and others.

DEDICATION

For my grandfather, Otis Arden Mercer (1924 – 2015): A man committed to a life of learning, doing, and teaching.

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LIST OF ABBREVIATIONS AND SYMBOLS

α	Inverse bubbling pressure head or capillary length
A'	Wetland area
CRHO	Canadian Rockies Hydrological Observatory
CS	Campbell Scientific, Inc.
Δh_o	change in observed water table depth
ΔH_V	Vertical change in hydraulic head
δ^2H	Isotopic ratio of $^2H/^1H$ of a sample compared to a standard (VMSOW)
$\delta^{18}O$	Isotopic ratio of $^{18}O/^{16}O$ of a sample compared to a standard (VMSOW)
DJF	December, January, and February
ET	Evapotranspiration
GW_{net}	Net groundwater flow
h	Water table position relative to the soil surface
H_V	Vertical hydraulic head
HCRB	Helen Creek Research Basin
JJA	June, July, and August
K_s	Saturated hydraulic conductivity
LMWL	Local meteoric water line
L_V	Distance between vertical hydraulic head measurements
m	A fitting coefficient equivalent to 1-1/n
MAM	March, April, and May
n	Pore-size distribution index
NM	Not measured
OTT	OTT Hydromet
ϕ	Porosity
P	Precipitation
PET	Potential evapotranspiration
PT	Priestley-Taylor combination equation
q_v	Vertical specific discharge (positive) or recharge (negative)
Q_i	Inlet discharge
Q_o	Outlet discharge

RETC	Retention curve computer program
RTK GPS	Real time kinematic global positioning system
SC	Specific conductance
SfM	Structure from motion with multi-view stereo
SOM	Soil organic matter
SON	September, October, and November
SWE	Snow water equivalent
SWRC	Soil water retention curve
Sy	Specific yield
Sy_h	Specific yield at water table depth h
$Sy_{h,t}$	Specific yield at water table depth h, and time t
Sy_{soil}	Specific yield of soil
Sy_{water}	Specific yield of water
θ_R	Specific retention
t	Time step
TE	Texas Electronics
VHG	Vertical hydraulic gradient
VSMOW	Vienna Standard Mean Ocean Water
Wx	Meteorological station
Z_{xy}	Microtopographic elevation at coordinates x and y

CHAPTER 1: INTRODUCTION

1.1. Introduction

High elevation wetlands are key in the provisioning of mountain watershed functions and services – they regulate runoff (Buytaert *et al.*, 2006; Mosquera *et al.*, 2016), provide habitat for rare and endangered species (Hauer *et al.*, 2007; An *et al.*, 2013), support the stabilization of riparian areas (Grab and Deschamps, 2004), and play a role in regulating greenhouse gases (Chimner *et al.*, 2002; Millar *et al.*, 2017). In this respect, changes in the functional capacities of mountain wetlands are likely to alter the overall functions of mountain ecosystems, which are themselves very important ecosystem service providers (Viviroli *et al.*, 2003). While it is generally expected that mountain wetlands will be sensitive to changes in climate, the current understanding of these sensitivities is nascent, particularly in the alpine zone, thus limiting predictive capacity.

Essential to the provisioning of wetland functions and services is hydrologic condition, which requires a long-term water balance that promotes near-surface soil saturation during some or all of the growing season (Mitsch and Gosselink, 2007). At high latitudes, such as in the Rocky Mountains of North America, snowmelt is thought to satisfy the water demands of alpine wetlands, with limited contributions from other flow paths, due to small upslope contributing areas and poorly weathered bedrock conditions (Burkett and Kusler, 2000; Winter, 2000). This conceptual model effectively treats alpine wetlands as bogs, whose hydrologic conditions are principally controlled by the balance of precipitation (i.e., snowmelt) and evapotranspiration (Brinson, 1993; NWWG, 1997). Under such conditions the atmosphere acts as the primary driver of wetland hydrodynamics, which has led some to conclude that alpine wetlands will be extremely sensitive to even small shifts in mountain climate (Burkett and Kusler, 2000; Winter, 2000; Poff *et al.*, 2002; Körner and Ohsawa, 2005; Lee *et al.*, 2015), like those expected with climate change (IPCC, 2014). There is some evidence to support the tight control of atmospheric processes over alpine wetland hydrodynamics (hereafter referred to as the "bog model"; **Figure 1.1**). For example, research on the Tibetan Plateau has found that increases in temperature of between 0.19 and 0.23 ºC have been accompanied by large losses of wetland area – as much as 29 % in some areas (Zhang *et al.*, 2011). However, it is still unclear as to whether these changes in wetland extent have also been influenced by other, more direct, changes to hydrologic condition, such as ditching and draining associated with an increased agricultural presence in the region (Mao *et al.*, 2014).

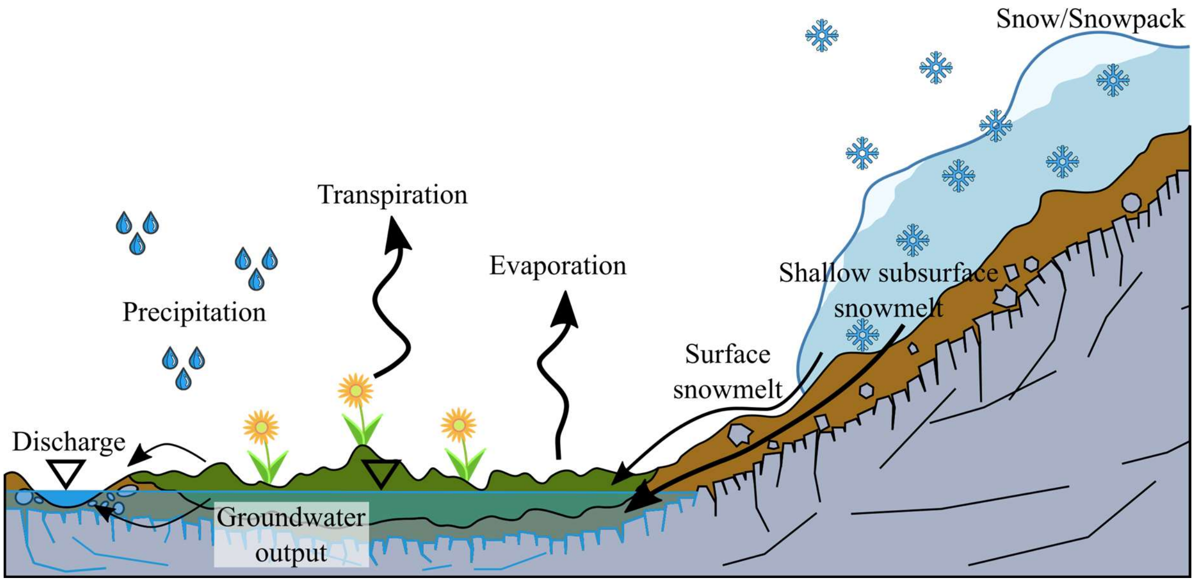

Figure 1.1. Current conceptual understanding of the most important hydrological processes to alpine wetlands. The relative size of an arrow represents its likely importance to the water balance of alpine wetlands.

Contrary to the bog model of alpine wetlands, there are a growing number of studies that illustrate the importance of groundwater in alpine watersheds (e.g., Sueker *et al.*, 2000; Manning and Caine, 2007; Hood and Hayashi, 2015). This merits interest, because subsurface contributions to alpine wetlands could promote a certain degree of temporal decoupling between wetland hydrodynamics and atmospheric conditions, with groundwater acting as a kind of hydrologic buffer (Kløve *et al.*, 2014). Under such conditions, alpine wetlands might be expected to be less vulnerable to changes in mountain climate. Evidence of groundwater contributions to alpine wetlands has largely been confined to areas outside of high latitudes, such as the tropical Andes (e.g, Caballero *et al.*, 2002; Cooper *et al.*, 2010; Gordon *et al.*, 2015). However, the presence of groundwater has been documented in at least one alpine meadow occurring at the base of a talus slope in the Canadian Rockies (McClymont *et al.*, 2010). It is still unclear as to how important groundwater storage is in the alpine landscape, and thus how common groundwater contributions are to alpine wetlands more generally.

Also unclear is if there are any landscape features that might allow generalization of the relative importance of groundwater contributions to alpine wetlands. Such relationships could assist managers in prioritizing preservation and restoration efforts in the context of changing hydrologic conditions by focusing on wetlands under more likely threat of conversion to uplands. One obvious first step towards such a generalization is an examination of the relationship between topographic context and wetland hydrodynamics. Topography has a long history of being used to infer hydrologic process, flow paths, and functions of wetlands within the watershed context, owing to its fundamental relationship to the physical drivers of water flow (Tóth, 1962; Brinson, 1993; Winter, 1999; Buttle, 2006).

1.2. Literature Review

The purpose of this literature review is to synthesize current scientific understanding and identify gaps in knowledge related to the hydrological conditions of alpine wetlands. Because there is so little information about alpine wetland hydrology, this review is largely focused on alpine hydrology in temperate environments, with an emphasis on the American Cordillera (the mountain ranges that compose the "spine" of western North and South America). Where appropriate, additional insights are garnered from the tropical mountain and lowland wetland literatures.

1.2.1. Wetlands Defined

Wetlands have been defined in a myriad of ways, but central to these definitions are the interactions between hydrology, soils, and biota (Mitsch and Gosselink, 2007). That is, wetlands are those ecosystems that are saturated for some duration of the growing season sufficient to promote biota (e.g., plants, microbes) that prefer or tolerate hydric soil conditions (NWWG, 1997). Further, hydric soils are not just those that are wet, but wet long enough to produce anaerobic conditions during some or all of the growing season (Richardson *et al.*, 2000). In this way, wetland boundaries are ultimately defined by interactions between hydrologic and edaphic conditions, as indicated and modified by vegetation and other biota (Carter, 1986; Tiner, 1993). Thus, the very definition of a wetland is dependent on their hydrologic condition both from a binary perspective (i.e., is it a wetland or not?) and functional perspective (i.e., hydrogeomorphic class).

1.2.2. Alpine Wetland Functions and Services

The functions and services of alpine wetlands in Canada are poorly documented. However, functions and services in other mountainous regions of the world have been examined to a limited extent. Alpine wetlands are known to support high rates of biodiversity and provide habitat for endemic vegetative and faunal species. In Peru, for instance, Cooper *et al.* (2010) found that 53% of species inventoried in a single wetland basin were found only in the Andes. In the Colorado Rockies, an extremely rare caddis fly larvae (*Allomyia bifosa*) has be found in alpine wetland streams, where it is thought that the wetlands provide the carbon inputs needed to support trophic dynamics important to the insect species (Hauer *et al.*, 2007). On the Qinghai-Tibetan Plateau, wetlands provide breeding habitat for rare migratory birds, such as the black-necked crane (*Grus nigricolis*) (Li et al., 2014). Further, alpine wetlands have also served as important human habitat, shaping the geographic pattern of settlement in the high Peruvian Andes some 5,000 years before present (Maldonado Fonkén, 2014).

As a result of wetland processes, alpine wetland soils serve a particularly important suite of functions in high elevation areas. For example, in Lesotho (southern Africa), wetlands soils are a key geomorphic regulator, helping to stabilize underlying mineral soils, thus preventing erosion and gully development, which would otherwise negatively impact watershed function (Grab and Deschamps, 2004). This outcome is largely the result of the presence of highly cohesive organic soils that maintain structural stability while damp, but become very fragile when dry. These organic soils also have a very high water holding capacity, with porosities as high as 81% (Crespo *et al.*, 2011) to 90% (Buytaert *et al.*, 2011), though it is thought that drought and other extreme conditions can severely reduce these capacities (Arnold *et al.*, 2014). Such high water holding capacities have been attributed to influencing regional climate. For example in Zoige (a region with the Qinghai-Tibetan Plateau), where lower temperatures and higher relative humidities are produced in watersheds with greater wetland extent (Bai *et al.*, 2013).

The soil-hydrological relationships of alpine wetlands can also impact their carbon management function, which then influences the carbon budget of the atmosphere and down-gradient waters. Depending on the position of the water table, biogeochemical status, vegetation present, and microbial activity, wetland soils can either act as a carbon sink or source, emitting both CO_2 and CH_4 (Koch *et al.*, 2007; Hao *et al.*, 2011; Franchini *et al.*, 2014; Kang *et al.*, 2014; Millar *et al.*, 2017) as well as dissolved organic carbon into streams and adjacent water bodies (Lou *et al.*, 2014). Such interactions between alpine wetland soils and hydrology are poorly understood at this point, but some data suggests that losses of peatland carbon, and thus wetland storage volumes, have disproportionately reduced discharge quantity and water quality from the Zoige region, a valuable headwater to much of Asia (Bai *et al.*, 2013). The applicability of these findings to other regions is uncertain, but such alterations to mountain systems are particularly worrisome in an uncertain climate future (Bales *et al.*, 2006).

1.2.3. Temperate Alpine Wetland Hydrology

Mountain wetland hydrology is the result of interactions between internal and external hydrologic processes occurring at multiple spatial and temporal scales (Loheide *et al.*, 2009; Lowry *et al.*, 2010). It is thus necessary to understand both basin-scale hydroclimatic processes and internal dynamics to untangle the mingled influence of hydrological processes supporting alpine wetlands (Woods *et al.*, 2006; Lowry *et al.*, 2010). To address the state of knowledge of these issues, this section (i.e., section 1.2.3) largely focuses on the mountain and alpine hydrological processes of the North American Cordillera, but occasionally draws on understanding generated from similarly situated (i.e., headwater) studies in temperate, boreal, or tropical alpine zones. Focus is generally maintained on growing season processes. However, due consideration is given to the snowmelt season, as it represents the initiation of the growing season.

1.2.3.1. Precipitation

Though patterns are variable, snow is often the dominant input to temperate alpine watersheds, constituting upwards of 95% of annual precipitation (Kattelmann and Elder 1991). However, snow distributions are uneven and influenced by a number of factors, such as topography and wind. For example, Grünewald *et al.* (2014) found that snow depth is elevation dependent, with a distinct peak followed by a decline at higher elevations due to avalanching (slope and poor adhesion) and wind redistribution. Wind redistribution, in turn, can cause snow to collect in topographic depressions, leeward slopes, and around exposed vegetation, which then has a pronounced impact on the partition of energy and hydrologic fluxes (DeBeer and Pomeroy, 2010). There is a very active research community exploring snow accumulation, redistribution and melt in the alpine of the Canadian Rockies (*c.f.*, Canadian Rockies Hydrological Observatory, http://www.usask.ca/hydrology/CRHOStns.php).

1.2.3.2. Discharge

Peak discharge in alpine areas is tightly coupled to snowmelt timing (e.g., Sueker et al. 2000; Liu et al. 2004), making snowmelt the major hydrologic event of the year (Clow *et al.*, 2003). However, the translation of snowmelt to streams is still poorly understood. Initial investigations hypothesized that overland flow was the dominant mechanism of transfer (Laudon and Slaymaker, 1997). This was predicated on the assumption that the thin soils/saprolite, short up-slope accumulation zones, and small subsurface storage volumes were limiting groundwater-surface water interactions (Winter, 2000), thus producing a reliance on intra-annual precipitation to support any flows (Clow *et al.*, 2003). However, as more observations have been made and new techniques developed, this paradigm has been challenged (Williams *et al.*, 2016). This paradigm shift is supported by a growing number of studies that have found groundwater is a significant component of spring discharge (e.g., Campbell et al. 1995; Sueker et al. 2000; Brown et al. 2007; Yang et al. 2012) and baseflow (Hood and Hayashi, 2015). For instance, Liu et al. (2004) found that discharge in the Green Lakes Valley, Colorado, USA was composed of as much as 64% groundwater, with a distinct absolute increase in groundwater during the melt season. These findings suggest a much more developed storage and subsurface flow network than previously thought.

The above circumstances may not be universal, however. For example, in the Sierra Nevada Range (California, USA) it has been observed that groundwater represents only 10-20% of stream discharge (Huth *et al.*, 2004). In contrast, Laudon and Slaymaker (1997) observed large variability in the contributions of groundwater to the stream, ranging from 25-90% in the Coast Range of British Columbia. There has thus been a push to understand this variability, as well as determine the importance of the various storage units potentially impacting runoff generation. Much of this research has since focused on talus slopes, moraines, glaciers, rock glaciers, and other cryogenic units (e.g., Caballero et al. 2002; Clow et al. 2003; Roy and Hayashi 2009; McClymont et al. 2012; Langston et al. 2013; Weekes et al. 2014); wetlands have received only limited attention (McClymont *et al.*, 2010).

While beyond the scope of this review, it should at least be noted that where glaciers exist, they are often, but not always (e.g., Brown *et al.*, 2007), a dominant source water to down gradient streams (Clow *et al.*, 2003; Penna *et al.*, 2014), effectively subsidizing streamflow during the lower summer flow periods (Viviroli *et al.*, 2003). The provisioning of this hydrologic function in the future is highly uncertain, as continued glacial loss is causing reductions in flow volumes throughout Canada and much of the world (Demuth *et al.*, 2008; Moore *et al.*, 2009). The implications of these losses to wetland water availability, and thus ecosystem stability, are largely unknown.

1.2.3.3. Groundwater

Spatial and temporal heterogeneities, as well as a poor knowledge of the geologic conditions of the alpine zone make it difficult to understand groundwater flow patterns, let alone generalize them. However, there is evidence to suggest that groundwater flow is strongly influenced by topography, geology, and recharge conditions (e.g., Ofterdinger *et al.*, 2014; Welch and Allen, 2014).

Recharge in the alpine depends on many factors such as the type and timing of precipitation, topography, soil/subsurface condition, vegetation type, and land use (Bayard *et al.*, 2005; Ofterdinger *et al.*, 2014), as well the balance of other flow paths, such as evaporation, runoff, and sublimation (Bales *et al.*, 2006). Snowmelt is the main recharge source to alpine aquifers (Bales *et al.*, 2006; Hood and Hayashi, 2015). However, the translation of snowmelt to recharge can be quite variable due to a number of factors. For example, topographic patterns can promote snow accumulation in depressions and other areas. Snowpack depth, in turn, is an important regulator of soil frost. Bayard *et al.* (2005), for example, observed that during a heavy snow year seasonal frost was limited, due to the insulating properties of snow, while in a low snow year the frost was deep and lasted well into the snowmelt season. These contrasting conditions then promoted 90-100% infiltration efficiency during the low frost year and only 65-75% efficiency during the deep frost year. However, design limitations of their study did not examine the role of local topographic depressions, which may act to concentrate recharge and alter the basin-scale recharge efficiencies. Aspect and slope angle are also important controls of recharge as the energy balance between different slopes will produce differences in evapotranspiration and melt timing, both of which are impacted by vegetation, subsurface architecture, soil moisture, and relative humidity (Drexler *et al.*, 2004; Carey and Quinton, 2005).

Hydraulic conductivity can be viewed as a synoptic variable, summarizing a number of geologic properties, such as porosity and fracture geometry (i.e., aperture size, density, and connectivity) (Welch and Allen, 2014). That said, hydraulic conductivities in the alpine are extremely variable, influenced by the scale of observation, geologic structures (including permafrost and preferential flow paths), and lithologic discontinuities (Kohl *et al.*, 1997; Stein *et al.*, 2004; Ofterdinger *et al.*, 2014).

Current understanding of patterns of hydraulic conductivity in the alpine are summarized by the critical zone concept (*sensu*, Welch and Allen, 2014), which emphasizes a five layered model of the subsurface. These layers include: soil, saprolite, fractured and weathered bedrock, fractured bedrock, and unweathered/competent bedrock (Anderson *et al.*, 2012; Welch and Allen, 2014). The first three layers are thought to be very variable in their bulk hydraulic conductivity values. Studies have documented a lower range of between 10^{-5} to 10^{-7} m/s with no apparent trend in anisotropy (Harr, 1977; Katsuyama *et al.*, 2005; Banks *et al.*, 2009; James *et al.*, 2010; Welch and Allen, 2014). The fractured bedrock layer has a documented bulk hydraulic conductivity that ranges from 10^{-6} to 10^{-8} m/s with a maximum documented depth of approximately 200 m (Welch and Allen, 2014). Last, the unweathered bedrock layer can extend to >2 km below the surface with hydraulic conductivities ranging from 10^{-6} to 10^{-9} m/s, the lower values being more representative (Welch and Allen, 2014). In contrast, saturated hydraulic conductivities in at least one alpine meadow (wetland) have been found to range from 10^{-5} to 10^{-7} m/s (McClymont *et al.*, 2010).

1.3. Research Gap

Alpine wetlands provide many important ecosystem functions and services, many of which are bounded by the water table. The hydrologic processes important to maintaining water tables in these wetlands are still poorly understood. Thus, it is unclear how these ecosystems will respond to climate-mediated changes in water availability. Because direct recharge from snowmelt is thought to be the dominant hydrological input to many alpine wetlands, the current paradigm suggests these ecosystems will be very sensitive to directional shifts in climate that promote warming and drying. However, mounting evidence suggests groundwater may also be an important source water to streamflow in some alpine watersheds. The conditions that promote groundwater importance, such as topography, remain unclear. There is thus a need to better understand the hydrological processes influencing water table dynamics of alpine wetlands and their geomorphic controls so we may better predict how they will respond to future climate conditions.

1.4. Purpose and Objectives

The purpose of this thesis is to explore the hydrologic conditions of alpine wetlands so as to begin to untangle the various factors influencing their vulnerability to climate change. Within this context, the water table is used as both an indicator of moisture status and as a means of understanding the hydrologic processes being integrated in wetlands. The objectives of this study are to: 1) characterize water table dynamics of alpine wetlands; 2) identify which key hydrologic processes influence water table position; and 3) evaluate whether topographic setting influences the hydrological processes critical to water table maintenance. To meet these objectives, three alpine wetlands situated in contrasting topographic positions were studied in the Canadian Rocky Mountains. These wetlands were examined over two growing seasons that differed by a month in the timing of snowmelt.

CHAPTER 2: METHODS

2.1. Study Site

This study is part of the Canadian Rockies Hydrological Observatory (CRHO; http://www.usask.ca/hydrology/CRHO.php), which focuses on better understanding and predicting changes in the processes influencing both the hydrologic flows from, and resilience of, mountain ecosystems. Our study focuses on the hydrologic conditions of wetlands in the Helen Creek Research Basin (HCRB) located in Banff National Park, AB, Canada (**Figure 2.1**). HCRB is a 315 ha, currently un-glaciated, U-shaped valley typical of the southern Canadian Rocky Mountains. Helen Creek is a tributary of the Bow River, which in turn flows to the Saskatchewan River – one of Canada's most economically important watersheds. Elevations of HCRB range from ~2,200 to ~2,900 m, with tree line occurring at ~2,350 m.

No long-term climate data is available for HCRB. However, regional weather station data are available for the area (**Figure 2.2**). Bow Summit is approximately 3.3 km (straight line distance) to the northwest of HCRB and located at an elevation of 2,080 m, while Skoki Mountain is approximately 31.2 km to the southeast at an elevation of 2,040 m. The earliest data available for Bow Summit is March 2006, while Skoki Mountain is March 2009. Similar weather observations were made at both sites. The average winter (DJF) air temperature at Bow Summit was -10.5 C, while the summer (JJA) temperature was 8.7 C. Fall (SON) and spring (MAM) air temperatures were -1.0 and -2.4 C, respectively. Daily average precipitation rates across seasons were fairly constant, ranging from a low of 1.7 mm/day in the winter to a high of 2.6 mm/day in the fall.

HCRB is largely underlain by the Miette Group, which consists mostly of sandstone, conglomerate, siltstone, and slate (Pana and Elgr, 2013). Above tree line, the vegetation is dominated by shrub tundra, which transitions to bare rock at about 2,500 m. The lateral edges of the watershed are also bare, with a lateral moraine on much of the west side. Wetlands and lakes

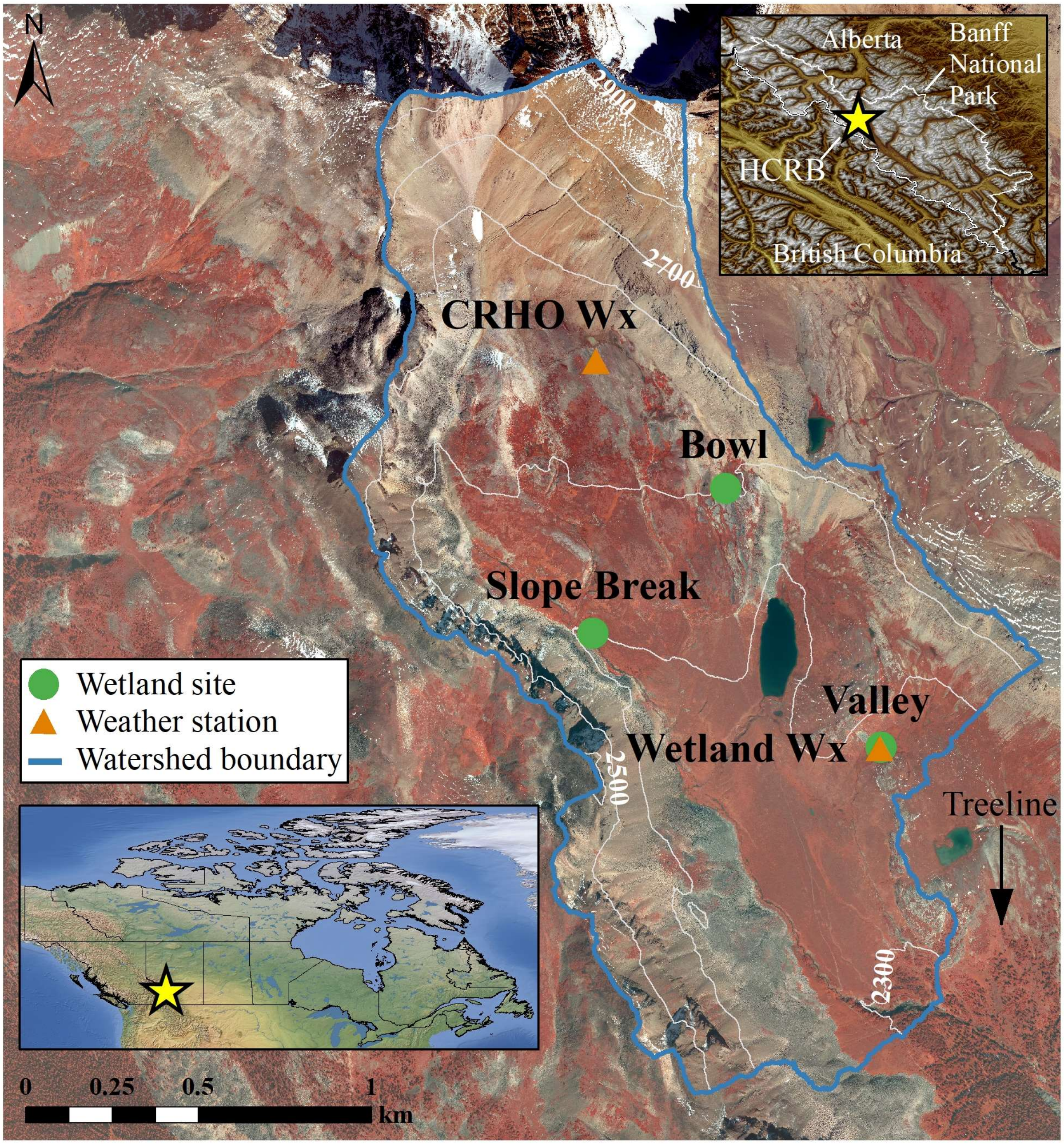

Figure 2.1. A color infrared aerial photo of the Helen Creek Research Basin (HCRB) with its continental (lower left) and regional (upper right) contexts. Orange triangles indicate the locations of weather stations: the Canadian Rockies Hydrological Observatory weather station (CRHO Wx), located in the upland, and the wetland weather station (Wetland Wx), located in a wetland. Grey lines represent 100 m elevation contours.

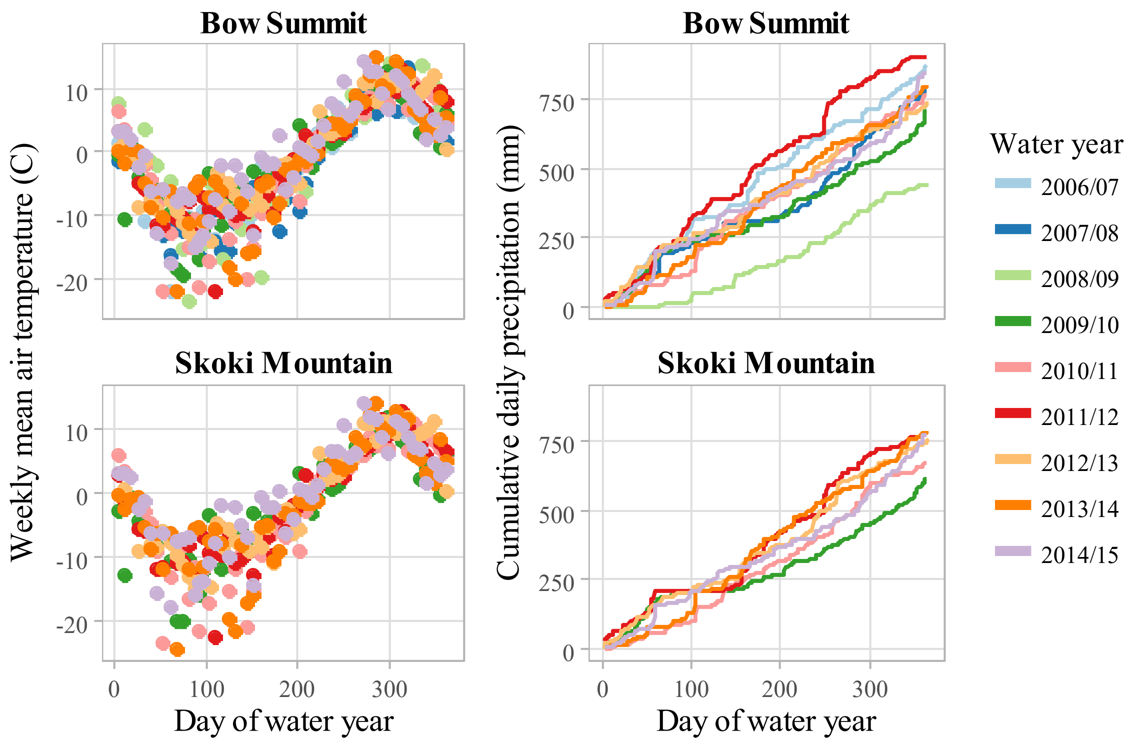

Figure 2.2. Climate data for two weather stations proximal to HCRB, based on water year (starting on Oct 1st of each calendar year). Both weather stations are approximately 300-900 meters lower than HCRB – thus the data are likely warmer and drier than that experienced by HCRB. Data available at: https://agriculture.alberta.ca/acis/alberta-weather-data-viewer.jsp.

cover approximately 5.3 and 1 % of the watershed, respectively. Common megafauna in the basin include ground squirrel, hoary marmot, and grizzly bear. Three wetlands in the basin (8.8% of the total wetland area) were selected for study based on differing profile (down slope) and planform (across slope) topographic geometries – Bowl, Slope Break, and Valley.

Bowl is the highest of the studied wetlands (2,490 m) and has profile and planform geometries that are concave (**Figure 2.3A**). Important hydrologic characteristics include surface inflows and outflows and a pond. Inflows occur as diffuse overland flow that disappear following the end of the snowmelt season. Perennial and well-defined outflow originates from a pond near the base of the wetland. Some seasonal frost was found in the spring of 2014, but quickly melted with the onset of summer. Plant cover is characterized by low sedges (*Carex* spp.), ericaceous shrubs (e.g., *Empetrum nigrum*) and mosses. The microtopography of the site is generally hummocky with some flat areas. Soils are a mix of gravels, silt, and organic matter. The area of Bowl is 0.15 ha and the area of its watershed (upslope accumulation area) is 0.99 ha, as determined by topographic boundaries.

The wetland Slope Break occurs at 2,385 m, is adjacent to Helen Creek and below a sharp break in slope such that its profile and planform geometries are concave and linear, respectively (**Figure 2.3B**). A small spring abuts the wetland, but is effectively parallel to the surface slope. No seasonal frost was detected at the site. Slope Break is moderately hummocky and has only thin organic soils. Vegetation cover is similar to Bowl's. It is the smallest wetland of the three, with an area of 0.11 ha and a watershed area of 0.54 ha.

Valley is located at 2,365 m, and is linear and concave in its profile and planform geometries, respectively (**Figure 2.3C**). The wetland has both a perennial inlet and outlet. Additionally, a number of perennial and intermittent seeps dot the margins of the wetland. The wetland's microtopography is characterized by low hummocks with occasional pools, flarks, and strings, the latter two of which are most evident during the high-water conditions of the spring snowmelt season (Mercer and Westbrook, 2016). Radiocarbon dating of the basal peat in this wetland indicate that organic matter started to form ~3,800 years before present (3820 ± 30 BP, Beta Analytic sample number 403430). Winter soil temperature records suggest the soil at this site does not freeze during the winter, and no frost was found at the site in the spring of 2014. The vegetation is characterized by low willows (*Salix* spp.), sedges, cotton grass (*Eriophorum* spp.), and moss. Valley is the largest wetland studied with an area of 1.23 ha and a watershed area of 21.0 ha.

2.2. Water Table Measurements

Water table dynamics were recorded in the field at monitoring wells near the geographic center of each study wetland. Water levels were measured using absolute pressure transducers every 15-minutes during the summers of 2014 and 2015 (Solinst Levelogger Edge 3001 and Onset HOBO 0-4m). A barometric pressure transducer (Solinst Barologger) was used to correct water level observations for atmospheric pressure fluctuations. The barometric transducer was installed in a well, always above the water table, at Valley to shade it and prevent any thermal artifacts from influencing the resultant water level measurements (McLaughlin and Cohen, 2011). Manual depth measurements were occasionally recorded as per Westbrook *et al.* (2006) to assure electronic data quality. Any missing water level data were linearly interpolated from data available immediately before and after the missing data (< 1% of data at Valley and Slope Break).

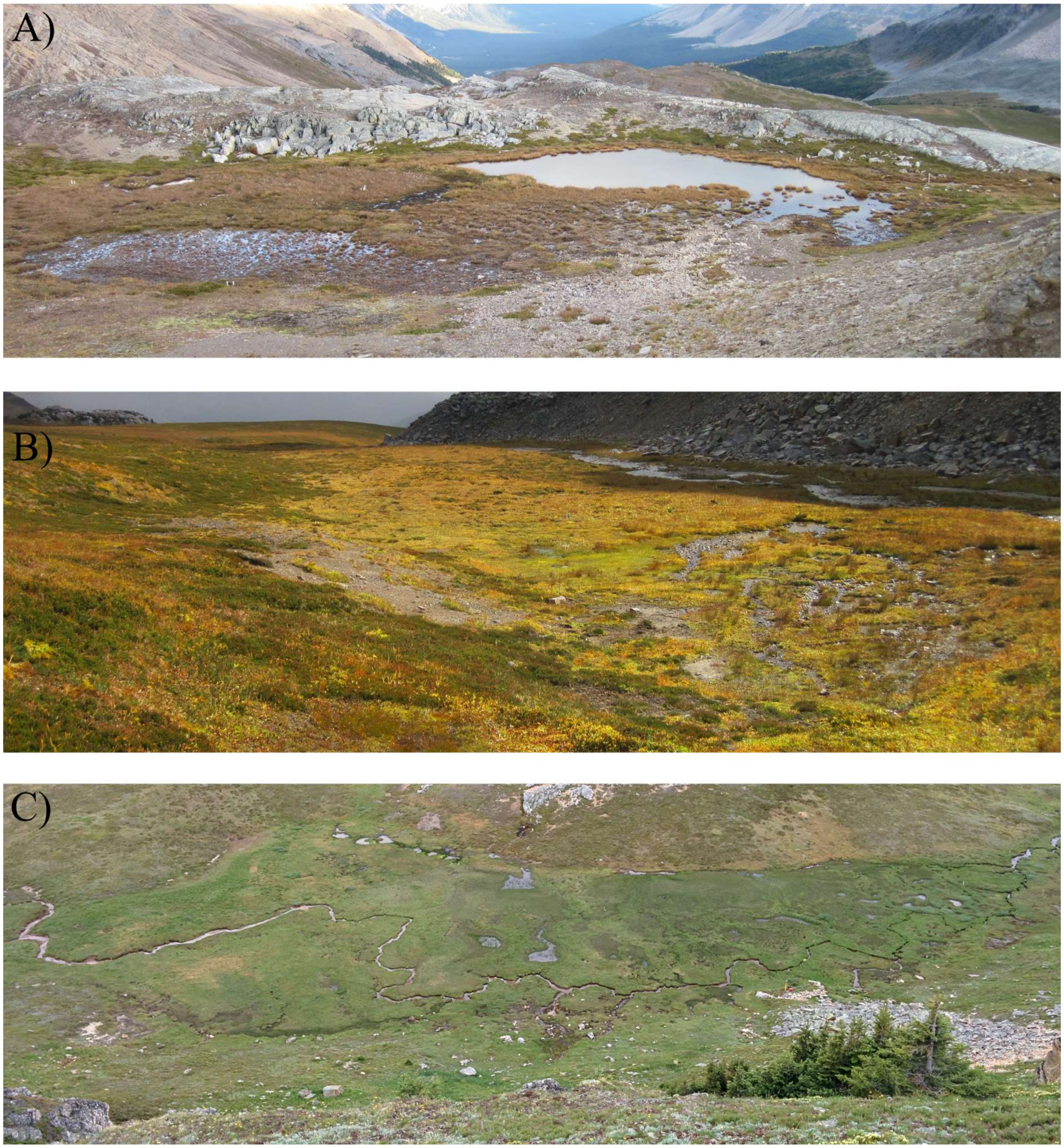

Figure 2.3. The wetlands, as viewed from their northern edges, are: **A**) Bowl, **B**) Slope Break, and **C**) Valley.

2.3. Hydrophysical Processes

2.3.1. Snowmelt Timing

A meteorological station was installed in an upland area of the HCRB (CRHO Wx) at 2,545 m during the fall of 2013, and operated throughout the study. Because this study was prompted by an interested in wetland hydrologic processes, an additional weather station was also established at the Valley wetland (Wetland Wx), which operated July 7th to September 19th, 2014. Similar observations were collected at both stations (**Table 2.1**) allowing gap-filling of any missing data at the wetlands, as well as estimations of uncertainties associated with spatially heterogeneous processes (e.g., rain). Snowmelt timing was determined using a SR50 (Campbell Scientific) located at CRHO Wx. The relationship between snow depth and snow water equivalent (SWE) was established using snow survey data from 2014 and 2015 ($R^2 = 0.98$, $p << 0.01$; **Appendix A**).

Table 2.1. The variables measured at each meteorological station, relevant to this study, including the make and model of instruments used for data collection. All data were collected in 15-minute intervals, with the exception of the weighing gauge, which was collected at variable time steps ranging from 5 to 15 minutes.

		Instrument make and model	
Variable	Units	Wetland Wx	CRHO Wx
Relative permittivity	[-]	CS-CS650	CS-CS650
Soil electrical conductivity	dS/m	CS-CS650	CS-CS650
Soil temperature @ 0.1 m	C	CS-CS650	CS-CS650
Air temperature @ 1.6 m	C	CS-HMP45C212	CS-HMP45C212
Relative humidity @ 1.6 m	%	CS-HMP45C212	CS-HMP45C212
Rain - Tipping bucket	mm	TE-525M	NM
Rain/snow - Weighing gauge	mm	NM	OTT-Pluvio
Snow depth	mm	NM	CS-SR50
Soil heat flux @ 0.1 m	W/m^2	CS-HFP01	CS-HFP01
Incoming shortwave radiation	W/m^2	CS-CNR1	CS-CNR4
Incoming longwave radiation	W/m^2	CS-CNR1	CS-CNR4
Outgoing shortwave radiation	W/m^2	CS-CNR1	CS-CNR4
Outgoing longwave radiation	W/m^2	CS-CNR1	CS-CNR4

CS: Campbell Scientific, Inc.; OTT: OTT Hydromet; TE: Texas Electronics; NM: not measured.

2.4. Water Table Modelling

To investigate the applicability of the bog model in the studied wetlands, a 1D water table model was applied that considered only surface and atmospheric volumetric fluxes: precipitation, evapotranspiration, and discharge. It was reasoned that if surface fluxes alone, accounting for error and process uncertainties, can explain the water table dynamics then groundwater would be expected to play a limited role in satisfying the moisture conditions of the study wetlands. The basic form of the water table model used is (modified from Sumner, 2007):

$$h_{t+1} = h_t + \frac{P_t - ET_t + Q_{i,t} - Q_{o,t}}{Sy_{h,t}} \tag{2.1}$$

where t and t+1 indicate some initial and later time step, respectively, h is the water table position relative to the soil surface [L] (negative below the soil surface), P is precipitation [L], ET is evapotranspiration [L], Q_i and Q_o [L] are discharge at the inlet and outlet (if applicable), respectively, and $Sy_{h,t}$ [-] is specific yield at water table h and time t. In this representation, all volumetric fluxes are areally weighted, relative to the surface area of a given wetland, resulting in a water depth [L]. Specific yield represents the relationship between change in combined fluxes (i.e., storage) and change in water table, with values ranging from between close to 0 and 1. When values are close to 0, small absolute changes in hydrologic flux will yield large water table changes, while values of 1 produce a 1:1 change in storage and water table position, which will occur in open water.

The relationship defined by **Equation 2.1** formed the basis for treatment of the water table with two different hydrological models. The first treatment only accounted for direct water fluxes to the study wetlands, meaning lateral flow from upslope areas was excluded. The second treatment included lateral flows from upslope areas, weighted by the total area of an individual wetland's contributing watershed (e.g., 21.0 ha in the case of Valley). This weighting factor applied only to precipitation and ET losses from upslope (upland) regions and not discharge. Thus, it can be thought to over-estimate the amount of lateral inputs to wetlands, since there may be some double accounting of waters in streams entering the wetlands (if applicable). Lateral flows from upslope areas were assumed to be instantaneous (i.e., conductivity is infinite or resistance is 0), which should be a reasonable approximation, given the small upslope watersheds of each wetland, the steep terrain, and the porous nature of the soil and regolith in the basin.

2.4.2. Precipitation, Evapotranspiration, and Discharge

Direct precipitation was measured at Wetland Wx using a tipping bucket rain gauge during the summer of 2014 (**Table 2.1**). Precipitation was also measured at the CRHO Wx using a weighing gauge for the entire duration of the study. Snowfall was observed every month of the year in HCRB, though in much smaller quantities and of limited persistence during the summer.

Precipitation phase was corrected for using the method of Harder and Pomeroy (2013). Corrections for catch efficiency for the weighing gauge were calculated for the snow portion of precipitation (Macdonald and Pomeroy, 2007).

Rates of evapotranspiration were estimated using the Priestley-Taylor combination (PT) equation (Priestley and Taylor, 1972). Though PT is an estimate of potential evapotranspiration (PET), the method has been found to provide a reasonable estimate of ET in well-watered conditions (Drexler *et al.*, 2004), and generally outperforms more complex ET formulations, such as the Penman-Monteith equation, in wetland environments (Gavin and Agnew, 2004; Rosenberry *et al.*, 2004). Given that the water tables remained near the ground surface in the study wetlands throughout the study, ET and PET should have a ratio close to 1:1 (Shah *et al.*, 2007). Energy balance terms for the wetlands were measured at Wetland Wx, while upland energy balance terms were measured at CRHO Wx. A standard scaling value of 1.26 was applied to the wetlands (Rosenberry *et al.*, 2004; Hood *et al.*, 2006), while a lower value of 1 was applied to the uplands (Pape *et al.*, 2009; Muir *et al.*, 2011; Hood and Hayashi, 2015). An empirical relationship was derived between wetland and upland PET, which was used to gap-fill data for wetland estimates of ET when the Wetland Wx was not operational (< 10% of data in 2014 and all of 2015) ($R^2 = 0.77$; $p < 0.05$; **Appendix B**).

Discharge was estimated from stream stage using rating curves. Stream stage was recorded with absolute pressure transducers (Solinst Levelogger Edge 3001) and corrected for barometric pressure (Solinst Barologger). Rating curves were developed via the area-velocity method (Dingman, 2008) using a current meter (Marsh-McBirney Flo-Mate 2000) (**Appendix C**). The rating curves had an R^2 of 0.76 ($p = 0.13$), 0.87 ($p = 0.06$) and 0.92 ($p = 0.04$) at the outlet of Bowl, and inlet and outlet of Valley, respectively. Although there were occasionally small volumes of diffuse flow into the wetlands, it is assumed those volumes were captured in the large uncertainties of the above relationships. Stream stage, and thus surface flows, were only measured during the growing season of 2014, for Bowl (outlet) and Valley (inlet and outlet), with an $n = 3$ for each site.

2.4.3. Specific Yield

Specific yield was estimated using a composite function that incorporates the (van Genuchten, 1980) soil water retention curve (SWRC) relationship (Cheng *et al.*, 2015). Parameters of the Sy function are porosity [L^3 L^{-3}], ϕ, specific retention [L^3 L^{-3}], θ_R, inverse bubbling pressure head (aka, capillary length when inverted) [L^{-1}], α, average absolute depth of the water table, pore-size distribution index [unitless], n, and m which is equivalent to 1-1/n. The function is represented as (Sumner, 2007; Cheng *et al.*, 2015):

$$Sy(Z_{xy}, h) = \begin{cases} Sy_{water} = 1 & h > Z_{xy} \\ Sy_{soil} = (\phi - \theta_R)(1 - [1 + (\alpha h)^n]^m) & h < Z_{xy} \end{cases} \quad (2.2)$$

where Z_{xy} is the small-scale elevation (relative to the zero datum at which the water table is being measured) at the x- and y-coordinates of a point in the wetlands (represented by a cumulative distribution function), Sy_{water} is the specific yield when the water table is above the soil surface, and Sy_{soil} is the specific yield of soil when the water table is below the soil surface. The ultimate Sy value for a given water table depth, Sy_h, accounts for wetland microtopography via (Sumner, 2007; Dettmann and Bechtold, 2016):

$$Sy_h = \frac{\sum Sy(Z_{xy}, h)}{A'} \quad (2.3)$$

where all Sy values are summed and averaged over a wetland's area [L^2], A'.

Equation 2.3 assumes that soils are at or near hydrostatic equilibrium (Nachabe, 2002; Acharya *et al.*, 2012; Cheng *et al.*, 2015), such that the hydraulic head is invariant with depth (i.e., hydraulic potentials are roughly equal down the profile). To determine a time-scale at which this condition could be approximated, the time-to-drain function developed by Nachabe (2002) was used. The function incorporates observed changes in water table height, soil water retention parameters, and saturated hydraulic conductivity [$L\ T^{-1}$], K_s. That analysis indicated that a daily time-step would be sufficient in approximating hydrostatic equilibrium in the wetlands studied.

To assess the soil hydraulic properties of the study wetlands, soils were randomly sampled from Bowl (n = 4), Slope Break (n = 5), and Valley (n = 12). Ideally, more samples would have been collected, but it is yet unclear as to how sensitive alpine wetlands are to soil disturbance in the Canadian Rockies, thus an attempt was made to try to find a balance between limiting disturbance and characterizing soil hydraulic parameters. Soils samples were collected in PVC pipe with a diameter of 5.2 cm and a height of 5 cm. At Bowl, all soil samples were randomly sampled from a depth of 15 cm (n = 4). At Slope Break, samples were also taken at 15 cm depth, with one of the samples being taken at 8 cm (the shallow depth was due to the rockiness of the soil; n = 4). At Valley, six samples were taken at 15 cm depth, two were taken at 35 and 40 cm, and the remainder were sampled at 50 cm (n = 4). Again, a more systematic approach to depth sampling would have been better, but a reduction in disturbance was necessary, considering the study took place in a national park of international importance.

The resulting soil samples were split in half (producing soil samples ~ 2.5 cm in height) for different purposes: 1) to develop soil water retention curves, and 2) to estimate soil organic matter content. Relationships between pressure and water content were determined via pressure plate extractor (Klute, 1986) at the University of Calgary. Bulk density [M L^{-3}] was also determined as part of that process. Soil hydraulic parameters were fit using RETC (van Genuchten *et al.*, 1991). Replicate soil samples were analyzed for soil organic matter (SOM) content via loss on ignition (Rydin and Jeglum, 2006). *In situ* peat depths were also assessed using a combination of soil auger, push probe, and soil pits.

Saturated hydraulic conductivity was determined via slug test (Hvorslev, 1951; Fetter, 2001) at piezometers in Bowl (n = 10), Slope Break (n = 12), and Valley (n = 14). Piezometer nests contained both a 'shallow' and 'deep' piezometer. 'Shallow' piezometers were generally within 20 cm of the soil surface, while 'deep' piezometers were located approximately 15 cm below a noticeable change in soil properties (i.e., texture, organic content), to provide a sense of exchange between wetland layers both within and near the edges of wetlands. Piezometers were made of PVC with an internal diameter of 3.4 cm. The base of each piezometer was slotted for approximately 10 cm, and thus the middle of the slot was 10 cm below any noticeable change in soil properties. Piezometers were developed after installation. The maximum piezometer depth was ~100 cm, located at Valley.

Microtopographic elevation at each wetland was determined using structure from motion with multi-view stereo (SfM) via the methods and recommended processing steps developed by Mercer and Westbrook (2016), who used Valley as a test case. Two additional processing steps were added to the methods of Mercer and Westbrook (2016) to better suite conditions at Bowl and the Sy formulation. First, a pond occurs at Bowl, which would cause an upward bias in the estimated soil elevation from ground-based SfM techniques owing to reflection from the water (points would be interpreted as the soil surface, rather than open water). To account for this bias, a high precision (±3 cm) global positioning system (Leica Geosystems Viva GS 15 & CS 15

RTK GPS) survey was conducted to map the bathymetry of the pond. Pixels representing the pond were then replaced in the SfM-MVS generated digital elevation model by GPS survey points using linear interpolation. Second, because water tables tend to follow mesotopographic trends (Winter, 1999; Haitjema and Mitchell-Bruker, 2005; Van der Ploeg *et al.*, 2012) microtopographic data were detrended. This had the added benefit of preventing the model from artificially inundating or drying areas upslope or downslope of the monitoring well, respectively. All photo processing and point cloud generation was performed using PhotoScan Professional 1.2.5 (Agisoft, St Petersburg, Russia), while all other geospatial tasks were performed in ArcGIS 10.3 (ESRI, Redlands, California).

2.4.4. Water Table Modeling Uncertainty

To better ensure that magnitude of water table monitoring results were reflective of hydrologic processes and not error, sensitivity analysis was performed. For this, a Monte Carlo approach was used, similar to other hydrological uncertainty methods (Beven and Binley, 1992, 2014). The Monte Carlo approach involved adding a suite of random values from empirical error distribution functions to the appropriate water budget component and re-running a model 10,000 times.

For snow error, guidance from Harder and Pomeroy (2013) was followed: a uniform 10% error in the snow component was added to all summer snow precipitation events. The empirical error distribution for PET was determined by resampling from the regression residuals relating PET from the wetland and upland weather station data. The same error distribution was used for the upland PET estimates as there was no independent error estimate for those data. In the case of surface water, error was assumed to be normally distributed based on the residuals defined by the stage-discharge relationships. Similar procedures were utilized for soil parameter uncertainties in estimating Sy. Unless otherwise indicated, all data analysis and model implementations were performed in R 3.4.3 (R Core Team, 2017).

2.4.5. Water Budget

To better understand the potential role of groundwater in controlling water table dynamics, **Equation 2.1** was rearranged to solve for net groundwater, GW_{net}:

$$GW_{net,t} = \Delta h_{o,t} Sy_{h,t} - P_t + PET_t - Q_{i,t} + Q_{o,t} \tag{2.4}$$

where Δh_o is the observed change in water table depth. In this application, the term $\Delta h_o Sy_h$ is effectively the change in storage commensurate with an observed change in water table position. Large errors are often associated with net groundwater formulations similar to **Equation 2.4**. To ensure that error could not explain the estimation of GW_{net}, a similar Monte Carlo approach to that described above was used, again incorporating hydrologic process and soil hydraulic parameter uncertainties. A total of 10,000 model runs were performed for all wetlands in 2014. Because no surface water measurements were made at Bowl or Valley in 2015, only Slope was evaluated that year.

2.4.6. Groundwater Flux

Darcy's Law provides a means of directly estimating vertical specific discharge or recharge of groundwater [L T^{-1}], q_v from observational data (Hunt *et al.*, 1996):

$$q_v = K_s \frac{\Delta H_V}{L_V} \tag{2.5}$$

where ΔH_V is the vertical change in hydraulic head [L], and L_V is the distance from head measurements [L], such that $\Delta H_V/L_V$ is the vertical hydraulic gradient [unitless], VHG. In this case, K_s was estimated using the harmonic bulk saturated hydraulic conductivity between the shallow and deep piezometer at each nest. Though isotropic conditions were assumed, it is just as likely that horizontal and vertical hydraulic conductivities are 1-2 orders of magnitude different. Thus, q_v values were treated as relative indicators of groundwater flux. VHG data were calculated from measurements at each piezometer nest. Weather and logistic difficulties (i.e., HCRB was closed due to bear activity for weeks at a time during the study) prevented collection of enough data to meaningfully quantify groundwater fluxes. However, there was sufficient groundwater flux data to provide a reasonable quantitative context for comparison to the water budget findings.

2.5. Hydrochemical Estimation of Hydrologic Processes

Hydrochemical data were used as an independent tool to evaluate source waters of the wetlands. Hydrochemical measurements included specific conductance, (SC: µS/cm normalized to 25 °C), and stable water isotopes of various source waters both in the wetlands and around HCRB during the summer of 2014 and 2015, near the end of the snowmelt season. Water samples were collected from lakes, ponds, streams, springs/seeps, and wetland wells. Rain samples were opportunistically collected over the study period. Snowpack samples were collected during the melt season. SC, which can indicate if waters have had significant contact with subsurface geologic material (Laudon and Slaymaker, 1997; Klaus and McDonnell, 2013; Mueller *et al.*, 2016), was measured using a YSI Professional Plus Multiparameter meter. Isotope samples, which can be useful in distinguishing between snow and rain water sources (Kendall and McDonnell, 1998; Mueller *et al.*, 2016), were collected at the same time as SC samples, and placed in 20 mL scintillation vials with zero headspace. Further, the tops of isotope sample vials were wrapped in plastic paraffin film and stored in a refrigerator until being analyzed for both $\delta^{18}O$ and $\delta^{2}H$ at the Watershed Hydrology Lab, University of Saskatchewan, using cavity ring down spectroscopy (Los Gatos Research). Results of that analysis are reported relative to VSMOW (Vienna Standard Mean Ocean Water).

Water sources were classified in the field to help determine if hydrochemical signatures were significantly different from one another. Classes included: rain, snow, groundwater (e.g., springs, seeps, wells) and surface water (e.g., lakes, streams, ponds). Differences between classes were individually evaluated for each hydrochemical constituent via Tukey's 'Honest Significant Difference' method (Yandell, 1997). To determine the proportion of a given water source, an end-member mixing model was used (Hooper *et al.*, 1990; Liu *et al.*, 2004).

CHAPTER 3: RESULTS

3.1. Snowmelt Timing and Mass Balance

Maximum SWE values measured at CRHO Wx were similar in 2013-14 (520 mm) and 2014-15 (513 mm). However, snowmelt occurred several weeks earlier in the spring of 2015 than in 2014 (**Figure 3.1**). In effect, the basin was snow free by June 17 in 2015, 22 days earlier than in 2014. Assuming maximum SWE at CRHO Wx was roughly representative across HCRB, snowpack volumes for each wetland (SWE multiplied by wetland watershed area) were roughly the same between years (**Table 3.1**). Similarly, if those volumes were translated into water depth (i.e., snowpack volume divided by wetland area), they were also relatively consistent across years. Based in nearby studies, it is expected that only 10-20 % of max SWE was translated to the subsurface hydrologic system (Hood and Hayashi, 2015). Recharge estimates in alpine areas are variable from year to year – depending on snowpack and subsurface conditions (Bayard *et al.*, 2005), so these values are just very rough estimates. Assuming 20% of max SWE recharged the subsurface, for example, recharge volumes would be equivalent to 1.04×10^3, 0.56×10^3, and 21.84×10^3 m^3 of snowmelt, which could have then been received by Bowl, Slope Break, and Valley, respectively, in the year 2014.

3.2. Observed Water Table Dynamics

Water table dynamics for each wetland are illustrated in **Figure 3.2**. None of the wetlands expressed a strong water table decline over the growing seasons. The maximum water table position was 3 cm above the soil surface and occurred at Valley in 2014 while the lowest water table position was 35 cm below the soil surface at Slope Break in 2015. Mean seasonal water table depths at Bowl and Valley ranged from 5.7 to 2.6 cm below the soil surface, respectively (**Table 3.2**). Bowl had a slightly lower mean water table in 2015 compared to 2014, but Valley was virtually unchanged between years. Slope Break, a mineral wetland, had a substantially deeper and more variable water table than the other two wetlands as evidenced by 25% less time spent above -20 cm from the soil surface. All the wetlands expressed some diel water table fluctuations, with nighttime increases in water table position, which are usually attributed to lateral or groundwater recharge (Loheide *et al.*, 2005).

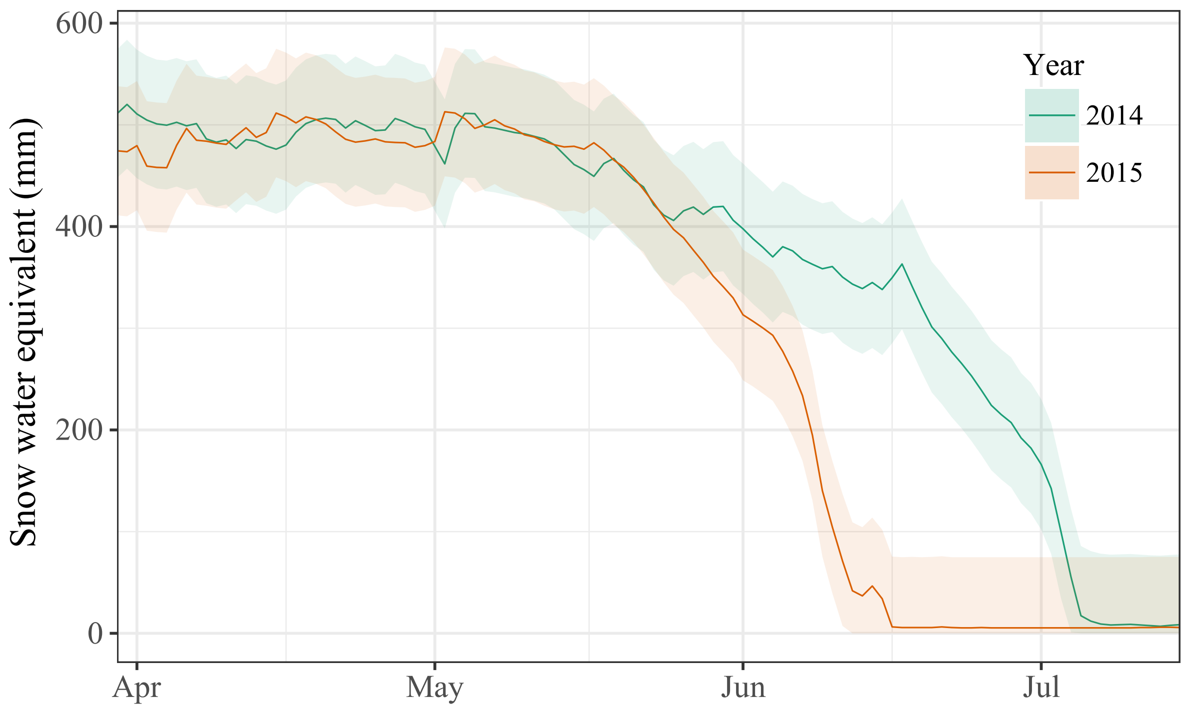

Figure 3.1. Spring SWE at CRHO Wx for the years 2014 and 2015. The last day of appreciable SWE was 22 days earlier in 2015 compared to 2014.

Table 3.1. Volumetric and depth equivalents of snow. Maximum wetland watershed SWE represents the volumetric SWE at each wetland, by year, determined by multiplying SWE estimated at CRHO by the upslope accumulation area for each wetland. The last two columns are the translation of that SWE, if the entirety of the snow volume were evenly distributed across each wetland's area. These results are not intended to represent the expected snowpack volumes that were translated to each wetland, but are simply used for comparative purposes with other water budget values. Note: Values in parentheses represent 2 standard deviations.

	Maximum wetland watershed SWE (m^3)		Maximum wetland SWE depth (cm)	
	2014	2015	2014	2015
Bowl	$5.2x10^3$ ($0.6 x 10^3$)	$5.1x10^3$ ($0.6 x 10^3$)	343 (42)	338 (42)
Slope Break	$2.8x10^3$ ($0.3 x 10^3$)	$2.8x10^3$ ($0.3 x 10^3$)	255 (31)	252 (31)
Valley	$109.2x10^3$ ($13.3 x 10^3$)	$107.7x10^3$ ($13.3 x 10^3$)	888 (108)	875 (108)

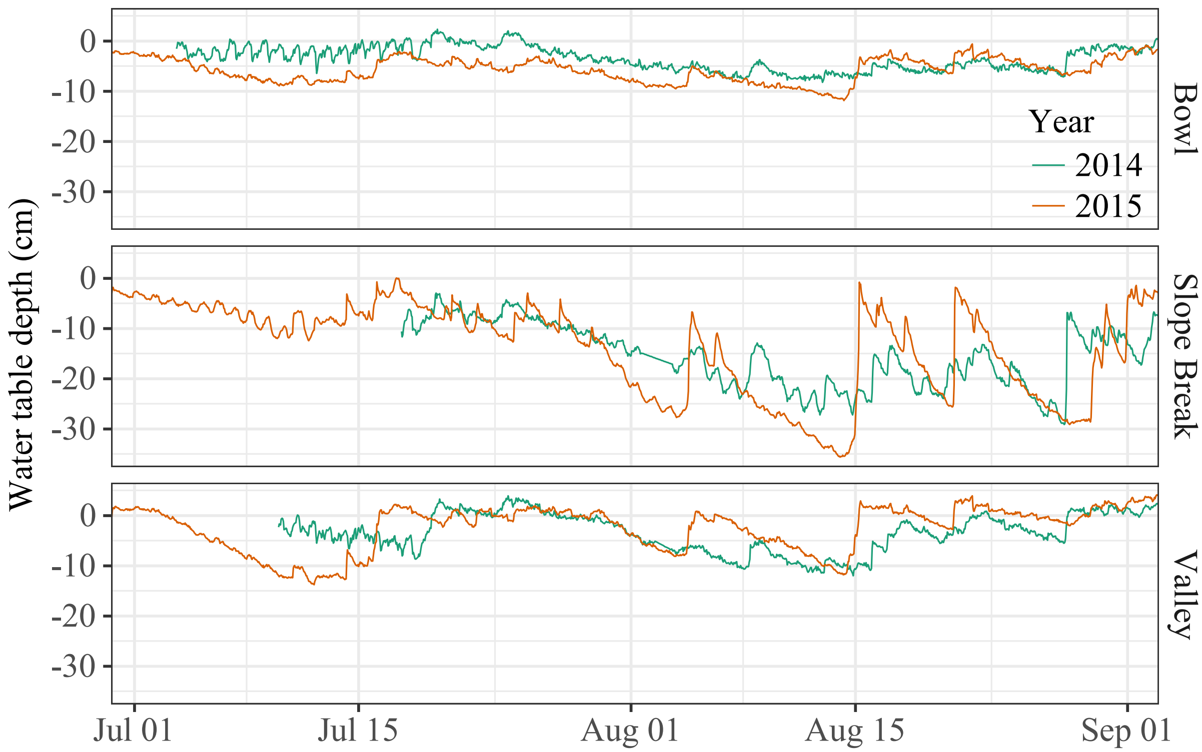

Figure 3.2. Hourly observed water table depths relative to the soil surface for each site and year.

Table 3.2. A summary of daily water table tendencies.

	Bowl		Slope Break		Valley	
	2014	2015	2014	2015	2014	2015
Mean position (SD) (cm)	-3.5 (2.3)	-5.7 (2.3)	-15.4 (5.8)	-13.6 (8.8)	-3.4 (3.7)	-2.6 (4.4)
Median change (cm/day)	-0.08	-0.47	-0.75	-1.36	-0.49	-0.49
Time above -20 cm (%)	100	100	73	74	100	100
Time above mean organic soil depth (%)	100	100	8	25	100	100

3.3. Soil Hydraulic Properties and Microtopography

Both Bowl and Valley contained peat (SCWG, 1998), but only Valley contained deposits deep enough for the wetland to be classified as a peatland according to the criteria outlined by

Canada's National Wetlands Working Group (NWWG 1997). Organic contents at Valley contained soil samples with the maximum observed proportion of peat, 69 % SOM by mass, while Slope Break had the lowest, with 1.8 % SOM by mass (**Table 3.3**). The deepest observed peat, >1.00 m, was located in Valley, whereas Bowl's deepest observed peat depth was 0.30 m. The minimum recorded bulk density, 0.05 g cm^{-3}, occurred at Valley and was associated with the highest organic content. The highest bulk density, 1.69 g cm^{-3}, was measured at Bowl and was associated with one of the lowest organic content soil samples, 3.6 % by mass. Overall, Slope Break had the highest average bulk density. Bulk densities associated with the peats of Bowl and Valley were much greater than values observed in other peatlands, which generally range from 0.05 to 0.25 g cm^{-3} (Lewis *et al.*, 2012).

Saturated hydraulic conductivity values ranged from 2.3 x 10^{-3} to 1.1 x 10^{-8} m s^{-1}, with the highest and lowest conductivities occurring at Bowl and Valley, respectively (**Table 3.3**). These estimates are similar to values observed in other mountain wetlands. For example, while studying a mineral alpine wetland in the Canadian Rocky Mountains of British Columbia, McClymont *et al.* (2010) found K_s values ranging from 1.4 x 10^{-5} to 2.5 x 10^{-7} m s^{-1}. These are also comparable to values found in fens located in the Colorado Rockies, which range from 3.6 x 10^{-4} and 4.6 x 10^{-6} m s^{-1} (Crockett *et al.*, 2015). Bowl ($R^2 = 0.44$; $p = 0.04$) and Slope Break ($R^2 = 0.56$; $p < 0.01$) had weak linear depth dependence in K_s, while Valley did not.

Specific retention values, θ_R, were fairly consistent between sites, but peat-bearing soils tended to have slightly larger porosity values. The peatlands also tended to have lower pore size distribution index values (n) than the mineral wetland, which indicates more negative pressures are required to remove equivalent amounts of water, all other parameters being the same. The largest and smallest capillary lengths (α^{-1}), 23.2 cm and 1.9 cm, respectively, occurred at Slope Break, illustrating the high variability in soil parameters at that wetland. However, on average Valley had the lowest capillary length while Bowl had the highest. These values are in line with previous estimates of peatland/wetland capillary lengths, which range from ~600 to 0.2 cm (Schwärzel *et al.*, 2006; Kettridge *et al.*, 2016).

During the period of observation, average Sy ranged from 0.34-0.45 at Bowl, 0.51-0.73 at Slope Break, and 0.62-0.75 at Valley. Sy decreased with water table depth in a predictable fashion at each wetland (**Figure 3.3**).

Table 3.3. A summary of the soil physical and hydraulic data for each wetland.

Bowl	No. samples	Mean	SD	Min.	Max.
Peat thickness (cm)	5	11	13	0	30
Soil organic matter (% mass)	4	22.5	23	3.8	51.6
Bulk density (g cm^{-3})	12	0.81	0.72	0.15	1.69
Log(K_S) (m s^{-1})	10	-3.61	0.63	-4.82	-2.64
ϕ [-]	4	0.68	0.24	0.45	0.97
θ_R [-]	4	0.28	0.24	0.00	0.49
α^{-1} (cm)	4	8.5	0.1	5.1	20.5
n [-]	4	1.74	0.47	1.16	2.20
Slope Break	No. samples	Mean	SD	Min.	Max.
Peat thickness (cm)	6	7	3	3	9
Soil organic matter (% mass)	5	6.8	4.9	1.8	14.8
Bulk density (g cm^{-3})	4	0.82	0.18	0.59	0.98
Log(K_S) (m s^{-1})	12	-4.61	1.08	-6.73	-3.31
ϕ [-]	5	0.65	0.15	0.49	0.85
θ_R [-]	5	0.28	0.25	0.00	0.55
α^{-1} (cm)	5	5.2	0.2	1.90	23.2
n [-]	5	1.82	0.65	1.18	2.55
Valley	No. samples	Mean	SD	Min.	Max.
Peat thickness (cm)	29	30	19	8	100+
Soil organic matter (% mass)	12	26.8	22.3	2.4	68.8
Bulk density (g cm^{-3})	4	0.43	0.29	0.05	1.03
Log(K_S) (m s^{-1})	14	-5.93	1.18	-7.95	-3.69
ϕ [-]	12	0.78	0.23	0.68	1.00
θ_R [-]	12	0.27	0.26	0.00	0.47
α^{-1} (cm)	12	3.8	0.3	2.7	15.8
n [-]	12	1.30	0.22	1.08	1.81

[-]: Unitless; SD: Standard deviation.

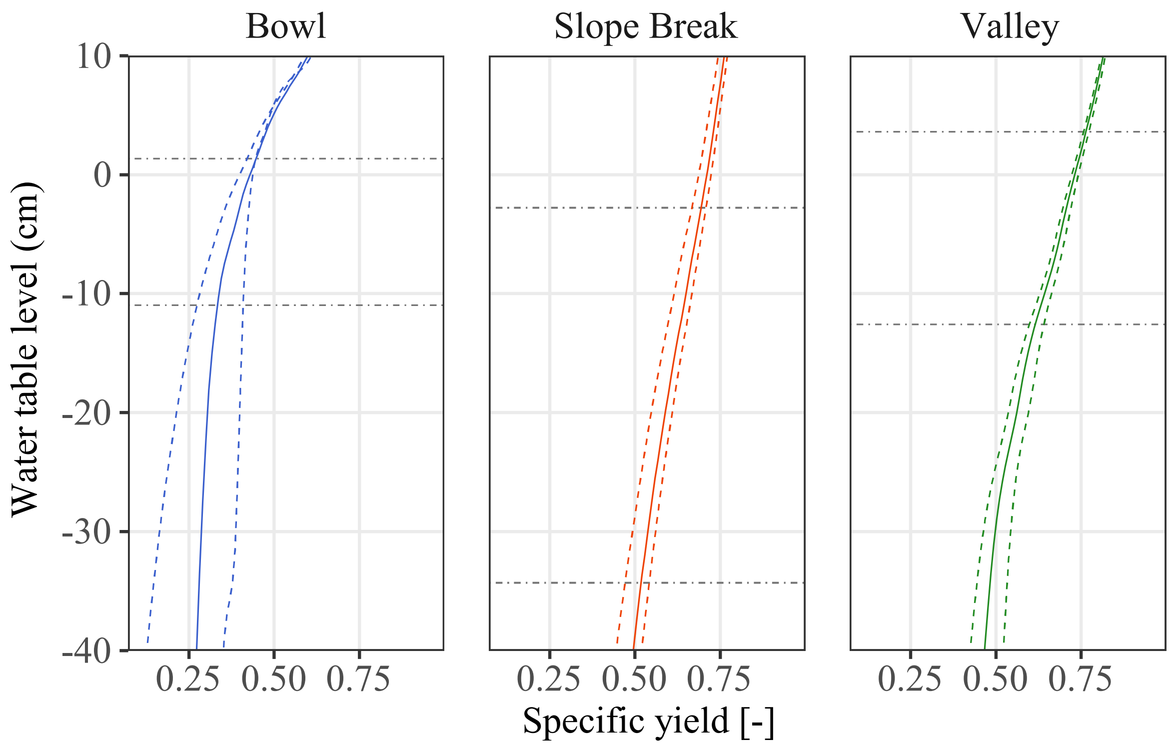

Figure 3.3. The relationship between water table depth (relative to the soil surface) and specific yield (colored, solid lines), as well as the relationship's sensitivity to soil heterogeneities (95% confidence intervals represented by colored, dashed lines). Minimum and maximum water table values observed during the summer are indicated by dash-dotted lines. Note that specific yield values do not reach 1 at the soil surface due to the influence of microtopgraphy.

Point cloud densities that resulted from the SfM inventory of microtopographic elevations were generally greater than 1 point cm^{-2}. Point clouds for Bowl and Slope Break are presented in **Figure 3.4** and **Figure 3.5**, respectively. A visual representation of the point cloud for Valley is presented in **Figure 3.6** (Figure 6 of Mercer and Westbrook 2016). Vertical RMSE values ranged from 8 cm to < 1 cm, the highest occurring at Valley and the lowest at Slope Break. The elevation range (maximum value – minimum value) of Bowl, Slope Break, and Valley is 2.0, 4.7, and 3.8 m, respectively. The median (IQR) elevation was determined to be 2486.3 m (0.3 m) at Bowl, 2383.2 m (1.5 m) at Slope Break, and 2362.8 m (0.8 m) at Valley.

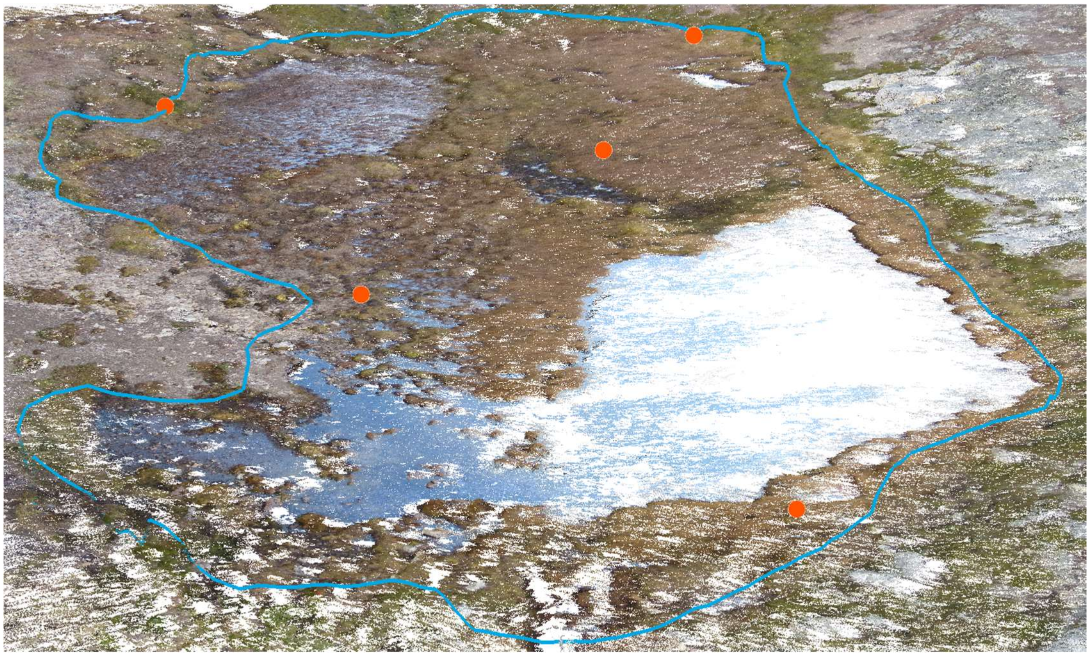

Figure 3.4. The point cloud generated for Bowl. Red dots are the locations of piezometer nests, while the blue line is the boundary of the wetland. Figure 2.2A was taken from a perspective to the upper left of the point cloud. Note that this image under-represents the true density of points acquired for the wetlands, due to visualization limitations.

Figure 3.5. The point cloud generated for Slope Break. Red dots are the locations of piezometer nests, while the blue line is the boundary of the wetland. Figure 2.2B was taken from a perspective to the upper left of the point cloud. Note that this image under-represents the true density of points acquired for the wetlands, due to visualization limitations.

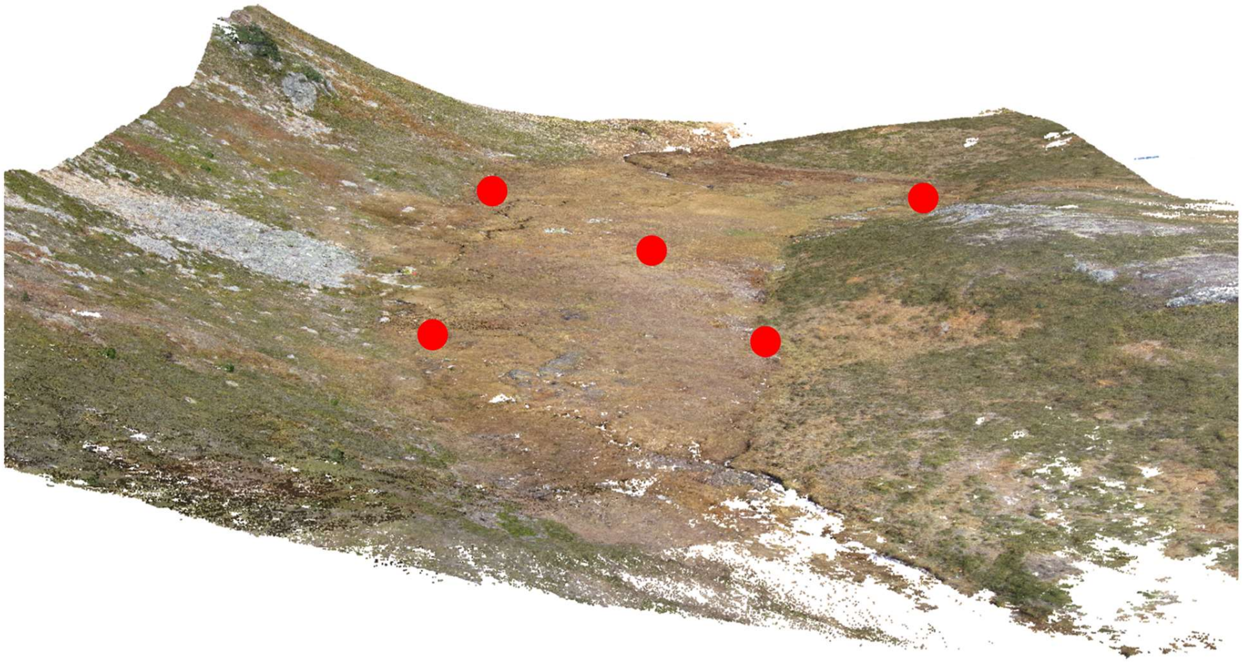

Figure 3.6. The point cloud generated for Slope Break. Red dots are the locations of piezometer nests. Figure 2-2C was taken from a perspective to the upper left of the point cloud (Mercer and Westbrook, 2016)[1].

3.4. Water Table Modelling, Water Balance, and Groundwater Flux

Daily atmospheric fluxes during the periods of observation in 2014 and 2015 are illustrated in **Figure 3.7.** Total precipitation during the 2014 period was 42 mm with 88 % occurring as rain, and 30 mm in 2015 with 86 % occurring as rain. Total potential evapotranspiration losses at the wetlands were estimated as 349 and 539 mm in 2014 and 2015, respectively. Differences were largely due to the lengths of the seasons. Average wetland ET rates were 4.3 and 5.1 mm day^{-1} in 2014 and 2015, respectively.

The use of precipitation, evapotranspiration, and discharge alone (i.e., the "bog model") did a poor job of explaining the observed water table dynamics at all three wetlands (**Figure 3.8**). Those model treatments that included direct flows only (no lateral flows) only did marginally better in producing water table simulations that were closer to observed water tables for 3 of the 4 site/year combinations investigated. In the case of Bowl, the mean model run indicated the

[1] This figure appears as Figure 6 in: Mercer JJ, Westbrook CJ. 2016. Ultrahigh-resolution mapping of peatland microform using ground-based structure from motion with multiview stereo. Journal of Geophysical Research: Biogeosciences 121 (11): 2901–2916 DOI: 10.1002/2016JG003478. Jason Mercer is the major contributor and lead author of the manuscript. Cherie Westbrook was the primary supervisor, as well as assisted with writing and structure.

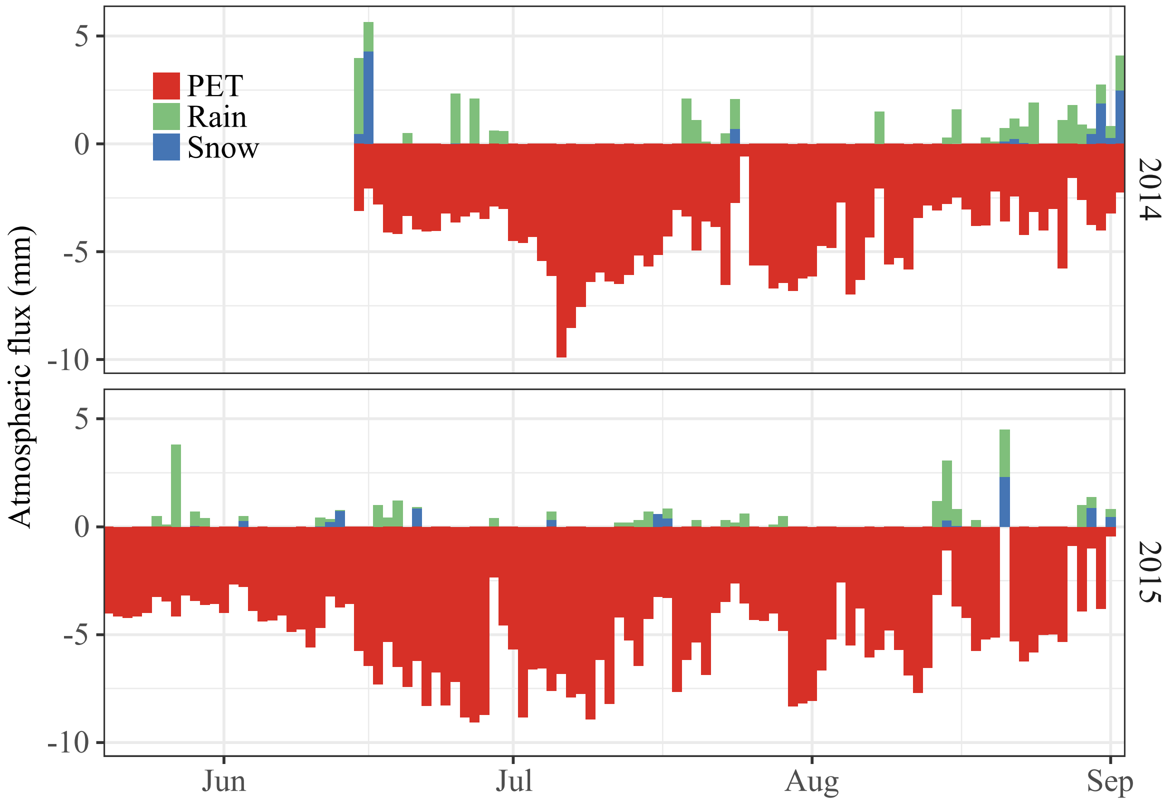

Figure 3.7. Daily wetland potential evapotranspiration loss (expressed as a negative) and rain and snow accumulations (positive) for the summers of 2014 and 2015.

water table depth at the end of the growing season should be 18.6 m deeper than was observed, representing a volumetric deficit of 27, 900 m^3 of water. A similar, though not as extreme, outcome was also observed at Valley. In that case, a water table of about 70 cm deeper than observed was predicted, representing a volumetric deficit of, on average, 8,600 m^3. In the case of Slope Break, however, the bog model predicted values within the 95% confidence interval 35% of the time in both 2014 and 2015. All model treatments that included lateral fluxes did equally poorly at predicting water table position, suggesting a fundamental flaw in the conceptualization of those models.

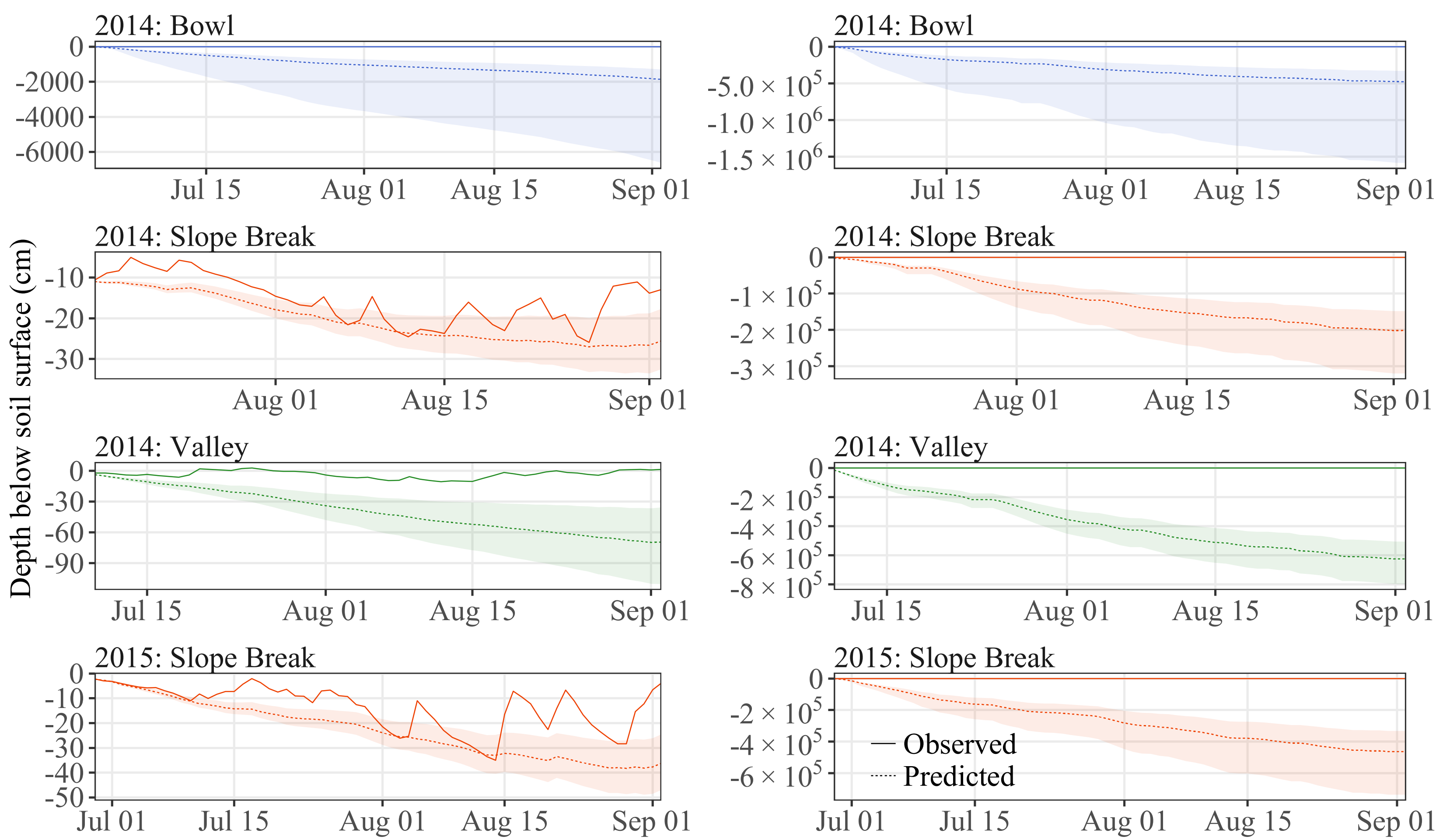

Figure 3.8. (Left column) Water table prediction scenarios with no lateral flow included. (Right column) Prediction scenarios that do include lateral flow from upslope areas. The observed (solid line) versus most likely (dashed line) daily water table values for a given year and site. Shaded regions represent 95% confidence interval for all model runs. Note the differences in scale for each plot.

3.5. Net Groundwater Contributions

3.5.1. Water Balance

For all wetlands, the cumulative net groundwater contributions were positive (**Table 3.4**), even after accounting for hydrologic uncertainties and edaphic heterogeneities (**Figure 3.9**). In the wetlands where there is peat, net groundwater was larger than precipitation inputs, implying groundwater is a dominant source water to both Bowl and Valley during the growing season. There was a large amount of uncertainty regarding the importance of net groundwater contributions to Slope Break, as illustrated in the overlap between the 95% confidence interval of our model and the zero-flux line. At Slope Break, net groundwater contributions were smaller than PET. However, this does not rule out groundwater importance at Slope Break, as subsurface fluxes (discharge and recharge) could have been balanced.

Table 3.4. 50th percentile change in volumetric water balance components for the observed periods of the growing season. Note that negative values indicate losses to the system.

	Bowl		Slope Break		Valley	
	2014	2015	2014	2015	2014	2015
Net groundwater (m^3)	8,200	ND	83	220	4,600	ND
Change in storage (m^3)	-1.0	4.0	-16	-12	310	260
Discharge – inlet (m^3)	NA	NA	NA	NA	6.8×10^5	ND
Discharge – outlet (m^3)	-7,900	ND	NA	NA	-6.9×10^5	ND
Rain (m^3)	170	170	100	130	1,300	1,400
Summer snow (m^3)	19	24	14	18	160	200
PET (m^3)	-420	-500	-220	-370	-2,900	-4,100

NA: Not applicable; ND: No data.

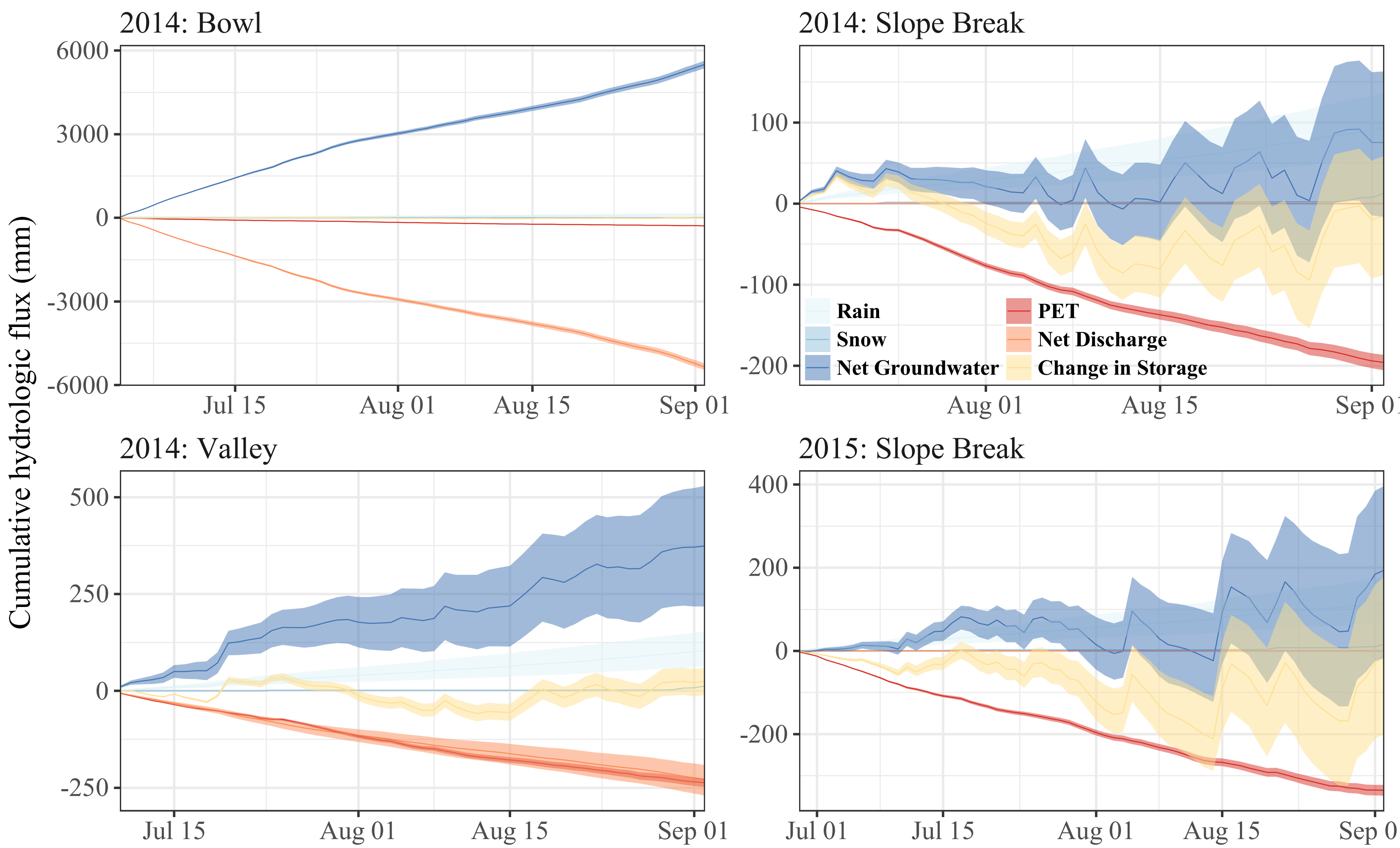

Figure 3.9. Cumulative changes in water balance components for the years when all surface fluxes could be accounted for at a given site. Net discharge represents the difference between inlet and outlet discharge, if applicable. Losses to the system are expressed as a negative. Data has been areally weighted, converting volumes into depths, to provide easier comparison between sites and years.

3.5.2. VHG and Specific Discharge

All wetlands expressed both positive and negative vertical hydraulic gradients (VHGs), implying both discharge (positive) and recharge (negative) processes occurring in different parts of each wetland. Valley had both the most positive and negative vertical hydraulic gradients during the summer of 2014, which were -0.7 and 1.48, respectively. VHGs at the other two wetlands were comparatively subdued. **Figure 3.10** illustrates both the spatial and temporal pattern of VHGs across wetlands, including one set of measurements made after the growing season. At Bowl, VHGs were positive during the summer near the top of the wetland, indicating discharge, and negative at this base, indicating recharge. Generally, the opposite pattern was observed at Slope Break, where the large gradients measured occurred at the base of the wetland, closest to Helen Creek. VHG values at Valley were transient, with the spatial distribution of hydraulic gradients reversing over the summer and again in the autumn – a fairly uncommon phenomenon.

During the summer, the average vertical specific discharge ($n = 3$) at Bowl was 128 cm day^{-1} while recharge was 518 cm day^{-1}. Mean vertical specific discharge and recharge at Slope Break ($n = 3$) was 65 and 3 cm day^{-1}, respectively, while they were 13 and 2 cm day^{-1}, respectively, at Valley ($n = 3$).

3.6. Hydrochemistry

The local meteoric water line (LMWL) for the study period ($R^2 = 0.98$) was $\delta^2H = 7.70 \cdot \delta^{18}O - 5.13$, which is in line with long-term LMWL estimates from Calgary, AB, Canada (Peng *et al.*, 2004). Snow and rain stable isotope values bracketed isotopic signatures of other source waters throughout the Helen Creek basin, making them good end members for distinguishing between snow and rain.

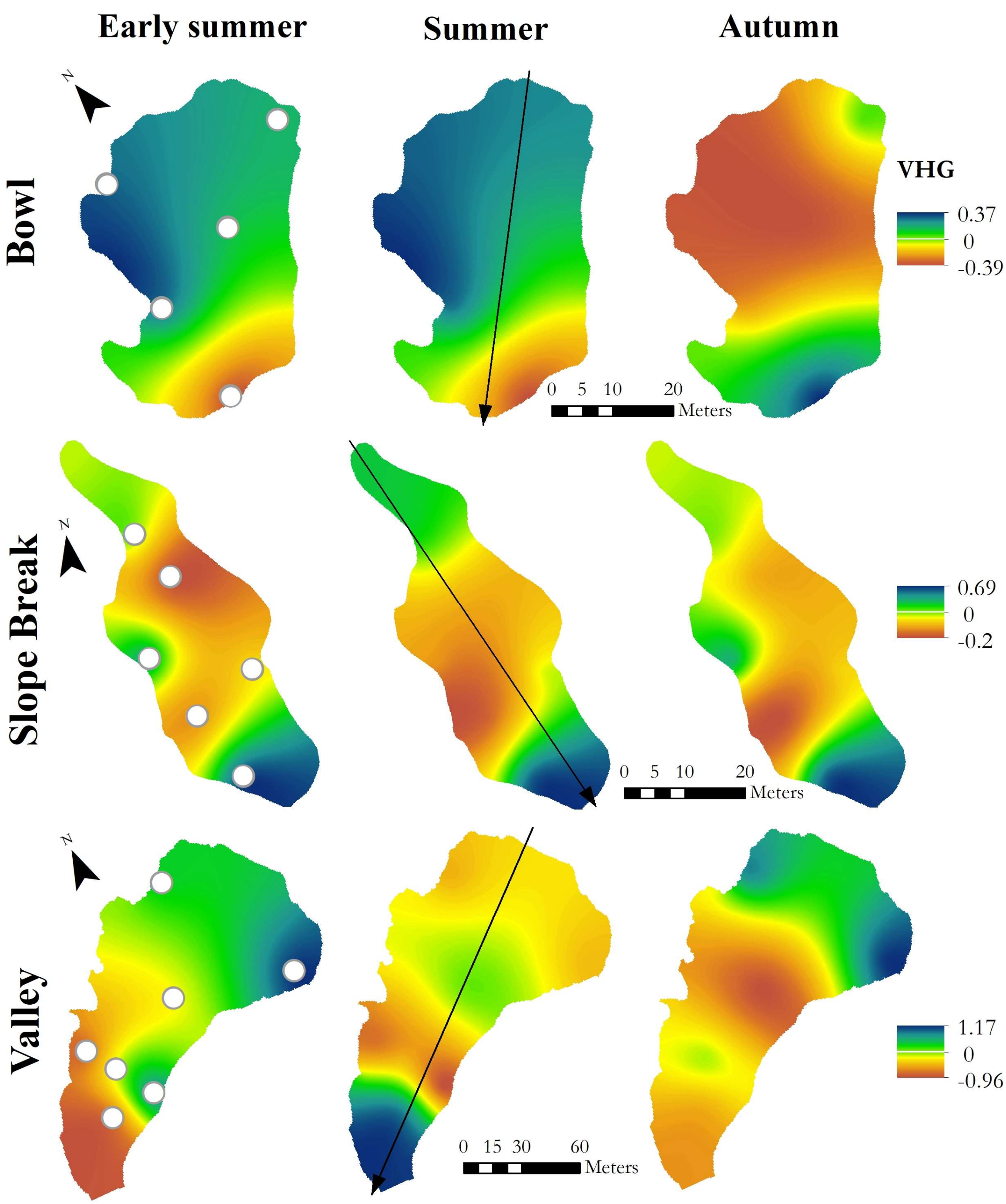

Figure 3.10. Select vertical hydraulic gradient maps for the summer (left two columns: 7/25 and 8/8/2014) and autumn (after the end of the growing season: 9/27/2014). Positive values are areas of discharge (blue-green), while negative values are areas of recharge (yellow-orange). Piezometer nests are indicated by circles (white) and the main vector of horizontal flow for each wetland is indicated by an arrow (black). Gradients were interpolated between nests using splines for visualization purposes only. Note the different spatial and legend scales for each wetland.

A summary of the hydrochemical data inventoried in the Helen Creek Research Basin (HCRB) is provided in **Table 3.5**. The isotopic signature of rain (mean of -86.7 δ^2H and -10.6 δ^{18}O) was significantly different from all other source waters (n = 5; $p << 0.01$ for both isotopes). However, δ^2H and δ^{18}O were not significantly different when comparing groundwater (n = 26) and surface water (n = 39; p = 0.99 for both isotopes) or groundwater and snow (n = 7; p = 0.12 for δ^2H and p = 0.31 for δ^{18}O). In regards to SC, groundwater (n = 38) and surface waters (n= 62) were not different from one another (p = 0.99). Groundwater was significantly different to both snow (n = 9; $p < 0.01$) and rain (n = 5; p = 0.05). Surface water SC was also significantly different to both snow (p = 0.02) and rain (p = 0.05). Given the differences in hydrochemical values for the source waters identified, and the limited evaporation signal, it may be considered reasonable to use δ^2H as the end member for separating rain and snow from each other.

Table 3.5. The mean (SD) values of different source waters inventoried in HCRB, by geochemical constituent. Data significantly different from one another are marked with a symbol. Letters indicate which source waters are different from each other within a column.

Source	δ^2H (‰)	δ^{18}O (‰)	SC (μS cm^{-1})
Rain (R)	-86 (27)	-10.6 (3.4)	0 (0)
Snow (Sn)	-162 (14)$^{‡(R)}$	-20.6 (1.7)$^{‡(R)}$	9.1 (13.3)
Groundwater	-150 (8)$^{‡(R)}$	-18.8 (1.5)$^{‡(R)}$	95.5 (75.9)$^{†(R,Sn)}$
Surface water	-150 (14) $^{‡(R)}$	-18.8 (2.7)$^{‡(R)}$	75.2 (58.5)$^{*(R,Sn)}$

Significance level determined at p = 0.05.

Groundwater specific conductance values in HCRB are similar to those found in alpine watersheds nearby, which range from 90-260 μS cm^{-1} (Roy *et al.*, 2011). However, groundwater and surface water could not be distinguished using SC, suggesting the two sources are well mixed. Because of this potential mixing, these end-members thus were lumped into a single ‘subsurface’ category. Further, because of its large SC and VHG values, well water from Valley was used as the end member for defining the subsurface group. End member values used in the end member mixing analysis are presented in **Table 3.6**.

Results of end member mixing analysis are illustrated in **Figure 3.11**. Of the waters sampled at the study wetlands and across HCRB, the analysis indicated Valley had the largest average contribution of subsurface water at 53 %. Snow and rain constituted the other 38 and 9

%, respectively. The average subsurface contribution of waters sampled at Bowl was 41 %, while snow was 54 % and rain was 5 %. Slope Break had the highest average contribution of snow at 65 %, while subsurface water and rain constituted the other 29 % and 6 %, respectively. In non-wetland areas (i.e., basin-scale) of HCRB, snow was the dominant signature at 49 %, while rain was 24 % and subsurface water was 27 %. Because all hydrochemical data were collected during or just after snowmelt, and are not integrated over time or flux weighted, it is expected that these values will be more reflective of peak snowmelt conditions, and not growing season conditions.

Table 3.6. End member hydrochemical signatures for the three source waters evaluated. Low and High values are associated with confidence intervals at the 95th percentile.

	δ^2H (‰)			SC ($\mu S\ cm^{-1}$)		
End member	Low	Median	High	Low	Median	High
Subsurface water	-153	-150	-146	120	181	250
Snow	-172	-162	-153	1	5	11
Rain	-106	-86	-66	0	0	0

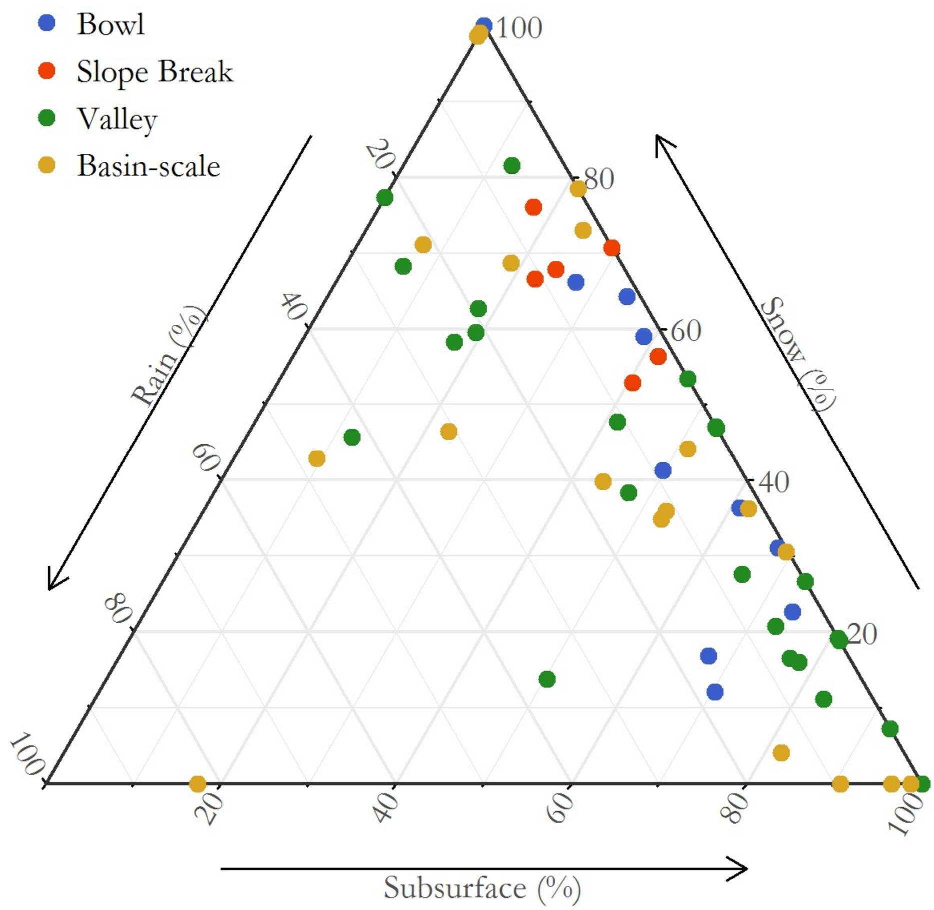

Figure 3.11. The proportions of end members for waters sampled at each site and across the Helen Creek Research Basin (basin-scale). The figure was made using the R package 'ggtern' (Hamilton, 2018).

CHAPTER 4: DISCUSSION

In this study, I investigated the hydrologic processes associated with the water table dynamics of alpine wetlands in contrasting topographic settings to better understand the potential vulnerability of alpine wetlands to changes in environmental condition. Several processes were considered as possible factors contributing to the observed water table dynamics. Groundwater is likely an important contributor to wetlands during the growing season, especially in peat-bearing wetlands. This suggests that the bog model of hydrology in alpine wetlands has limited utility, at least in HCRB. Meso-scale topography (as opposed to microtopography) did not play an obvious role in controlling water table dynamics. Soil hydraulic properties, including specific yield, do seem to play a particularly important role in regulating water table position. Together, these observations suggest that peat-bearing alpine wetlands may be more resilient to changes in climate than previously expressed in the literature, but there are still many process uncertainties that require further investigation.

4.1. Hydrological processes contributing to water stability

It was expected that water tables would be lowest during the summer in 2015, when snowmelt occurred earlier, due to the prolonged period over which ET losses could occur. However, despite a nearly one-month advance in snowmelt timing, average and minimum water table depths were not substantially different between the two study years. The unexpected lack of correlation between timing of snowmelt and water table depth is unusual. While there are no studies documenting the effect of advanced snowmelt timing on wetland water table dynamics, there is empirical evidence for the opposite condition. Millar *et al.* (2017) showed, in the US Rocky Mountains, that a later melt season (shifting from March to June) was associated with as much as a 35 cm drop in the mean growing season water table. In their case, the wetland being studied was dependent on groundwater, suggesting transience in the groundwater system may be as important as snowpack dynamics for some mountain wetlands. At HCRB, upslope SWE volumes were smaller than lateral inputs (i.e., net groundwater and discharge at the inlet) to Bowl and Valley for 2014 (the only year lateral inputs were estimated). This discrepancy in the water balance, in conjunction with other hydrological process estimates, highlights the importance groundwater must play in water table maintenance. In such cases, the contributing area to the wetland is likely to be much larger than that estimated from the surface watershed. The importance of groundwater to the wetlands at HCRB is consistent with findings from other studies of aquatic alpine ecosystems (**Table 4.1**).

Table 4.1. A comparison of the importance of groundwater in various aquatic alpine ecosystems.

Ecosystem	Study	Range and region	Groundwater (%)
Wetlands	Nielson (2008)	Rocky Mountains, CO, USA	44-54
	Present study*	Rocky Mountains, AB, Canada	29-53
Lakes	Gurrieri and Furniss (2004)	Rocky Mountains, MT, USA	58-84
	Hood *et al.* (2006)	Rocky Mountains, BC, Canada	30-74
Streams	Caine (1989)	Rocky Mountains, CO, USA	50
	Williams et al. (1993)	Sierra Nevada, CA, USA	62
	Campbell et al. (1995)	Rocky Mountains, CO, USA	10-75
	Mast et al. (1995)	Rocky Mountains, CO, USA	45
	Laudon and Slaymaker (1997)	Coast Range, BC, Canada	25-90
	Sueker et al. (2000)	Rocky Mountains, CO, USA	37-89
	Clow et al. (2003)	Rocky Mountains, CO, USA	75
	Huth et al. (2004)	Sierra Nevada, CA, USA	10-20
	Liu et al. (2004)	Rocky Mountains, CO, USA	18-64
	Brown et al. (2006)	Pyrénées, France	10-53

* Based on mean mixing model values from spring hydrochemical samples - potentially a low estimate of groundwater importance.

Despite the apparent prevalence of groundwater, it is still expected that snowpack volumes are exceptionally important to alpine wetlands – including those that seem dependent on groundwater – considering snowmelt is the major hydrologic input to temperate alpine watersheds (DeBeer and Pomeroy, 2010), and so is critical for recharging groundwater (Hood and Hayashi, 2015). However, snowpack volumes in the Helen Creek Research Basin were not different between years, limiting inference related to the interactions between SWE and water table depths. However, there are a growing number of mountain wetland studies that have recorded SWE and water table depths for wetlands that are likely supported by groundwater systems. By extracting information from these studies and adding wetlands from HCRB, a correlation between SWE and mean growing season water table emerges (**Figure 4.1)**. While there is still a lot of unexplained variance, the correlation between peak SWE and water table position is fairly strong ($R^2 = 0.49$, $p << 0.01$). The relationship between SWE and water table depth indicates that, on average, a 10 mm increase (decrease) in peak SWE is associated with a 1.2 cm increase (decrease) in mean growing season water table, relative to the soil surface. Though the available information used to generate **Figure 4.1** was restricted to the Rocky Mountains (because this is where most data are available), the wetlands included in the analysis span large latitudinal, elevational and temporal (3 decades) gradients, suggesting the correlation between SWE and mean water table position might be regionally generalizable. The correlation is likely to be widely valuable, especially to hydrological modelling. For example, Huntington and Niswonger (2012) recently used wetland elevation to improve groundwater process estimations for stream discharge modelling.

It may at first seem inconsistent that peak SWE is so strongly correlated to the mean growing season water table of groundwater-dependent wetlands. However, the correlation may be reasonable in the context of piston or translatory flow. Translatory flow occurs when infiltrating water displaces older water in the subsurface network, thus promoting downgradient discharge (Mast *et al.*, 1995; McGlynn and Seibert, 2003; Williams *et al.*, 2016). In this context, snowmelt may have both direct and indirect influences on alpine wetlands. Directly, *in-situ* (and nearby) snowmelt can set the initial water table depth during the spring. Indirectly, snowmelt is recharging the groundwater system, thus providing hydraulic pressure, which causes discharge to down gradient wetlands. Future studies that explore the heterogeneities that regulate the partitioning of snowmelt between surface and subsurface flowpaths would be valuable for understanding alpine wetland source waters.

Regardless of mechanism, groundwater expression appears to be very pronounced at the wetlands studied herein, as evidenced by the large vertical hydraulic gradients, the large specific discharge values, and the end member analysis. While the VHG values in wetlands of HCRB are somewhat comparable to those found by Woods *et al.*, (2006), who studied a subalpine wetland in the Colorado Rockies, the extremes seem to be much greater than those that have been recorded in lowland wetland systems (**Table 4.2**). A similar sentiment can be expressed for vertical specific discharge. While no comparable mountain wetland studies have estimated groundwater specific discharge, lowland wetland values of discharge and recharge range from

0.001 to 4.65, and 0.004 to 14 (cm day^{-1}), respectively (Koerselman, 1989; Roulet, 1990; Devito *et al.*, 1996; Hunt *et al.*, 1996; Choi and Harvey, 2000; Ferone and Devito, 2004; Todd *et al.*, 2006). As there are so few studies of alpine groundwater hydrology, it is not entirely clear why wetland specific discharge values in HCRB are so high. Determination of transit times in alpine groundwater systems are needed to better understand how changes in climate might alter subsurface flows contributing to wetlands (e.g, Lessels *et al.*, 2016; Mosquera *et al.*, 2016), and thus wetland resilience.

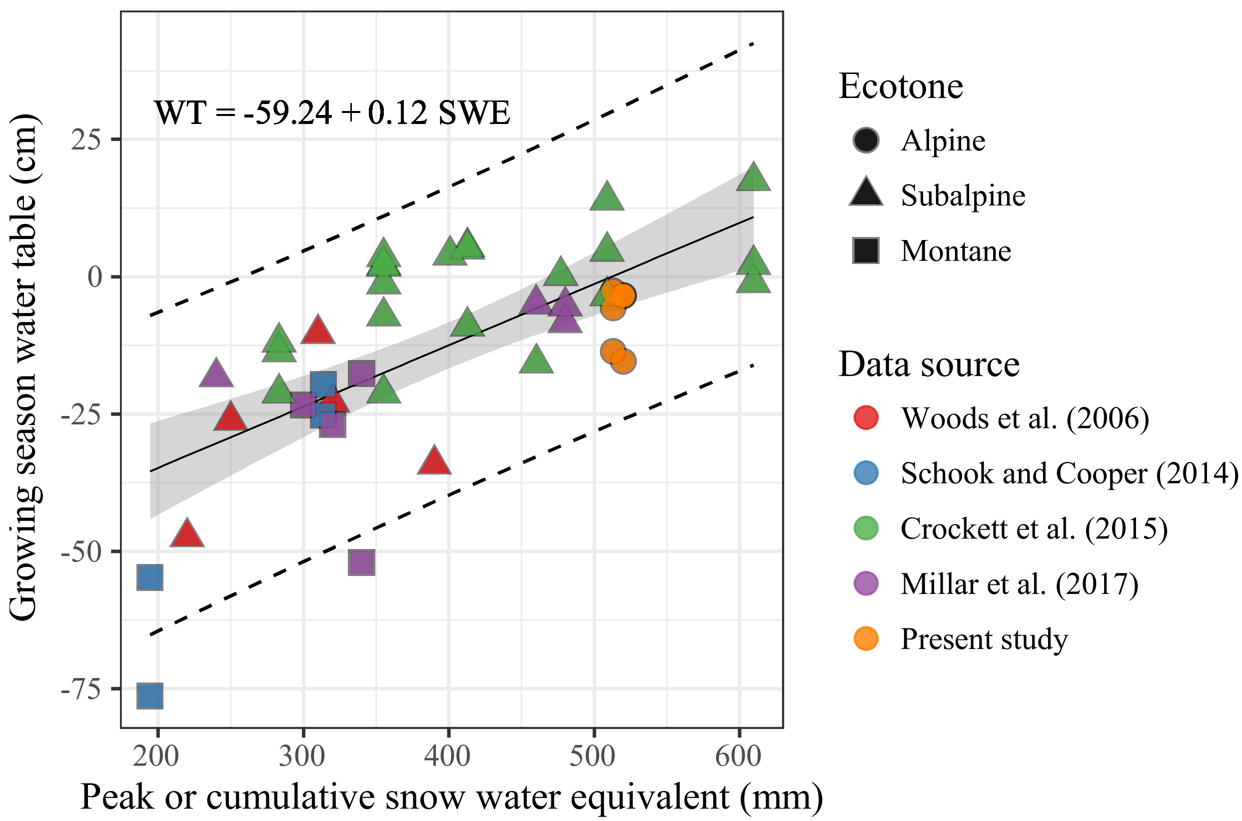

Figure 4.1. A comparison of SWE to mean growing season water table depth, relative to the soil surface, for published studies of groundwater-dependent mountain wetlands and peatlands. The solid line represents the relationship defined by the equation in the upper left-hand corner of the figure ($R^2 = 0.49$, $p << 0.01$), while the shaded area is the 95% confidence interval, and the dashed lines are the 95% prediction interval. Data Thief III (Tummers, 2006) was used to extract data from (Schook and Cooper, 2014) and (Crockett *et al.*, 2015).

Summer precipitation may also be contributing to the water table stability of alpine wetlands. Though summer precipitation was small compared to other water budget inputs, all three wetlands expressed disproportionately large responses to summer precipitation events (**Figure 3.2 and 3.6**) – especially Slope Break. This phenomena has been previously observed in a number of wetlands, including an alpine wet meadow (McClymont *et al.*, 2010). Such responses have been attributed to both internal and external conditions (Heliotis and DeWitt, 1987; Gerla, 1992; Sumner, 2007; McClymont *et al.*, 2010), the mechanisms of which are considered later in this section. Regardless of mechanism, however, these observations suggest that even small summer precipitation events can be important in attenuating the water table decline triggered by a high summer evapotranspiration demand (Duval and Waddington, 2011). Considering alpine zones tend to receive more precipitation than their lowland counter parts due to cooling and environmental lapse rates (Grabherr *et al.*, 2010), this could mean that summer precipitation may be an increasingly important water table control in a future wherein snowpacks are shallower.

Table 4.2. A comparison of vertical hydraulic gradients measured in the wetlands of this study, lower elevation wetlands, and one study focusing on alpine streams.

Study	Location	Ecosystem	Min. [-]	Max. [-]
Present study	Canadian Rockies, AB, Canada	Alpine wetlands	-0.7	1.48
Woods *et al.* (2006)	Rocky Mountains, CO, USA	Subalpine wetland	-2	0.6
Moorhead (2003)	Appalachian Mountains, NC, USA	Montane wetland	-0.13	0.15
Devito *et al.* (1996)	ON, Canada	Lowland wetland	0.01	0.1
Choi and Harvey (2000)	Everglades, Florida, USA	Lowland wetland	-0.1	0.3
Almendinger and Leete (1998)	MN, USA	Lowland wetland	0.037	0.146
Mann and Wetzel (2000)	AL, USA	Lowland wetland	-0.2	0.24
Frisbee *et al.* (2011)	Rocky Mountains, CO, USA	Alpine streambeds	0.03	0.2

Evapotranspiration was prolonged and rates generally higher in 2015, compared to 2014. However, these changes did not produce notable changes to average summer water table depths. This result contrasts with the findings of Zhang *et al.* (2016) who showed that climate changes that are prolonging the ET period in the Tibetan Plateau are leading to surface drying and runoff reductions from alpine wetlands. This may indicate that the hydrological response of alpine wetlands to climate change is not uniform. Needed, then, is a better understanding of the linkages between water table dynamics of alpine wetlands in the Canadian Rockies and a changing ET regime. For example, eddy covariance systems (Drexler *et al.*, 2004; Rosenberry *et al.*, 2007) could be used to better determine which alpine wetlands are potentially more or less vulnerable to climate change.

4.2. Topographic and geologic considerations in regulating mountain wetland hydrology

Topography is generally considered a first-order hydrological control (Tóth, 1962; Beven and Kirkby, 1979; Brinson, 1993; Winter, 1999; Buttle, 2006). As such, topography is a key component of wetland functional classifications, such as the hydrogeomorphic classification system (Brinson, 1993; Smith *et al.*, 1995). Thus, wetlands that occur near breaks in slope (i.e., concave profile geometry) are generally thought to be more groundwater dependent (i.e., slope wetlands; Woods *et al.*, 2006), while wetlands with both concave profile and planform geometries are considered to be largely precipitation-dependent (i.e., depressional wetlands). Based on these criteria, it was expected that Slope Break would be groundwater dependent, Bowl would be more-or-less snowmelt dependent, and Valley would represent a mix of both water sources. Instead, all three wetlands were similarly dependent on groundwater, as discussed earlier, suggesting surface topography may be a poor predictor of alpine wetland hydrology.

Other mountain studies have also found that topography may not be a key regulator of wetland hydrology. For example, while investigating landscape patterns associated with fens in the Cascade Range, USA, Aldous *et al.* (2015) found that geologic factors were a better predictor of wetland type and hydrology than topography. This observation is consistent with a growing number of studies that have linked mountain wetland locations to geologic features. In particular, mountain and alpine wetlands have been associated with volcanic deposits (Aldous *et al.*, 2015), bedrock discontinuities (Stein *et al.*, 2004; Cooper *et al.*, 2010), alluvium/colluvium deposits, including talus fields (Caballero *et al.*, 2002; Stein *et al.*, 2004; McClymont *et al.*, 2010; Baraer *et al.*, 2015; Gordon *et al.*, 2015), and faults (Stein *et al.*, 2004). This seems to signify that the geologic framework on which wetlands sit may play a more fundamental role in controlling alpine wetland hydrology, than does topography – especially for those wetlands reliant on groundwater.

In some respects, strong associations between alpine wetlands and the hydrogeologic system could mean alpine wetlands are more resilient to changes in climate than currently expected, as groundwater can often act as a buffer to other hydrological changes (Kløve *et al.*, 2014). Practically speaking, this may mean that managers could make use of existing geologic information (e.g., geologic maps, borehole logs) to inform predictions of wetland climate vulnerabilities. However, not all geologic networks are the same. For example, Aldous *et al.* (2015) found that many of the groundwater wetlands that they studied in the Cascade Range were supported by intermediate groundwater flow networks. Generally, these intermediate networks are expected to be less resilient to changes in climate than are regional aquifer systems given differences in residence times. Thus, the nature of the hydrogeologic network will be an important aspect of wetland resilience. In this regard, future research is needed to investigate linkages between the hydrogeologic network, water transit times, aquifer size, and landscape settings of alpine wetlands, and develop a mountain-specific wetland classification system.

Based on hydrological insights from this research and geologic information from the literature, **Figure 4.2** provides an updated conceptual understanding of the hydrological processes important to alpine wetlands.

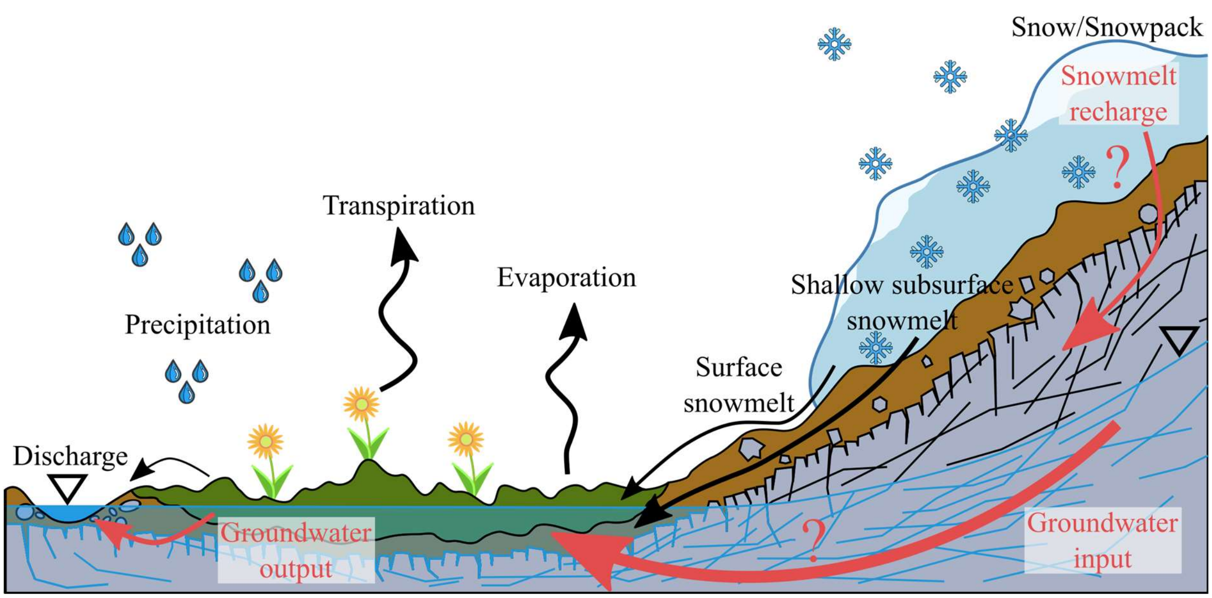

Figure 4.2. Updated conceptual understanding of the hydrological processes and geological conditions that may be important to alpine wetlands. Red arrows represent likely flow paths important to the hydrological functioning of alpine wetlands, but that required much more investigation, while the relative size of the arrows represent their potential importance.

4.3. Soil hydraulic properties in relation to water table stability and resilience

4.3.1. Water table response to summer precipitation in the context of soil hydraulic parameter uncertainties

As mentioned earlier, wetlands in HCRB often expressed a disproportionate water table response to summer precipitation events. This phenomena is commonly associated with both capillary (Heliotis and DeWitt, 1987; Gerla, 1992) and storage effects (Heliotis and DeWitt, 1987; Nachabe, 2002; McLaughlin *et al.*, 2014). Capillary effects occur when new water fills the partially filled pores above the water table, promoting a rapid transition from negative to positive pressure, which is manifest as a water table rise in unconfined aquifers, like those in most wetlands. Storage effects, on the other hand, are water table changes due to changes in storage, as predicted by specific yield. Because capillary effects were explicitly represented in the specific yield function of both the 1D water table model and the water budget, capillarity is not expected to explain the observed responses of the water table to summer precipitation. However, there are still a number of uncertainties associated with capillarity, and other soil hydraulics, that could influence the interpretation of results from this study.

One such uncertainty was the estimation of hydraulic parameters using a pressure plate extractor. Pressure plates are used to determine volumetric relationships at very negative pressures (generally between -100 and -1,500 kPa; ~-1 x 10^3 to -1.5 x 10^4 cm of water), representing dry to very dry conditions – conditions not experienced by the wetlands in this study, as indicated by the water table depths. Because capillary effects occur at pressures very close to zero, fitting soil water retention curve models from data generated at very negative pressures could, conceivably, result in highly non-unique estimates of capillarity, as well as the other soil hydraulic parameters used in the van Genuchten equation. Such uncertainties could be improved by using instruments meant to be operated at pressures closer to 0 to -100 kPa. These instruments/techniques include tempe cells (Figueras and Gribb, 2009), evaporation methods (Gardner and Miklich, 1962; Wind, 1968), centrifugation (Šimůnek and Nimmo, 2005), or the use of a hanging water column (Hillel, 2004).

The use of pressure plates also requires the destruction of secondary soil structure, which is an important control of wetland soil hydraulics, including capillarity (Hayward and Clymo, 1982; Rezanezhad *et al.*, 2016). This destruction results in the collapse of macropores, such that laboratory estimates of hydraulic parameters are no longer representative of field conditions.

Disruption of the pore network is likely not an important factor at very negative pressures, because larger pores would have already drained, but could be a significant issue at pressures closer to 0 kPa, thus biasing parameter estimates.

Hysteresis in the water table can often be prominent in wetland soils, but was not considered in this study. This is likely appropriate for the conditions found in HCRB. Price and Schlotzhauer (1999) found that hysteretic effects became more muted in peat soils as bulk density increased. Soils in this study were an order of magnitude denser than those analyzed by Price and Schlotzhauer (**Table 3.3**), thus reducing the likelihood of hysteretic effects. Further reducing the likelihood of the importance of hysteresis was the use of a time-step that approximated hydraulic equilibrium for the soil and hydraulic head conditions observed (Nachabe, 2002).

Spatial heterogeneity in parameter estimates was also not considered here. In particular, depth-dependent changes in soil hydraulic parameters were not used. Generally, wetland soils express strong vertical gradients (Quinton *et al.*, 2008), but considering changes in K_s were only weakly correlated with depth for two of the wetlands (Bowl and Slope Break), this bulk parameter assumption may be reasonable. Specific yield values were found to decrease with depth, which is consistent with both theoretical (Sumner, 2007; Dettmann and Bechtold, 2016) and empirical treatments (Sherwood *et al.*, 2013; McLaughlin and Cohen, 2014) of Sy, again indicating the bulk treatment of wetland soil hydraulic parameters was reasonable for the study wetlands.

The net effect of the above soil hydraulic uncertainties is likely an underestimation of the importance of capillary action, promoting an overestimate of net changes in lateral flows, and thus an over-prediction of the importance of groundwater. However, the combination of uncertainty analysis and consistency across methods (i.e., mixing models and discharge measurements) suggest that even though there are potentially still large uncertainties in the estimated hydrological processes, the overall signal is greater than any methodological noise.

Given the above interpretation, it seems likely that much of the water table response to summer precipitation events was due to storage effects associated with an increase in net inflows. Though direct precipitation did not explain observed changes in the water table, the methods used here cannot rule out the possibility that summer precipitation is important. That is, it is still possible that summer precipitation is being translated to the study wetlands via relatively shallow subsurface flow paths. This would be consistent with the translatory flow mechanism discussed earlier, but could also be a function of a transmissivity feedback, or some other near-surface flow processes, such as pipe flow.

4.3.2. Peat response to climate change may alter alpine wetland water table dynamics

Peat was found at both Bowl and Valley, where peat is defined as soil containing approximately 30% soil organic matter, or 17% organic carbon by mass (SCWG, 1998). Valley, however, was the only wetland to have deep enough peat deposits, > 40 cm, to be considered a peatland (NWWG, 1997). Peatlands are, generally, lumped into three categories based on their physical hydrology, hydrochemistry, and vegetation. The water table, water chemistry, and vegetation at Valley are all suggestive of a moderate-rich fen, which are considered to be groundwater-dependent peatlands (Zoltai and Vitt, 1995; NWWG, 1997). This study, then, is the first to describe the hydrology of an alpine peatland in Canada, which is remarkable considering the volume and quality of peatland research that has occurred in the country. However, considering the limited attention alpine wetlands have received, it remains unclear as to how transferable the results of this study are to other alpine wetlands in the Rocky Mountains.

Aside from the utility of the presence of peat in helping to understand a wetlands' hydrology, peat is also an important hydrologic control. For example, the porosity of peat, which is usually above 0.8 (Letts *et al.*, 2000), is generally greater than that associated with mineral soils, allowing peat and peatlands to store more water than their mineral wetland counterparts. The high porosity acts as a buffer to water loss. An association between peat amount and porosity was observed in this study. For example, porosities of soils at Valley were highest (which had the most peat), while those at Slope Break were lowest (which had no peat). The implications of an association between peat and mineral soil storage could mean that alpine peatlands with greater peat, or SOM in general, may be more resilient to changes in climate, due to greater internal hydrological buffering capacities. The factors leading to differential peat accumulation in wetlands close to one another in the alpine remains an open question.

Another important feature of peat, in the context of climate change resilience, is that it is fairly responsive to hydrological and ecological processes (Arnold *et al.*, 2014; Kettridge *et al.*, 2016). Some of these responses are known to promote stabilization of wetland water table dynamics, while others may amplify (reinforce) those dynamics (Waddington *et al.*, 2015). Mechanism that reduce fluctuations of a given hydrological process are considered stabilizing (negative) feedbacks (Chapin *et al.*, 2009). For example, the "peat decomposition feedback" described by Waddington *et al.* (2015) is a stabilizing feedback that occurs when water depths increase, causing decomposition rates to increase, promoting pore collapse, which reduces specific yield values. The net result of such a feedback is that reduced water inputs are required to produce the same water table depths prior to drying and soil consolidation. Amplifying (positive) feedbacks, on the other hand, result in increased process fluctuation rates (Chapin *et al.*, 2009). These feedbacks are largely associated with afforestation and shrubification, but are also associated with specific yield responses, in certain circumstances. See Waddington *et al.* (2015) for a more detailed explanation of these and other feedbacks associated with peatlands.

This combination of feedbacks is important to consider as they suggest that peat-bearing wetlands are complex adaptive systems (*sensu* Levin, 1998; Chapin *et al.*, 2009; Petraitis, 2013) capable of some self-regulation, but are also highly responsive to internal conditions and external forcings, which may produce alternative stable states (Belyea and Baird, 2006; Dise, 2009). While there is some differing opinion as to whether state shifts in peatlands are real, or simply the result of mathematical simplifications (van Nes and Scheffer, 2005; Baird *et al.*, 2012), there are an increasing number of both theoretical (Hilbert *et al.*, 2000; Belyea and Baird, 2006; Ridolfi *et al.*, 2006; Rennermalm *et al.*, 2010; Baird *et al.*, 2012; Moffett *et al.*, 2015) and empirical studies (Srivastava and Jefferies, 1996; Jefferies *et al.*, 2006; Heffernan, 2008; Dise, 2009; Ireland *et al.*, 2012) that have illustrated alternative stable states occur in peatlands, as well as mineral wetlands.

From a management perspective, the existence of alternative stable states in ecosystems can be particularly challenging, because these ecosystems will often exhibit highly non-linear responses to change, which are very difficult to predict, and even harder to remediate (Chapin *et al.*, 2009). For example, it is conceivable that alpine wetland water table depths could express only marginal changes in response to shifts in climate, due to stabilizing feedbacks. However, drought conditions could promote the expression of an amplifying feedback, which causes a permanent drop in the water table, even if pre-drought conditions return. Similar 'tipping points' have been increasingly identified in a number of ecosystems (Scheffer *et al.*, 2012), including socio-hydrological systems (Sivapalan *et al.*, 2012).

In this respect, shrubification and afforestation may be a particular threat to alpine peatlands, because of the amplifying feedbacks associated with those mechanisms (Waddington *et al.*, 2015). There are an increasing number of examples and studies illustrating the invasion of shrubs and trees in the alpine (Grabherr *et al.*, 2010), including in wetlands (Moradi *et al.*, 2012). Such invasions could result in a non-linear lowering of water tables, which would impact peat accumulation rates. However, much of what is known about afforestation and shrubification feedbacks is from the boreal peatland literature. Study of alpine wetlands is required to determine if these types of climate changes are likely to yield similar water table results.

Last, it is unclear how quickly peat and high organic soils might take to respond to changes in hydrology. Kettridge *et al.* (2016), for example, found that low elevation peats respond to climate signals operating at centennial to millennial timescales. Arnold *et al.* (2014, 2015), however, found that drought promoted peat hydraulic changes within a matter of years to decades in high elevation wetlands of the Sierra Madres, California, USA. Future research focusing on the properties and processes contributing to these differences in peat response will be very valuable in understanding how alpine peatlands will respond to changes in hydrology and climate.

CHAPTER 5: CONCLUSIONS

The overall goal of this thesis was to better understand the hydrological processes contributing to alpine wetlands in the Helen Creek Research Basin, so that we might start to understand the vulnerability of these wetlands to environmental shifts associated with changing climate conditions. This was achieved by monitoring and characterizing water table dynamics of wetlands in differing geomorphic settings, and using a number of complementary hydrochemical and hydrophysical methods to infer the hydrological processes critical to maintaining near-surface water table stability.

Groundwater was found to be a potentially important source water to at least two of the alpine wetlands studied in the Helen Creek Research Basin, suggesting the bog model of alpine wetland hydrology has only limited utility. While the methods used to determine hydrological contributions were relatively simple, they were surprisingly consistent in indicating the importance of groundwater. This finding is important in understanding how alpine wetlands may respond to changes in climate, as groundwater is generally considered a more stable water source than other inputs, such as precipitation. Thus, having groundwater as the predominate water source may help alpine wetlands adapt to climate-driven hydrological changes. While this assessment is contrary to previous reviews of alpine wetland sensitivity to climate change (Winter, 2000; Williamson *et al.*, 2008), it is consistent with the growing number of studies that show groundwater is an important hydrological element of the alpine zone (e.g., Hood and Hayashi, 2015; Williams *et al.*, 2016).

Surface topography did not correlate well with the observed water table dynamics. This was unexpected, considering the emphasis that both the hydrological and wetland literature put on this landscape factor, and the complex topography that characterizes alpine environments. However, the observations made in this thesis are consistent with at least one other mountain wetland study that have found topography to be of limited utility in understanding wetland hydrology (i.e., Aldous *et al.*, 2015). Instead, geologic considerations may be more fundamental in determining hydrological conditions, especially in relation to groundwater flows.

Surprisingly, we found peat in two of the three studied wetlands in the Helen Creek Research Basin – with one of the wetlands likely qualifying as a peatland. This is significant in considering the vulnerability of alpine wetlands to climate change because: 1) peat may act as an indicator of long-term water availability (more peat could mean more stable groundwater conditions); and 2) peat is subject to strong stabilizing feedbacks that may reduce sensitivities to changes in water availability. This also suggests a potential hierarchy in the vulnerability in alpine wetlands, with well-established peatlands being more resilient to changes in climate, compared to their mineral wetland counter-parts. However, strong amplifying feedbacks, like those associated with shrubification, may negate this resilience.

Though this thesis represents an important advance in our understanding of alpine wetland hydrological conditions, and vulnerability to changes in those conditions, it is not without its limitations. For example, the focus on three wetlands in a single watershed limits the ability to generalize the findings to other alpine watersheds. Also, the hydrophysical and hydrochemical methods used did not allow for a clear understanding of groundwater contributions to the wetlands from shallow versus deep groundwater flow paths. In the case of the water balance method, groundwater was treated as a residual. Although logistically challenging, it would be useful to have had a continuous record of discharge and recharge at the site to independently evaluate groundwater inputs to and outputs from the wetlands. Also, both the water budget and water table estimates of hydrological processes would have been improved if lab methods for estimating soil water retention curve data were better matched with conditions in the field, thus improving estimates of capillarity. Additionally, compared to the 1D model used in this study, a distributed (2D or 3D) water table model may have yielded more insights into the differential roles that soils and water balance components (e.g., evapotranspiration) played in regulating water table dynamics. It is also unclear if the groundwater end-member for the hydrochemical methods was representative of the larger groundwater systems. This is a typical constraint on end-member mixing models (Christophersen and Hooper, 1992), but considering the consistency across methods, the results seem reasonable.

There are a number of elements that future research could focus on to address the above short-comings and advance our understanding of alpine wetland hydrology and resilience to shifts in climate. One such element is simply a better understanding of the distribution of alpine wetlands (and mountain wetlands, more generally). With few exceptions, even basic distributional information related to alpine wetlands is not available for most of the world. Such inventories would improve basic understanding of how common alpine wetlands are. Complementary efforts to characterize soil, hydrological, and landscape features (e.g., geologic and topographic features) of alpine wetlands would also be useful and could provide some preliminary insights in the controls of their distribution and resilience.

Additional understanding of alpine wetland function would also be useful, especially in the context of hydrology. While there is some documentation of alpine wetland functions and ecosystem services, there is still much to know. For example, how much water is stored in alpine wetlands? Are alpine wetlands important nodes in the larger hydrological network? To answer these questions, it would be useful to virtually experiment with wetland coverage and hydrological condition in alpine hydrological models.

Improved insights into connectivity between upland and wetland areas will also be valuable in better understanding alpine wetland resilience. This includes more refined approaches to partitioning 'shallow' and 'deep' flow paths, estimations of the size of the groundwater system(s) contributing to wetlands, and the transit times associated with incoming flows. Such information would be useful in understanding how well buffered alpine wetlands might to be to changes in snowpack.

A focus on understanding ecohydrological feedbacks in alpine wetlands will also be very valuable. While feedbacks have been relatively well studied in lowland areas, the understanding of alpine processes is still extremely limited, making prediction very difficult. In this regard, the work done by Waddington *et al*. (2015) could serve as a template for determining which feedback mechanisms require the most immediate focus. Of particular importance will be mechanisms that have relatively strong amplifying impacts, such as those associated with shrubification and afforestation, and to a lesser extent, specific yield. Together, such insights would be extremely valuable in understanding how wetland resilience outcomes might influence the larger hydrological network, thus improving predictive capacity, and allowing scientists and managers to better plan for an uncertain water future.

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APPENDIX A : DEPTH-SWE RELATIONSHIP

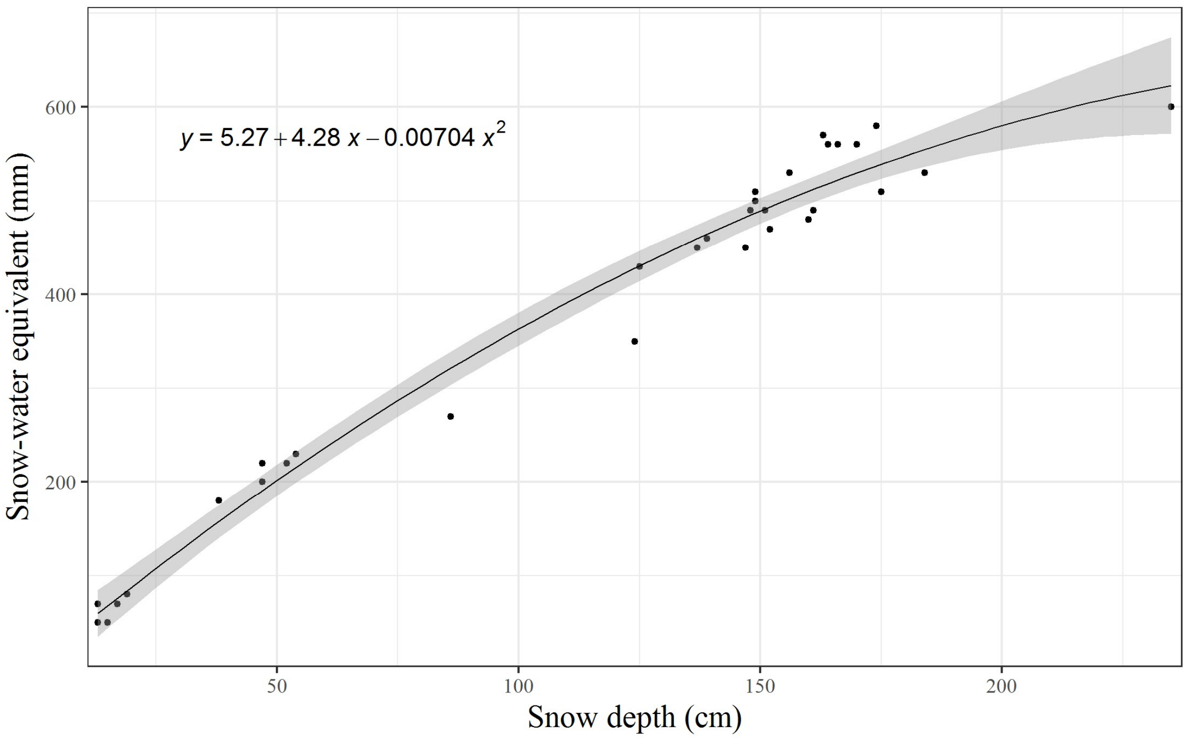

Figure A.1. Plot of snow depth (x) and snow water equivalent (y), including the 95% confidence interval (shaded area) of their relationship.

APPENDIX B : RELATIONSHIP BETWEEN WETLAND AND UPLAND PET

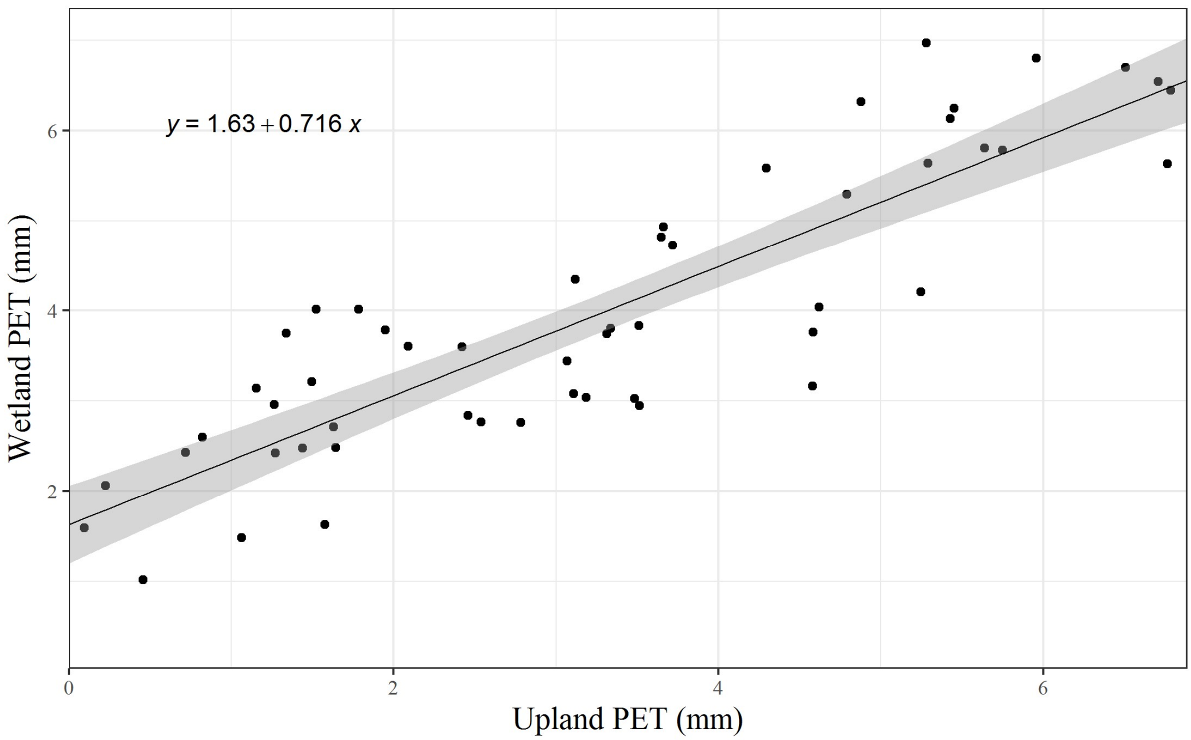

Figure B.1. Plot of upland (x) and wetland (y) potential evapotranspiration, including the 95% confidence interval (shaded area) of the relationship.

APPENDIX C : STAGE-DISCHARGE CURVES

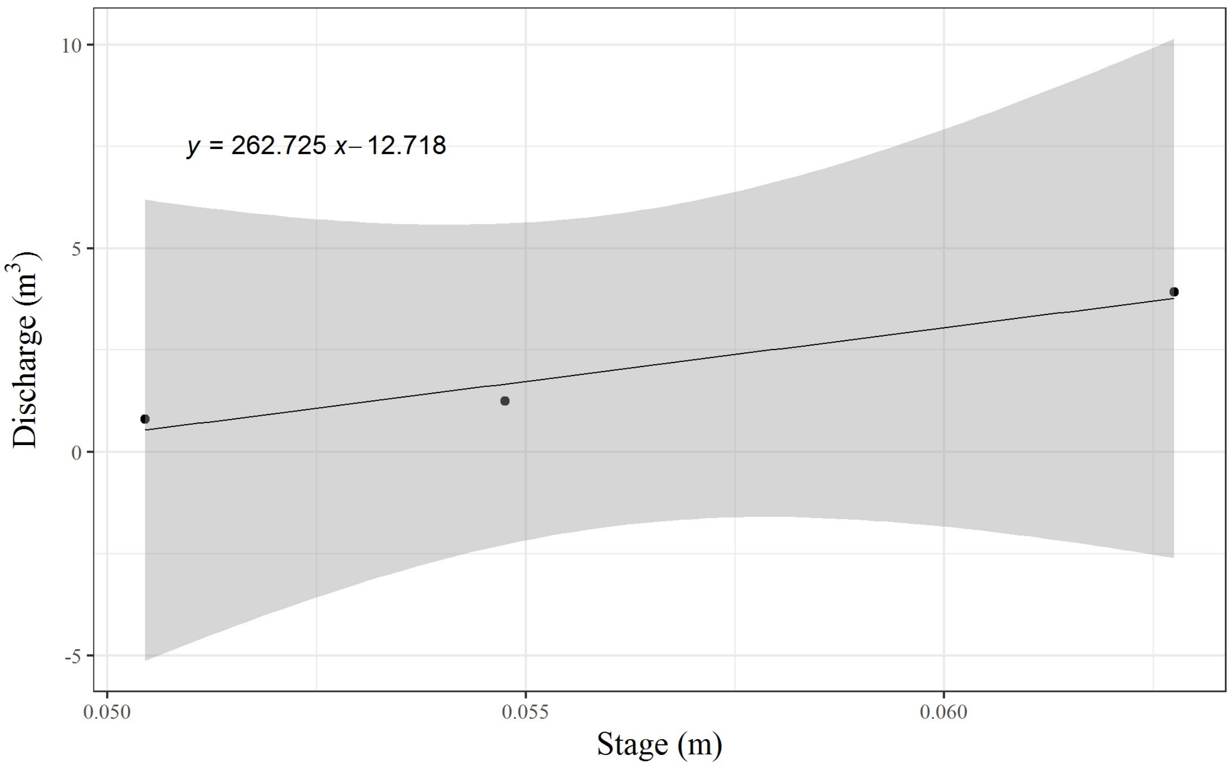

Figure C.1. Rating curve for the outlet of the Bowl wetland. The shaded region represents the 95% confidence interval. Note: Negative discharge values were not considered in any part of this study (i.e., the 1D water table and water budget models).

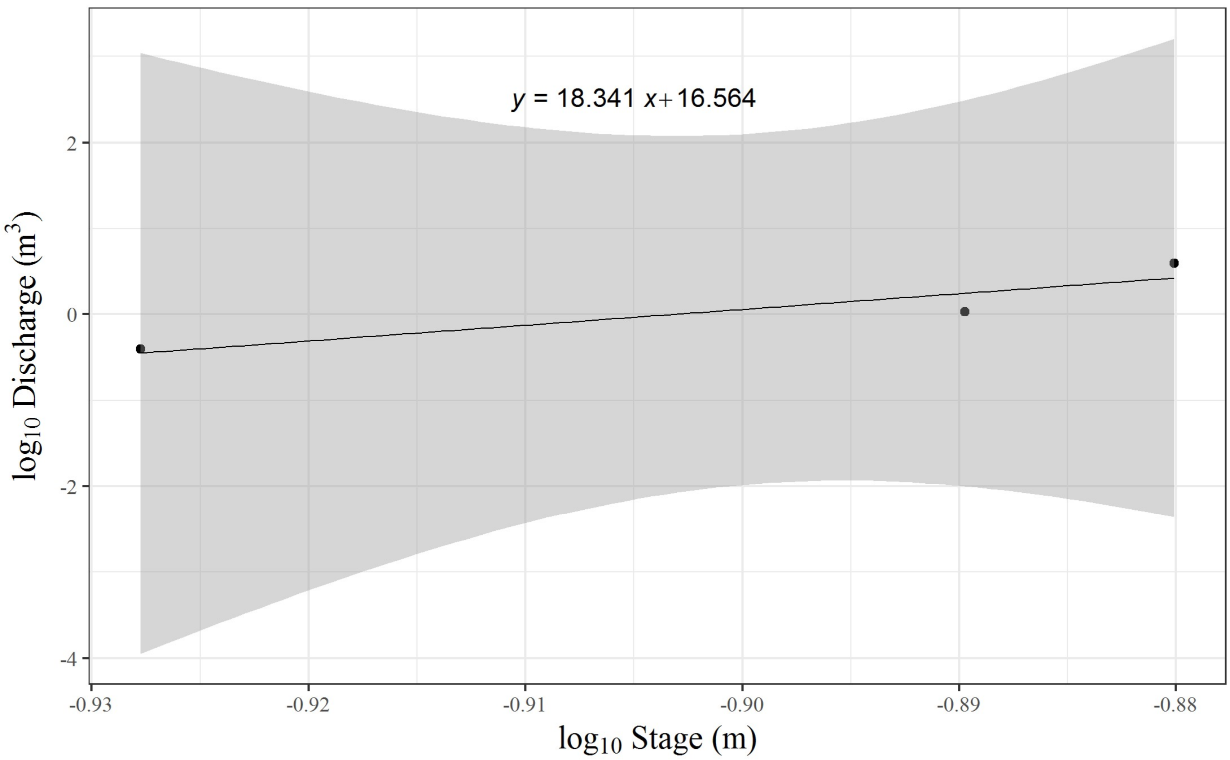

Figure C.2. Rating curve for the inlet of the Valley wetland. The shaded region represents the 95% confidence interval.

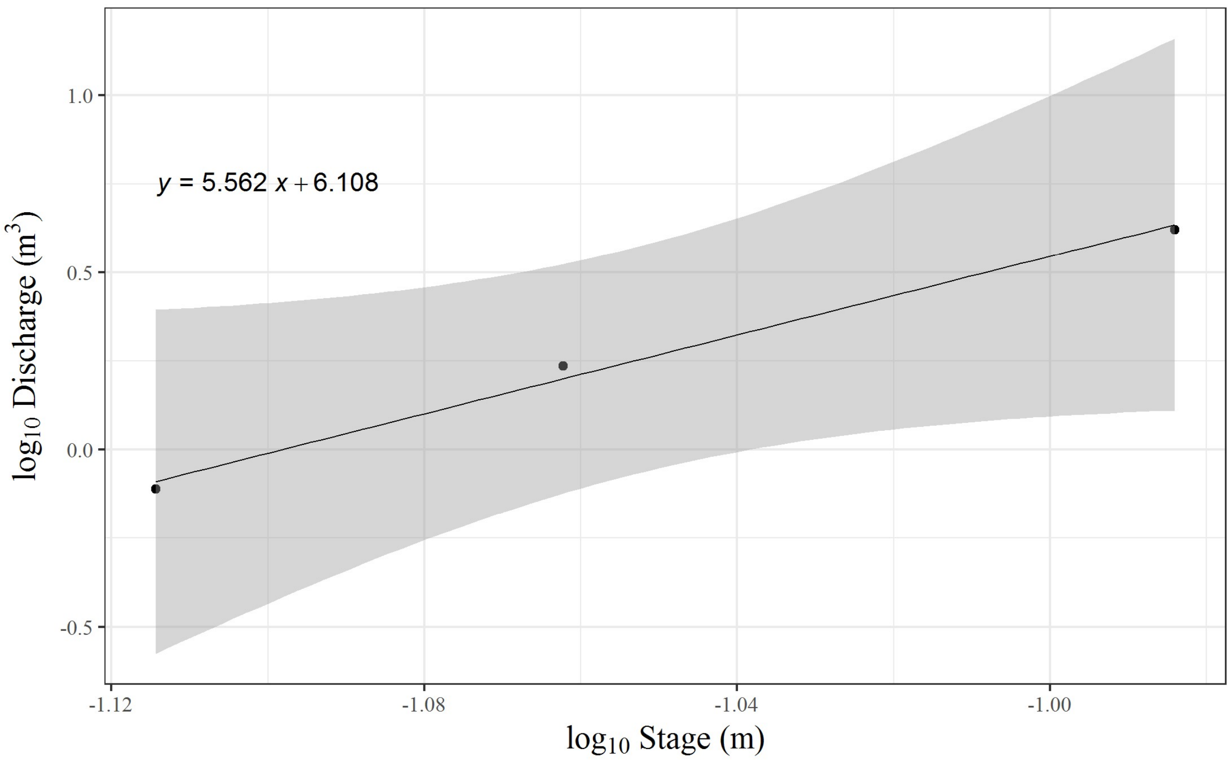

Figure C.3. Rating curve for the outlet of the Valley wetland. The shaded region represents the 95% confidence interval.