RUNOFF GENERATION OVER SEASONALLY-FROZEN GROUND: TRENDS, PATTERNS, AND PROCESSES

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ABSTRACT

Understanding and modeling runoff generation over seasonally-frozen hillslopes is a major challenge in hydrology. On the Canadian Prairies, snowmelt drives up to 80% of annual runoff, but the hydrological regime is vulnerable to changing precipitation states, snowpack persistence, snowmelt timing and rates, and frozen ground states. Our ability to understand and predict water partitioning and availability is being challenged by a lack of hillslope-scale climate-runoff observations, the presence of multiple interacting controls, and occurrence of spatial and temporal nonlinearity in runoff responses. I undertook long-term analyses of a 52-year dataset (1962-2013) of climate, snow cover, soil water content, and runoff from three 5 ha hillslopes in Saskatchewan. The aim was to determine how recent changes in climate have impacted upon hillslope rainfall- and snowmelt-runoff, and to unscramble the hierarchy of controls on hillslope snowmelt-runoff generation. These analyses then provided a multi-decadal contextual backdrop to an intensive field campaign that I led during the 2014 snowmelt season. I measured the spatial patterns of controls on runoff to assess the mechanisms behind connectivity and threshold delivery of snowmelt over frozen ground. There are three main conclusions from this research. First, differences between frozen and unfrozen soil infiltrabilities caused contrasting long-term snowmelt- and rainfall-runoff trends: no statistically significant changes were observed for rainfall-runoff amounts, but snowmelt-runoff showed statistically significant decreases over the 52-year record. Second, snowmelt-runoff was driven by hierarchical and condition-dependent controls related to snowfall, snow cover, antecedent soil moisture, and melt season dynamics. Third, for an individual melt season, filling and spilling of micro- and meso-depressions by snowmelt over frozen ground was the driver of hillslope connectivity and runoff delivery. Through a coupled analysis of trends, hierarchies and patterns, this research has advanced our understanding of runoff generation over seasonally-frozen ground. The long-term decrease in spring soil water recharge and snowmelt-runoff is a threat to dryland crop production and economic prosperity in farming. These findings have implications for modeling these threats by guiding new empirical frameworks for lumped hillslope runoff based on what we found in our long terms analysis and identifying what micro- and meso-scale features are important to now include in our process-based distributed snowmelt models.

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LIST OF ABBREVIATIONS

CART	Classification and Regression Tree
DI	downslope index
FA	flow accumulation
ha	hectare
ScWE	snow cover water equivalent
SfWE	snow fall water equivalent
SWE	snow water equivalent
TWI	topographic wetness index
vwc	volumetric water content

CHAPTER 1

INTRODUCTION

1.1 Introduction

The economic prosperity of the Canadian Prairies, at the northern limit of the Great Plains of North America, is heavily dependent on water (Pomeroy *et al*., 2009). The Prairies have a cold, semi-arid climate (300-400 mm of precipitation each year, approximately one-third of which falls as snow), with seasonally-frozen ground, and low-angled, undulating topography (Pomeroy *et al.*, 2010). Agriculture is a key land use of the region. The majority of the region sees continuous snow cover and frozen soils during the four to five months of winter. Approximately 90% of the South Saskatchewan River's (the region's main watercourse) streamflow is generated remotely, in the Rocky Mountains (Martz *et al.*, 2007). Snowmelt accounts for approximately 80% of the local runoff generation.

The hydrology of the Canadian Prairies is vulnerable to changing precipitation states, snowpack persistence, snowmelt rates, and frozen ground states (Fang and Pomeroy, 2007; Tetzlaff *et al.*, 2013). Projected warming and wetting over the coming decades in this region (Barrow, 2009) have the potential to yield cascading influences on hillslope hydrological regimes, runoff generation, soil water availability, agricultural productivity, and downstream water resources.

However, climate impacts on the hydrology of prairie hillslopes are poorly understood. Hillslopes are key landscape units because they are the scale at which we observe runoff generation processes that deliver water to soil water recharge, groundwater recharge, and streamflow. The hydrology of hillslopes at the sub-catchment scale is most important for agriculture. For example, runoff from these hillslopes delivers water to dugouts (small excavated storage reservoirs), which are important sources of water for livestock and farm household use. At the larger scale, runoff from hillslopes provides water to streams and glacially-formed topographic depressions (*i.e.* prairie potholes or wetlands), which are important sources of water for wildlife habitat and ecosystem functioning. Hillslope surface runoff also effects downstream flooding and water quality. The relevance of runoff is highly variable across the Prairies. Minimising runoff and maximising infiltration is often desirable both for alleviating flood risks and for promoting soil water recharge for crop productivity. Therefore, hillslope-scale hydrology and runoff generation mechanisms are highly important for the region's surface and near-surface water availability. Yet, it is very difficult to relate existing catchment-scale observations of climate effects on hydrology (*e.g.* Dumanski *et al.*, 2015) back to the hillslope scale due to the influence of sloughs (water-filled depressions), riparian zones, and groundwater on catchment-scale streamflow signals. We have therefore little knowledge of how runoff generation processes or hillslope-scale water budgets have responded to recent decadal changes in temperature or precipitation, making it difficult to extrapolate to the potential effects of these future climatic changes.

The dominance of snowmelt for generating runoff in the region is due to the rapid release of water from snowpacks over frozen ground, which typically reduces the infiltration capacity of the soil and encourages surface runoff (Granger *et al.*, 1984; Fang *et al.*, 2007). Infiltration into frozen soil is therefore an important flux on the Prairies, which is difficult to predict due to the complex effects of coupled heat and mass transfer with phase changes (Kane and Stein, 1983; Zhao and Gray, 1999). The activation of hillslope runoff is spatially and temporally unstable. There are multiple interacting factors – including snow accumulation, distributed melt inputs, seasonally-frozen ground, ice lenses, land cover, topography, and variable pre-melt soil moistures – which combine to drive runoff responses that are non-uniquely related to precipitation (Fang *et al.*, 2007; DeBeer and Pomeroy, 2010; Ireson *et al.*, 2013). We do not fully understand how the multiple controls on runoff interact to drive the nonlinear relationship between snowfall and runoff. This is largely because our cold regions process understanding is based mostly upon short-term experiments and single-season runoff events, where nonlinearities and interactions between the various process controls are not observable. We need much longer records and associated analysis in order to witness hierarchies, combinations, and interactions of process factors, to reflect the differing states of catchment response, and to provide context in the form of event, seasonal, or annual variability.

Further, in striving to understand critical thresholds and feedbacks in runoff generation, connecting point-scale runoff generation across hillslopes is now seen as fundamental to field and modeling campaigns (Bracken and Croke, 2007; Bracken *et al.*, 2013). Nonlinear, threshold-like runoff responses occur when continuous flow fields are generated across a plot, hillslope, or catchment. These responses can be directly linked to specific geomorphic processes and controls (Phillips, 2003). Therefore, studies that have measured key features and processes at dense spatial resolutions have led to deeper mechanistic understanding of connectivity and water delivery (*e.g.* Tromp-van Meerveld and McDonnell, 2006). New concepts, such as the fill and spill mechanism, have emerged that describe nonlinearities, thresholds, and storage-mediation in internal catchment response to precipitation inputs (Spence, 2010; Ali *et al.*, 2012; McDonnell, 2013). These provide opportunities to enhance predictability, improve interpretation of historical evidence, and inform modelling and experimental designs (Phillips, 2003). We do not know how the spatial patterns of geomorphic features and processes drive runoff connectivity over frozen ground, and how these might be similar or different to mechanisms of connectivity in other environments, such as fill and spill. The next phase to improve our understanding of threshold runoff generation must integrate the leading edge of process- and field-based understanding (*i.e.* spatial patterns, connectivity, *etc*.), with the use and development of associated long-term datasets to test and quantify change under variable hydro-meteorology.

1.2 Research goals and thesis outline

The overall objective of my research was to mechanistically assess hillslope runoff on the Canadian Prairies within the context of long-term change. I developed and performed my work at a long-term research site at the Swift Current Research and Development Centre, in Swift Current, managed by Agriculture and Agri-food Canada. Specifically, the site is a set of three adjacent agriculturally-managed hillslopes, each around 5 ha in area, instrumented and researched since 1962. I organized my research into three sections – the three research chapters presented in this thesis – each of which sought to address one outstanding question with regards to hillslope runoff generation on the northern Great Plains:

1. How have recent changes in climate impacted upon hillslope rainfall- and snowmelt-runoff on the northern Great Plains?
2. What is the hierarchy of controls on hillslope snowmelt-runoff generation over frozen ground?

3. What controls connectivity and resultant threshold delivery of snowmelt over frozen ground?

I began with comprehensive analyses of a previously unpublished 52-year (1962-2013) hillslope-scale dataset of climate, soil water content, snow pack, and runoff data from the Swift Current hillslopes (Chapters 2 and 3). These analyses provided a multi-decadal contextual backdrop to an intensive field campaign measuring the spatial patterns of controls on runoff during the 2014 snowmelt season at the Swift Current hillslopes (Chapter 4).

Specifically, in Chapter 2, my main objective was to determine the multi-decadal trends in precipitation (both rainfall and snowfall) and the resultant runoff events. Many long-term climate records from the northern Great Plains show climate trends over the last 50-100 years over the region (Akinremi *et al.*, 1999; Cutforth *et al*., 1999; Mekis and Vincent, 2011; Shook and Pomeroy, 2012). However, combined long-term climate-runoff records from the region are much less common, and there have been relatively few analyses of such datasets (Dumanski *et al.*, 2015; Ehsanzadeh *et al.*, 2016). For example, Dumanski *et al.* (2015) showed much more amplified streamflow trends than the corresponding precipitation trends over a 40 year period, for a catchment highly impacted by land use changes. But as noted above, it is very difficult to scale down understanding from the catchment-scale to the hillslope-scale. Therefore, for understanding how long-term changes in precipitation have affected hillslope runoff generation and water availability, we need hillslope-scale climate-runoff data, which are rare.

In Chapter 2 I first sought to leverage the rich 52-year dataset collected at the Swift Current hillslopes to determine how hillslope-scale runoff has responded to changes in precipitation quantity, timing, and phase. Second, I sought to elucidate any seasonal (snowmelt- *vs.* rainfall-runoff) differences in runoff response to long-term trends in precipitation. Finally, I explored how (if at all) the hillslope-scale responses differed to already-published catchment-scale responses. This research rested upon time series analysis of the climatological and hydrological variables, and referred to existing knowledge on prairie hydrological processes to explain the differences observed across seasons and geographic scales. I concluded this research with an outlook for the future, considering the importance of snowmelt soil water recharge and growing season precipitation for agricultural productivity in this region. This study was submitted in August 2016 for potential publication in Journal of Hydrology and is currently under peer review. [Coles, A.E., McConkey, B.G., and McDonnell, J.J. (2017) Climate change impacts on hillslope runoff on the northern Great Plains, 1962-2013, *Journal of Hydrology*, in review].

In Chapter 3, I aimed to determine the key controls on snowmelt-driven runoff generation over the same 52-year data record at the Swift Current hillslopes. Current snowmelt-runoff process understanding is based typically on short-term experiments, single-season runoff events, or point-scale experiments. Infiltration into frozen soil, and factors affecting it, are known to be important (Fang *et al.*, 2007; Ireson *et al.*, 2013), and the research presented in Chapter 2 supported this with results of its long-term analysis. However, we still do not understand the hierarchies, interactions, and feedbacks between these controls, and their condition-dependent behaviour. Hierarchical understanding has been shown to be important for model development (Uchida *et al.*, 2005), spatial extrapolation (Cammeraat, 2002), and runoff classification schemes (Barthold and Woods, 2015). Such understanding would be hugely beneficial on the Canadian Prairies, where the highly nonlinear runoff response to snowfall and snowmelt makes it notoriously difficult to model (Gupta and Sorooshian, 1997; Pomeroy *et al.*, 2007).

In Chapter 3 I used decision tree learning (De’ath and Fabricius, 2000), a data mining approach, to extract information from the long-term dataset on the interactions between controls (*e.g.* topography, vegetation, land use, soil characteristics, and precipitation dynamics), their hierarchical order, and their condition-dependent importance. I compared the resultant decision tree model for the prediction of snowmelt-runoff ratio to an existing, widely-used model for infiltration into frozen ground. This study was submitted in October 2016 for potential publication in Hydrology and Earth System Sciences and is currently under peer review. It has been published in Hydrology and Earth System Sciences Discussions. [Coles, A.E., Appels, W.M., McConkey, B.G., and McDonnell, J.J. (2016) The hierarchy of controls on snowmelt-runoff generation over seasonally-frozen hillslopes, *Hydrology and Earth System Sciences Discussions*, doi: 10.5194/hess-2016-564].

Finally, in Chapter 4, I aimed to determine how the spatial patterns of the controls and processes identified in the long-term analyses (Chapters 2 and 3) interact to drive hillslope-wide connectivity and threshold-like runoff over frozen ground. Connectivity studies in other regions (*e.g.* Darboux *et al.*, 2002; Tromp-van Meerveld and McDonnell, 2006; Detty and McGuire, 2010) and at larger wetland-dominated scales (*e.g.* Leibowitz and Vining, 2003; Shaw *et al.*, 2012) have unveiled deeper process understanding and helped develop new, potentially-unifying concepts such as the fill and spill mechanism (*e.g.* Spence and Woo, 2003; Tromp-van Meerveld and McDonnell, 2006; McDonnell, 2013). However, runoff connectivity is still poorly understood on prairie hillslopes, where long periods of snow-covered frozen ground with very shallow slopes mask spatial patterns.

In Chapter 4, I sought to establish, for the 2014 snowmelt season, the spatial patterns that affect connectivity over frozen ground, and whether or not these are consistent with the fill and spill mechanism. I undertook digital topographic analysis of Hillslope 2 of the Swift Current hillslopes to develop a working hypothesis of flowpath locations and the extent of downslope impedance of runoff. In the field, I measured soil water content, thawed layer depth, snow cover, and snow water equivalent at a high resolution over Hillslope 2, through the melt season. I combined these measurements with measurements of snow, soil water, ponded water, and hillslope runoff stable isotope composition. I evaluated the importance of soil moisture, topography, and the fill and spill mechanism for runoff connectivity over frozen prairie hillslopes. This study is due for submission for potential publication in Hydrological Processes. [Coles, A.E. and McDonnell, J.J. (2017) Fill and spill drives runoff connectivity over frozen ground, *Hydrological Processes*, for submission].

This thesis adopts a ‘dissertation by manuscript’ style. Following this introductory chapter, Chapters 2, 3, and 4 are structured into three manuscripts. Finally, Chapter 5 sets out the conclusions of my research, discusses some linkages between the three manuscript chapters, and suggests avenues for future research.

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CHAPTER 2

CLIMATE CHANGE IMPACTS ON HILLSLOPE RUNOFF ON THE NORTHERN GREAT PLAINS, 1962-2013

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2.1 Abstract

On the Great Plains of North America, water resources are being threatened by climatic shifts. However, a lack of hillslope-scale climate-runoff observations is limiting our ability to understand these impacts. Here, we present a 52-year (1962-2013) dataset (precipitation, temperature, snow cover, soil water content, and runoff) from three 5 ha hillslopes on the seasonally-frozen northern Great Plains. In this region, snowmelt-runoff drives *c.* 80% of annual runoff and is potentially vulnerable to warming temperatures and changes in precipitation amount and phase. We assessed trends in these climatological and hydrological variables using time series analysis. We found that spring snowmelt-runoff has decreased in response to a reduction in winter snowfall, but that rainfall-runoff has shown no response to increases in rainfall or shifts to more multi-day rain events. In summer, unfrozen, deep, high-infiltrability soils act as a kind of sponge to rainfall, buffering the long-term runoff response to rainfall. Meanwhile, during winter and spring freshet, frozen ground limits infiltration and results in runoff responses that more closely mirror the snowfall and snowmelt trends. These findings are counter to climate-runoff relationships observed at the catchment scale on the northern Great Plains where land drainage alterations dominate. At the hillslope scale, decreasing snowfall and spring soil water content is causing agricultural productivity to be increasingly dependent on growing season precipitation, and will likely accentuate the impact of meteorological droughts.

2.2 Introduction

Climate impacts on the hydrology of the Great Plains of North America are poorly understood. Any such impact may have enormous consequences for agriculture on the Great Plains, where 80% of the region is under agricultural management, has a crop market value of approximately $92 billion USD (Hatfield *et al.*, 2014), and accounts for about half of the world's wheat production (Wishart, 2004). On the northern Great Plains, the focus of this study, water regimes are being threatened by warming temperatures and changes in precipitation amount and phase. For future sustainable agricultural production, it is crucial to understand the long-term climate-induced shifts in water availability. To do this, we need long-term records of climate and runoff.

While many long-term climate records exist on the Great Plains, there are relatively few sites with long-term combined climate-runoff records for this region (Figure 2.1). Most of these are in the southern Great Plains (Garbrecht, 2008; Harmel *et al.*, 2006; Heppner and Loague, 2008; Wine and Zou, 2012). The only published analyses of long-term climate-runoff records pertaining to the seasonally-frozen northern Great Plains are 40-year datasets from agriculture- and wetland-dominated catchments on the Prairies of Canada (Dumanski *et al.*, 2015; Ehsanzadeh *et al.*, 2016). All are catchment-scale streamflow observations. On the southern Great Plains, Harmel *et al.* (2006) and Wine and Zou (2012) found statistically significant trends in precipitation, but no resultant shifts in streamflow. Meanwhile, Garbrecht (2008) showed nonlinearly greater streamflow trends compared to the precipitation trends. On the northern Great Plains, Ehsanzadeh *et al.* (2016) showed little significant annual climate-runoff changes, with some slight wet and dry regime changes, while Dumanski *et al.* (2015) found, for the Smith Creek Research Basin in southeast Saskatchewan, much more amplified seasonal streamflow trends than the corresponding precipitation trends. (Dumanski *et al.* (2015) analysed snowmelt- and rainfall-driven events separately, while Ehsanzadeh *et al.* (2016) conducted an annual precipitation and streamflow analysis.) Most of the catchment-scale studies on the Great Plains, like many catchment-scale studies in other regions (Woo *et al.*, 2006), demonstrate that catchments can act as nonlinear filters of climatic signals to either possibly damp or enhance the resultant runoff signal.

Figure 2.1 The Great Plains of North America, indicating the locations of study sites with existing climate-runoff datasets. 1: A 69-year dataset from the USDA-ARS Grassland Soil and Water Research Laboratory experimental watershed in the Texas Blacklands Prairies near Riesel, Texas, USA (Harmel *et al.*, 2006). 2: A 65-year dataset from crop- and pasture-land of the Fort Cobb Reservoir watershed in central Oklahoma, USA (Garbrecht, 2008). 3: An 8-year dataset from the R-5 rangeland catchment in the USDA-ARS Washita River Experimental Watershed in central Oklahoma, USA (Heppner and Loague, 2008). 4: A *c.* 54-year dataset from Council Creek watershed in the tallgrass prairie of north-central Oklahoma, USA (Wine and Zou, 2012). 5: A 40-year dataset from an agriculture- and wetland-dominated catchment, Smith Creek Research Basin, on the prairies of Canada (Dumanski *et al.*, 2015). 6: This paper's study site, the Swift Current hillslopes, at South Farm, Swift Current, SK, Canada with a 52-year dataset.

The hydrology of the uplands at the sub-catchment-scale is most important for agriculture. For instance, dugouts (small excavated storage reservoirs), which collect water from adjacent hillslopes, are an important source of water for livestock watering and farm household use on the Canadian Prairies. These dugouts are purposefully not located on or within significant watercourses, so all inflow is determined from local hillslope hydrology. The hillslope scale is also the scale at which we observe runoff generation processes that ultimately deliver water to soil water recharge, groundwater recharge, and streamflow. To date, there have been no published long-term climate-runoff observations at the hillslope scale on the Great Plains. Further, it is very difficult to relate catchment-scale observations back to hillslope-scale water trends and resources when sloughs (water-filled depressions), riparian zones with possible groundwater contribution, and other geomorphic zones in the landscape influence the catchment-scale integrated streamflow signal (McGuire and McDonnell, 2010). As a result, we do not know how, if at all, runoff generation processes and hillslope-scale water availability have responded to changes in, for example, temperature and precipitation, and whether or not hillslope-scale runoff generation is coupled or decoupled from climate variations. Therefore, for understanding water availability for dryland agriculture in this region, we need observations of hillslope-scale runoff.

In the seasonally-frozen northern Great Plains, snowmelt in the spring freshet drives *c.* 80% of the annual runoff and thus dictates much of the surface water and soil water availability (Fang *et al.*, 2007). Snowmelt-runoff is generated typically as infiltration-excess overland flow (Granger *et al.*, 1984) when rapid release of water from the snowpacks, usually in a short, one to three week long snowmelt season, occurs over frozen ground of limited infiltration capacity on low relief slopes. However, as cold regions lose their cold, snowpack persistence, frozen ground, and snowmelt rates (important controls on the amount of spring runoff) are particularly vulnerable to warming and shifts in precipitation phase (Tetzlaff *et al.*, 2013).

Decreased winter snowfall has been observed on the northern Great Plains (*e.g.* Akinremi *et al.*, 1999; Cutforth *et al*., 1999; Mekis and Vincent, 2011), as has increased spring and fall rainfall fractions (*e.g.* Mekis and Vincent, 2011; Shook and Pomeroy, 2012). One might hypothesize that climate-related changes will yield cascading effects on hydrological regimes, runoff generation, and ultimately water resources available for agriculture and other uses. In the summer months, hillslope-runoff occurs occasionally during intense, one-day convective rainstorms that may generate infiltration-excess overland flow. But recent observations show decreasing one-day rain events, and an increase in less-intense, multi-day frontal rain events with greater overall magnitude (Shook and Pomeroy, 2012). As yet, for both snowmelt- and rainfall-driven runoff events, the effects of these precipitation trends on hillslope-scale runoff generation and water availability are unknown.

Here, we use a 52-year hillslope-scale dataset of climate and runoff data from three 5 ha agricultural hillslopes on the northern Great Plains to quantify changes in precipitation and identify if/how they relate to changes in runoff and water availability. Specifically, we ask the following questions:

i) How have hillslope-scale snowmelt- and rainfall-runoff events responded to changes in precipitation quantity, timing, and phase?
ii) Do hillslope-scale snowmelt- and rainfall-runoff responses differ in their response to long-term trends in precipitation?

The dataset we present here offers a unique and unprecedented ability to answer these questions, in order to understand if and why runoff responses and hillslope-scale water availability are shifting in response to changes in precipitation amount, phase, and timing.

2.3 Study site

The study site (South Farm, Swift Current Research and Development Centre, Agriculture and Agri-Food Canada, Swift Current, Saskatchewan, Canada; 50°15'53"N 107°43'53"W; hereafter referred to as the Swift Current hillslopes) is situated in the Brown Soil Zone on the northern Great Plains of North America (Figure 2.1). The northern Great Plains hillslopes are characterized generally by low relief and deep, well-drained soils of high unfrozen infiltration capacity (Elliott and Efetha, 1999). The northern Great Plains' greatest source of water for agriculture comes from surface and near-surface sources. In the South Saskatchewan River Basin of the northern Great Plains, agriculture dominates the share of surface water extraction (86.5%), and it is also reliant on shallow soil water storage (Pomeroy *et al.*, 2009). This is in contrast to the southern Great Plains, where groundwater is the primary source of water (Barnett *et al.*, 2005).

Specifically, the Swift Current hillslopes are a set of three adjacent 5 ha agricultural hillslopes with undulating topography and 1-4% north-facing slopes. Grassed berms around the perimeters of the hillslopes prevent runoff from transferring between hillslopes. The soil is a Swinton silt loam (Cessna *et al.*, 2013). The groundwater table is several meters below the soil surface (Maathuis and Simpson, 2007) and is not thought to contribute to runoff from the hillslopes. Coles and McDonnell (2017) (Chapter 4 of this thesis) used stable water isotope analysis to show that there is very little contribution of 'old' soil water to runoff from these hillslopes. The hillslopes are under an annual rotation of wheat (*Triticum aestivum*) and fallow, with some interspersions of grass (*Agropyron cristatum*) and pulses (lentils and peas; *Lens culinaris* and *Pisum sativum*, respectively). In addition, a nearby (*c.* 700 m to the south-southeast) Environment and Climate Change Canada standard meteorological station has recorded precipitation and temperature daily from 1886 to present and hourly from 1995 to present, as well as daily snow depth and wind speed (at 2 m and 10 m above the ground surface) from 1960 to present. During the 4-6 month winter season, the soils of the northern Great Plains are frozen from the soil surface to a depth of typically >1 m (Ireson *et al.*, 2013). At the Swift Current hillslopes, the ground freezes typically in late October, and begins to thaw in March during the snowmelt freshet, shown by soil temperature data from the meteorological station and observations on the hillslopes.

2.4 Dataset and methods

We use daily precipitation amounts and phase (rainfall or snowfall, where snow is given as snow water equivalent – SWE) from 1962-2013, measured at the Environment and Climate Change Canada meteorological station using a Belfort weighing gauge, where snowfall and rainfall were distinguished using air temperature data (measured inside a Stevenson Screen). From these data, we determined annual and seasonal totals of rainfall, as well as annual and seasonal occurrences, durations, and sizes of one-day and multi-day rain events. Each season was defined as follows: winter (December, January, February), spring (March, April, May), summer (June, July, August), and fall (September, October, November). These seasonal demarcations were used so that we could assess any changes in rainfall and rainfall-runoff regimes at different times of year (*e.g.* shortly after spring snowmelt *vs.* later in the year after the growing season). One-day rain events are defined as days with rainfall that are preceded and followed by days with no rainfall, while multi-day rain events are defined as two or more continuous rain days.

Snow cover for each hillslope was measured by manual snow surveys each year from 1965-2013. Snow depth and density were measured, and SWE calculated, at nine points on each hillslope, and means of each were calculated to give three hillslope-averages. These snow surveys were repeated several times from January to March, including one snow survey that was intended to capture the maximum snowpack before the onset of spring snowmelt. To explore the transformation of seasonal snowfall amounts into the amount of snow cover accumulated on the ground before the onset of spring snowmelt, we used temperature data and 10 m wind speed data (daily maximum, minimum and mean). From these data, we determined the occurrence of above-freezing winter days, and the likely occurrence of over-winter melt events and blowing snow ablation or sublimation.

Gravimetric soil water content (water fraction by volume of soil) was measured twice per year from 1971-2013 on each hillslope. In October (prior to freeze-up) and April (following spring snowmelt) each year, the gravimetric soil water content was measured for five depth intervals in the soil profile (0-15 cm, 15-30 cm, 30-60 cm, 60-90 cm, and 90-120 cm, where the soil water content was measured from a subsample of the entire mixed interval and reported for the mid-point of the interval: 7.5, 22.5, 45, 75, and 105 cm, respectively) on a permanent nine-point grid on each hillslope. These were converted to volumetric soil water contents using bulk density data. The bulk density of each depth interval in the soil profile was observed to be 1.22, 1.25, 1.36, 1.39, and 1.63 g cm^{-3}, respectively. The porosity of the soil at each depth interval in the soil profile was calculated to be 54.1%, 53.0%, 48.9% 47.7%, and 38.7%, respectively. Hillslope-averaged soil water content at each depth was calculated from the point-scale data. Both hillslope-averaged and point-scale data were recorded from 1980-2013. From 1971-1979 only hillslope-averaged data were recorded.

Runoff was measured from 1962-2013 using a Stevens water level chart recorder in the stilling well of a heated H-flume at the surface outflow of each hillslope. Rating curves for each flume were used to calculate daily runoff depths (mm d^{-1}). No runoff was measured during 1970. Flow rates exceeded the flume capacity during heavy rainfall on June 14 1964, and runoff during that event was estimated by McConkey *et al.* (1997). Rainfall-runoff events were identified as occurring when rainfall and runoff occurred on the same day. We calculated the runoff amounts derived from snowmelt and rainfall, distinguishing between one-day and multi-day rainfall-runoff amounts. No rain-on-snow runoff events were observed.

For all annual calculations and analyses, we used the hydrological year, October 1 – September 30. This demarcation is consistent with the hydrological regime of the region which sees snowfall and snow accumulation from October onwards, followed by snowmelt-runoff in the spring and rainfall through the summer. For all variables of interest, we used the Mann-Kendall test, a common statistical test used for the analysis of trends in climatological and hydrological time series (Burn and Hag Elnur, 2002). To assess the significance of these trends, we computed the p-value. We determined a trend to be significant if the p-value was below a significance level of 0.05. We used linear regression to determine the direction, gradient, and percentage change over time, of the trend.

2.5 Results

2.5.1 Precipitation

The long-term (1962-2013) average annual precipitation was 360 mm, of which 76% fell as rain and 24% as snow (Figure 2.2). Runoff is not continuous from these hillslopes: it occurs intermittently through the hydrological year, and is event-based (Figure 2.2). Over this long-term period, total annual precipitation increased by 90 mm (a 28% increase) (Figure 2.3a); total annual rainfall increased by 112 mm (a 53% increase) (Figure 2.3b); and total annual snowfall decreased

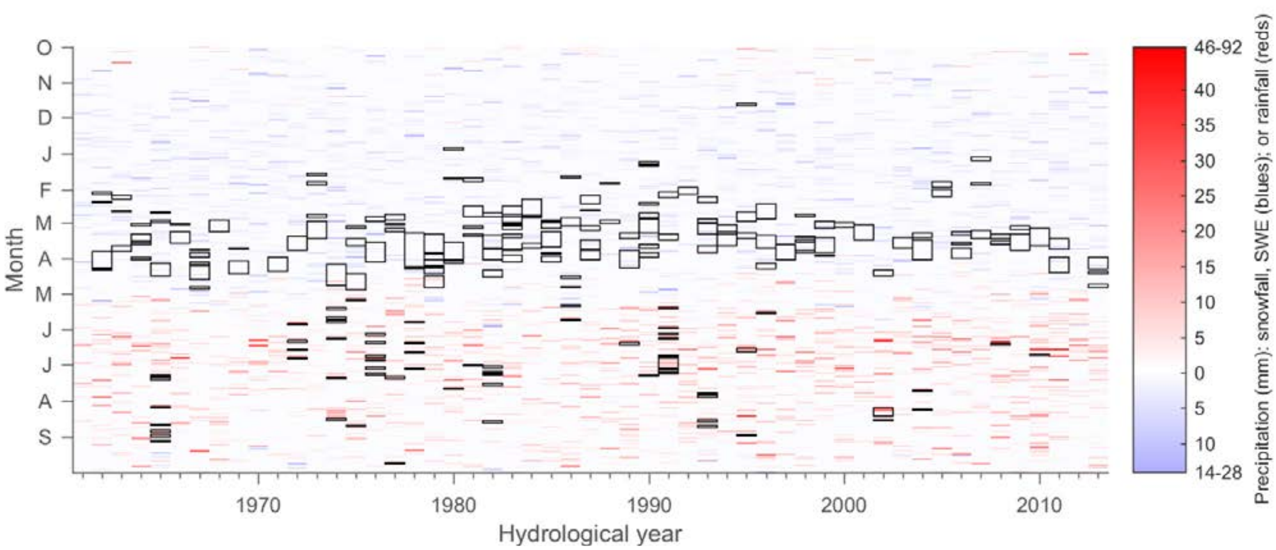

Figure 2.2 Daily precipitation and runoff at the Swift Current hillslopes, 1961-2013. Daily snowfall as SWE (blues) and rainfall (reds) at the site, with colour shade corresponding to daily volumes. Occurrences of snowmelt- and rainfall-runoff from the three hillslopes (combined) are indicated by black rectangles.

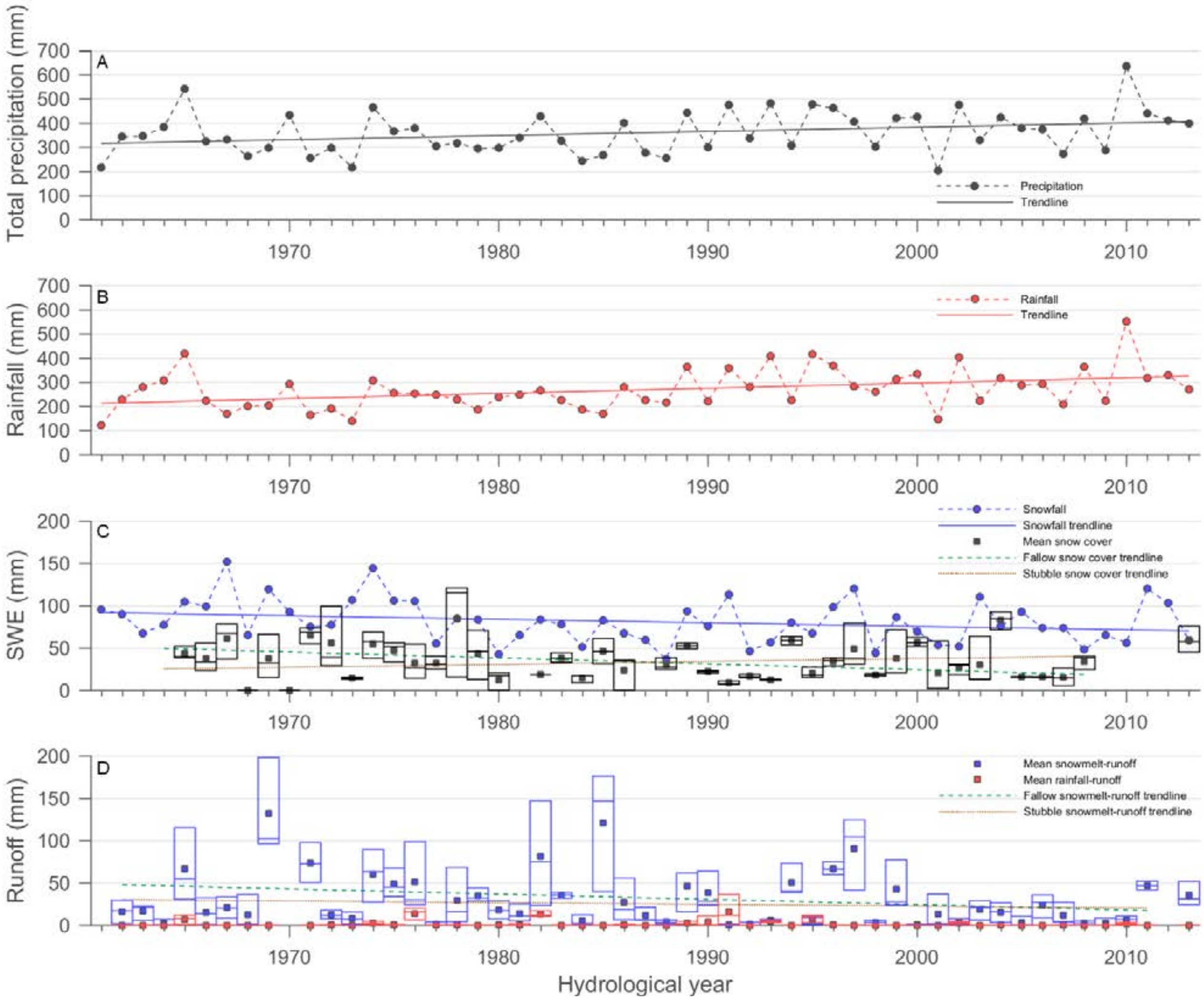

Figure 2.3 Annual (hydrological year) precipitation and runoff at the Swift Current hillslopes, 1961-2013. A) Total precipitation depths (rainfall and snowfall combined), with trendline. B) Rainfall depths, with trendline. C) Snowfall depths as SWE (blue circles), with trendline. Also shown is snow cover depth (as SWE) on the hillslopes. The boxes indicate the maximum, median and minimum hillslope snow cover SWE, with the mean seasonal snow cover amount indicated with black squares. Trendlines for snow cover depth under different land cover types are given. D) Annual snowmelt-runoff amounts (blue) and rainfall-runoff amounts (red). The boxes indicate the maximum, median and minimum runoff amounts. Trendlines for snowmelt-runoff amount under different land cover types are given.

by 22 mm (an 18% decrease) (Figure 2.3c). These latter two trends are due to shifts in precipitation phase and timing: more precipitation fell as rain and less as snow in winter and spring (no similar trend for fall) over the study period. For non-winter months over the period 1962-2013, we observed significant shifts in the delivery of rainfall from multi-day rain events (Figure 2.4a). The number of multi-day rain events, volume of rain that fell during each event, and proportion of

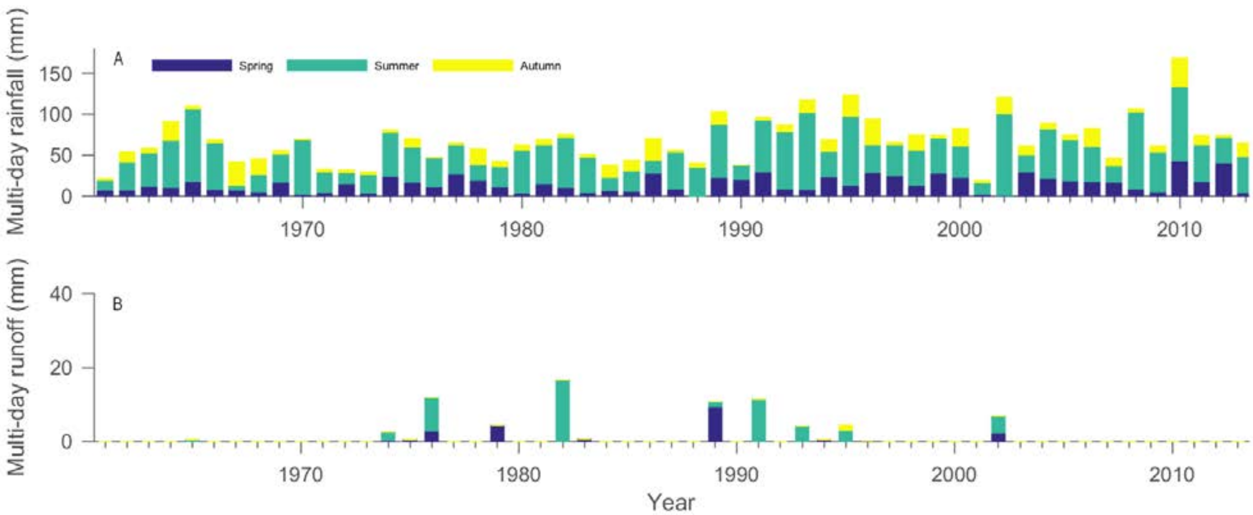

Figure 2.4 A) Total rainfall from multi-day rain events for spring (blue), summer (green), and autumn (yellow). Trendlines indicated for spring (dashed line), summer (dotted line), and autumn (solid line). B) Total runoff from those multi-day rain events, following the same seasonal colour scheme. No significant trends in runoff from multi-day rain events.

summer rainfall delivered by each event (as opposed to one-day storms), all increased ($p<0.05$). There were no equivalent trends in the delivery of rainfall from one-day rain events (Figure 2.5a).

2.5.2 Snow accumulation and melt

The SWE of the snow cover before spring snowmelt was, on average, 43 ± 31% of total snowfall (where ± here and throughout denotes the standard deviation) (Figure 2.3c; Figure 2.6). This suggests that, on average, 57 ± 31% of snowfall ablated through the winter via a combination of evaporation, sublimation, wind redistribution, and mid-winter melt and infiltration. Like snowfall, there was also a trend over the 1962-2013 period towards less snow cover (SWE; measured prior to spring snowmelt) retained on the hillslopes (Figure 2.3c). In years where the hillslopes were left fallow, snow cover SWE decreased by 88% over the 1962-2013 period (Figure 2.3c). The decrease in snow cover was over four times greater than the decrease in snowfall (a 21% decrease). No significant trends in snow cover were found over the same period when the fields were covered in stubble (Figure 2.3c).

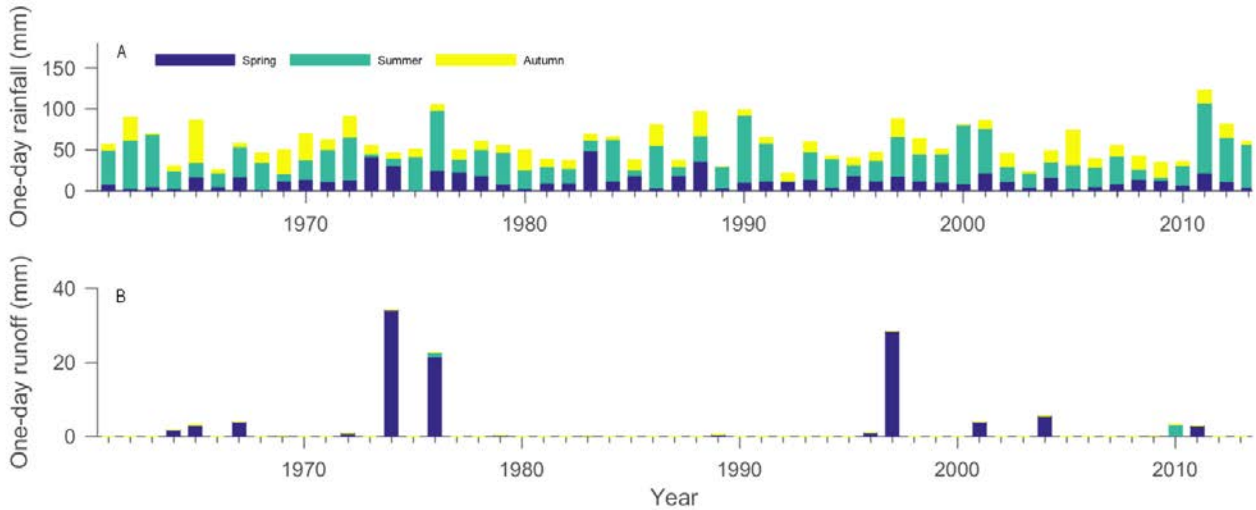

Figure 2.5 A) Total rainfall from one-day rain events for spring (blue), summer (green), and autumn (yellow). No significant trends in one-day rain events. B) Total runoff from those one-day rain events, following the same seasonal colour scheme. No significant trends in runoff from one-day rain events.

2.5.3 Soil water content

Over the period 1971-2013, mean volumetric soil water content measured in the spring was 0.22 at the surface (wettest depth) and 0.18 at 105 cm depth (driest and deepest measuring depth). The hillslopes were typically drier in the fall, when mean volumetric water content of the soil was 0.19 at the surface, and 0.17 at 105 cm depth. Between 1971 and 2013, the hillslope-averaged spring soil water content decreased for all hillslopes and at all depths; however, this trend was only significant for the surface where soil water content decreased by between 8.7% (Hillslope 1) and 9.5% (Hillslope 3) over the 43-year study period. There were no consistent trends in the equivalent, hillslope-averaged fall soil water content. Soil water content time series on Hillslope 3 (Figure 2.7), which had a consistent wheat-fallow rotation, is perhaps most reliable for climate-runoff analysis since any changes do not reflect the effects of land management. On Hillslope 3, between 1971 and 2013, hillslope-averaged spring soil water content showed a decreasing trend for all depths (although only significant at the soil surface), but there were no apparent changes in the fall. At the point scale, however, for which we have data from 1980-2013, there were significant trends at some points, depths, and hillslopes, for both spring and fall wetness conditions. Spring data showed decreasing soil water content at all depths, while fall data showed decreasing soil water content at the surface and increasing soil water content at the lowermost depths (75 cm and

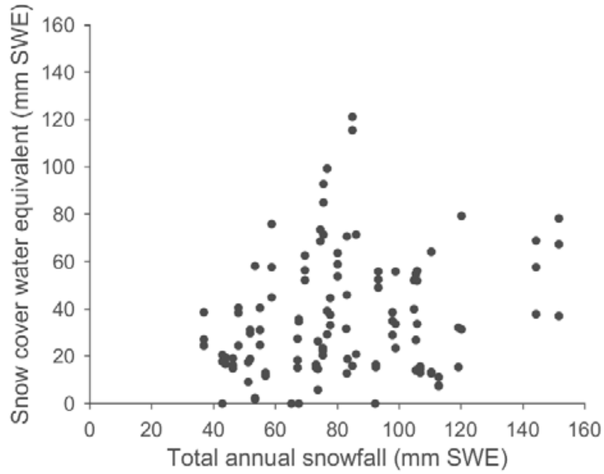

Figure 2.6 Total annual snowfall (mm) and snow cover water equivalent (mm) measured before the onset of spring snowmelt for each year and each hillslope over the period 1965-2013.

105 cm). We also examined the difference in soil water content from the fall to the spring (over winter and snowmelt), and from the spring to the following fall (over summer), for all hillslopes and at all depths over the period 1970-2011 (Figure 2.8). This showed a decreasing trend in the amount of soil water that was added to the soil profile over winter and during snowmelt, and an increasing trend in the amount of soil water that was added to the soil profile from rainfall in the summer.

2.5.4 Runoff

Over the period 1962-2013, mean snowmelt-derived spring runoff for each hillslope was 26 mm (Hillslope 1), 39 mm (Hillslope 2), and 22 mm (Hillslope 3). Over the same period, snowmelt-runoff decreased on each hillslope by 68%, 59%, and 51%, respectively (Figure 2.3d), although the trend was only significant on Hillslope 1. Snowmelt-runoff ratios also decreased over the period 1962-2013 (Figure 2.9), regardless if calculated based on total snowfall data or snow cover data. This implies that progressively less SWE was translated into runoff from the hillslopes with more going to a combination of infiltration, sublimation, evaporation, and blowing snow. We found no relationship between fall or spring rainfall fraction and the amount of snowmelt-runoff. We observed a significant relationship between SWE and the volume of snowmelt-runoff generated during the spring freshet, with both decreasing over the study period.

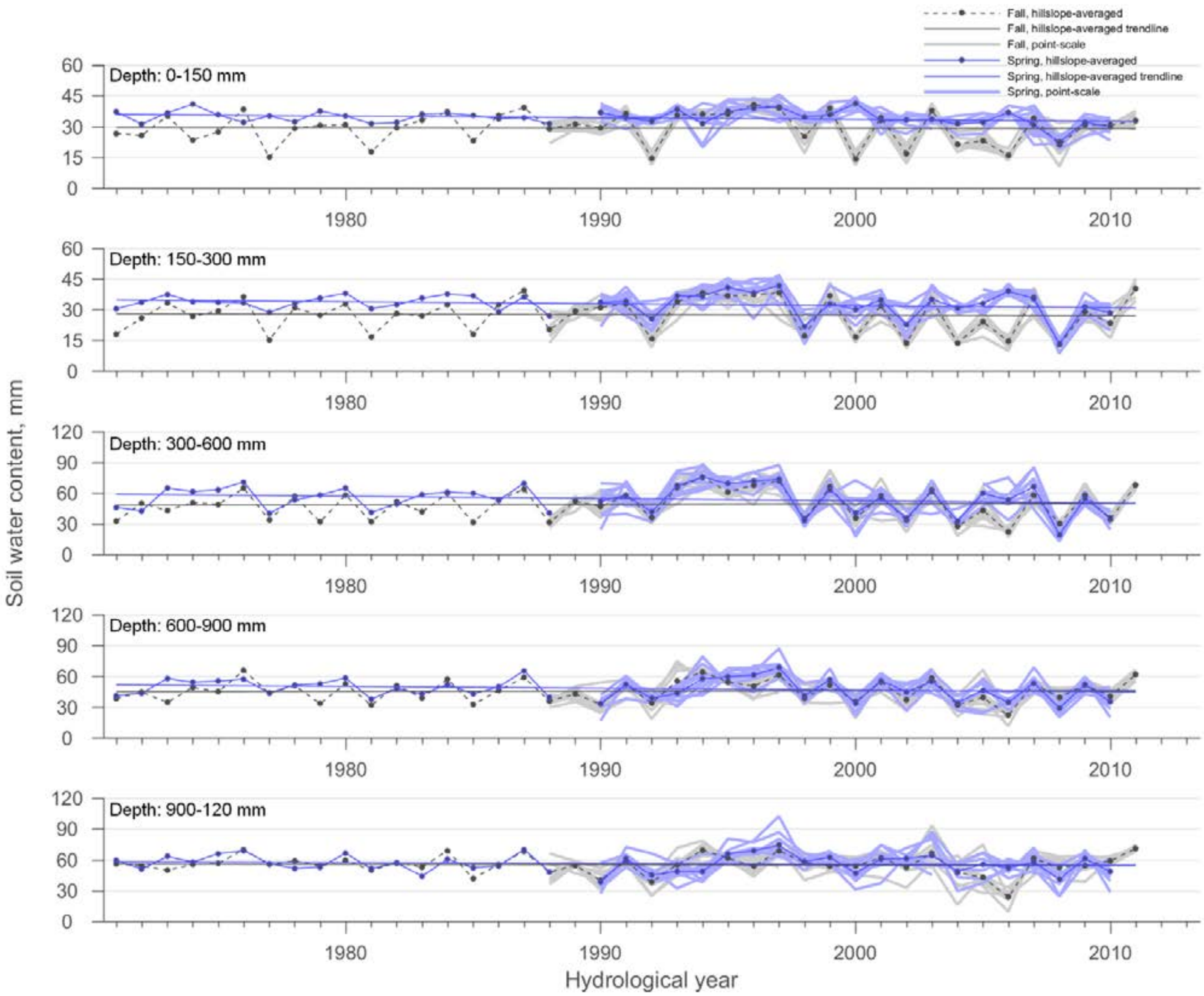

Figure 2.7 Seasonal soil water content (mm) measurements for Hillslope 3 for five depth layers (0-150 mm, 150-300 mm, 300-600 mm, 600-900 mm, 900-1200 mm below the soil surface) for fall (grey) and the following spring (blue). Grey circles and dashed line indicates the hillslope-averaged fall soil water content, while blue circles and solid line indicates the equivalent for spring. These data are the mean of point-scale soil water content measurements (9 points on each hillslope, 27 points total), for which the data were only archived from 1988 onwards. The solid grey and blue lines (data only available from 1988) indicate those point-scale soil water content measurements for spring and fall, respectively. Trendlines are indicated for the hillslope-averaged fall and spring soil water content data; no hillslope-averaged trendlines are significant.

Rainfall-runoff events occurred in 28 years out of the 52-year study period. For those years in which rainfall-runoff events occurred, mean runoff generated was 5 mm (Figure 2.3d). The majority of runoff (60% of the total volume) was generated by one-day rain events, and the remainder by multi-day rain events. A single one-day rainfall-runoff event was on average 20% larger in volume than a single multi-day rainfall-runoff event. While multi-day rain events

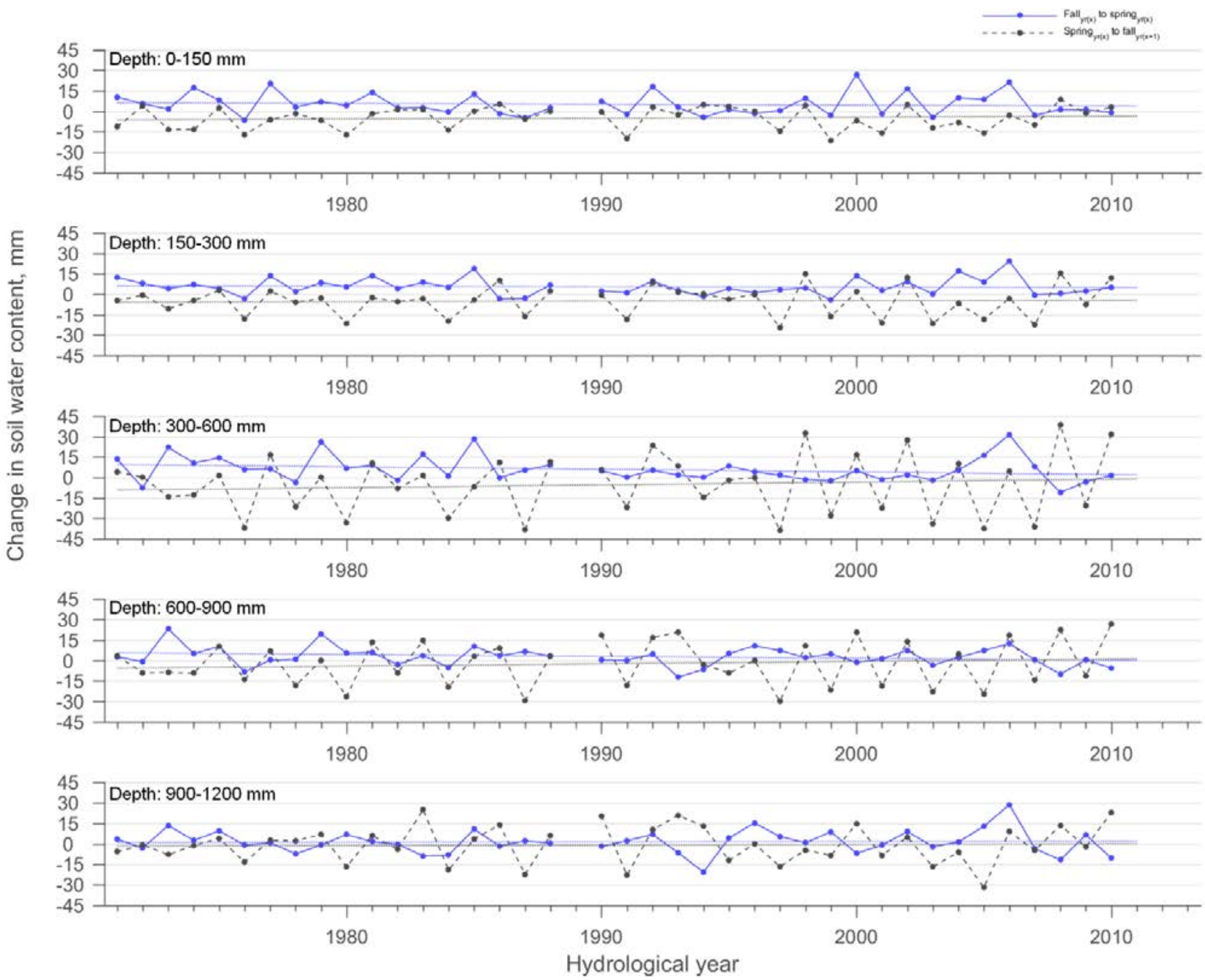

Figure 2.8 Change in hillslope-averaged seasonal soil water content (mm) for Hillslope 3 for five depth layers (0-150 mm, 150-300 mm, 300-600 mm, 600-900 mm, 900-1200 mm below the soil surface) for fall to spring (over winter and snowmelt; blue) and spring to the following fall (over summer; grey). Trendlines are indicated.

increased in occurrence over the period 1962-2013 (Figure 2.4a), there was no corresponding increase in occurrence of runoff events generated by those multi-day rain events (Figure 2.4b). Instead, there were shifts in rainfall-runoff timing and type: prior to 1976 and after 1996, rare runoff events were triggered predominantly by one-day rainfall events in March or April (Figure 2.5b), while in the intervening years runoff events were triggered predominantly by multi-day rain events throughout the summer months (Figure 2.4b).

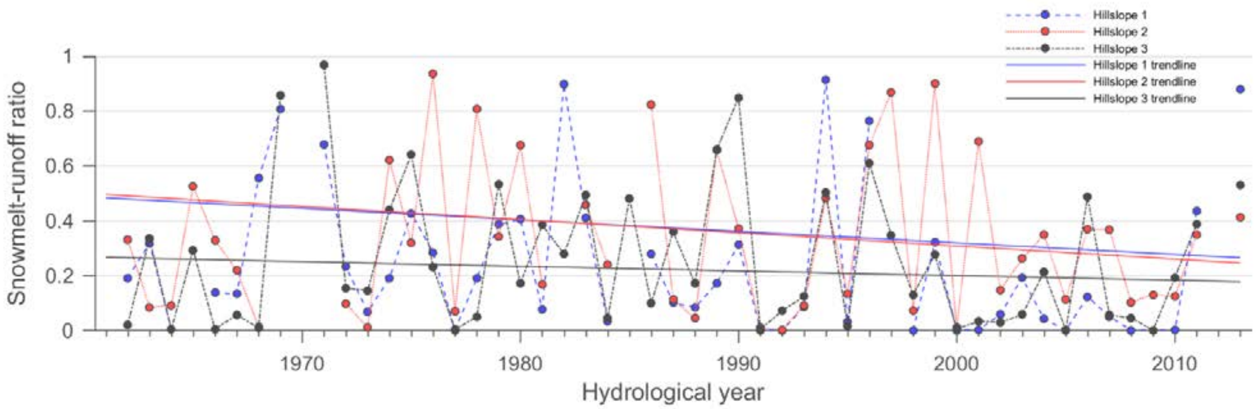

Figure 2.9 Snowmelt-runoff ratio for each hillslope, where runoff ratio is calculated as snowmelt-runoff divided by total snowfall. Trendlines are indicated.

2.6 Discussion

These 52 years of data on hillslope precipitation and runoff amounts from a research site on the northern Great Plains are the first such published data of their kind. For predicting and managing hillslope-scale water resources and sustainable agricultural production, it is essential to understand whether changing temperature and precipitation trends have induced changes in runoff and water availability at the hillslope scale. The observed climate trends over these 52 years are consistent with observations from elsewhere on the northern Great Plains (DeBeer *et al.*, 2015; Dumanski *et al.*, 2015; Mekis and Vincent, 2011; Shook and Pomeroy, 2012; Vincent *et al.*, 2007; Vincent and Mekis, 2006; Zhang *et al.*, 2000) and show: increased total precipitation, increased rainfall, increased winter and spring rainfall fraction, decreased snowfall and snow cover, and more multi-day rainfall events. However, our observed runoff trends are much less clear and in many cases not related to climate trends as we discuss in further detail in the following sub-sections.

2.6.1 More rainfall, but not more rainfall-runoff at the hillslope scale

Our findings show that the marked increase in rainfall has not yielded any increase in rainfall-runoff events at the hillslope scale. Further, despite the increase in multi-day rainfall events as compared to one-day rainfall events, there has been no similar change in the proportion of hillslope-scale rainfall-runoff events generated by those types of rainfall events.

High unfrozen infiltration capacities are a feature of hillslopes on the northern Great Plains (Elliott and Efetha, 1999). At the Swift Current hillslopes, measured unfrozen surface infiltration capacities range from 0.4 to 63.5 mm hr^{-1}, with a median of 13.9 mm hr^{-1} (field observations on Hillslope 2 in July-August 2013; reported in Seifert, 2014). Their spatial distribution, however, means that any runoff generated on limited patches of low infiltration is likely to run-on to areas of higher infiltration and infiltrate, especially since the downslope portion of Hillslope 2 has the greatest infiltration capacities (Seifert, 2014). This is consistent with the partial area concept of Betson (1964), who found that because of the spatial variability of soil properties, infiltration capacities and precipitation inputs, infiltration-excess runoff does not necessarily occur over an entire catchment or hillslope, but instead over small portion(s), during a rainfall- or snowmelt-runoff event (Tarboton, 2003).

Rare rainfall-runoff events at the hillslope scale are triggered by high intensity rains that exceed the soil's infiltration capacity along the full length of the flowpath, via infiltration-excess overland flow. Rainfall events that have triggered runoff on the hillslopes since 1995 (from when we have rainfall data at an hourly timescale) had peak rainfall intensities ranging from 0.6-14.8 mm hr^{-1}. Since multi-day rain events tend to be frontal and of lower intensity than one-day convective rainstorms, an exceptional frontal system would be needed to generate rainfall intensities that can exceed the infiltration capacity of the soil on the Swift Current hillslopes. Consequently, although the nature and total amount of rainfall has changed, the frequency of high-intensity rainfall has remained similar, at least since 1995. We hypothesize that, over the full 52 years of study, the number of rain storms of sufficient magnitude to create rainfall-runoff has not changed, so the occurrence of rainfall-runoff has not responded to the increase in rainfall.

Furthermore, we suggest that saturation-excess overland flow has not become a feature of these hillslope's surface runoff regime in recent years, despite increasing rainfall volumes. The soil remains unsaturated over the summer because the soil moisture at the start of summer has been steadily declining over the long-term, thus the 'starting' point for summer soil moisture is ever reduced, and because potential evapotranspiration over the summer still exceeds rainfall inputs.

2.6.2 Less snowfall, and also less snowmelt-runoff at the hillslope scale

While summer rainfall-runoff events have shown no response to changing rainfall, snowmelt-runoff has decreased nonlinearly in response to decreasing snowfall (snowmelt-runoff decreased by 59%, while snowfall decreased by 18%). We hypothesize that these seasonal differences are a result of the frozen and reduced infiltrability of the soil profile in the winter. A third of the annual precipitation melts onto the soil within a short, 1-2 week time period, generating large meltwater volumes at relatively fast rates that readily exceed the infiltrability of the soil. In the summer, the deep soils, with their high infiltrabilities and high evapotranspiration losses mean runoff is seldom generated despite increasing rainfall, and therefore the long-term runoff response to rainfall inputs is muted. But, this feature disappears for the winter season and spring freshet when the ground is frozen. The long-term decreasing snowmelt-runoff is not occurring because of any increase in infiltration due to thawed spring soils, since, for the years where we have soil temperature data, the snowmelt-runoff period occurred always over ground that was still frozen at the soil surface (data not shown). Indeed, total volume of infiltrated water appears to have actually decreased, as shown by the long-term decreasing trend in the amount of soil water added to the soil profile over winter and during snowmelt. Interestingly, there was little change in the general soil water content profiles over the winter months, from fall to the following spring, post-melt (Figure 2.10a): at greater depths, wet fall soil profiles remained wet in the following spring, and dry fall soil profiles tended to remain dry. This indicates that the infiltrated snowmelt-water is restricted to the surface layers, at least immediately following the snowmelt season, and that there is minor deep percolation of over-winter precipitation or spring snowmelt water into the deepest parts of the observed soil profile, consistent with the measured data shown in Figure 2.8.

By comparison, there is no distinction in fall soil water content profiles based on the previous spring's soil water content (Figure 2.10b). Therefore, the non-winter months exhibit vertical redistribution of soil water of over-summer precipitation and evapotranspiration. At the onset of spring snowmelt, the soil is still frozen and its infiltration capacity is greatly reduced: measured frozen surface infiltration capacities range from 0.09 to 2.57 mm hr^{-1}, with a median of 0.33 mm hr^{-1}. (These data were obtained from snowmelt-runoff laboratory experiments using intact soil cores, of different soil water contents, extracted from the Swift Current hillslopes (Appels *et al.*,

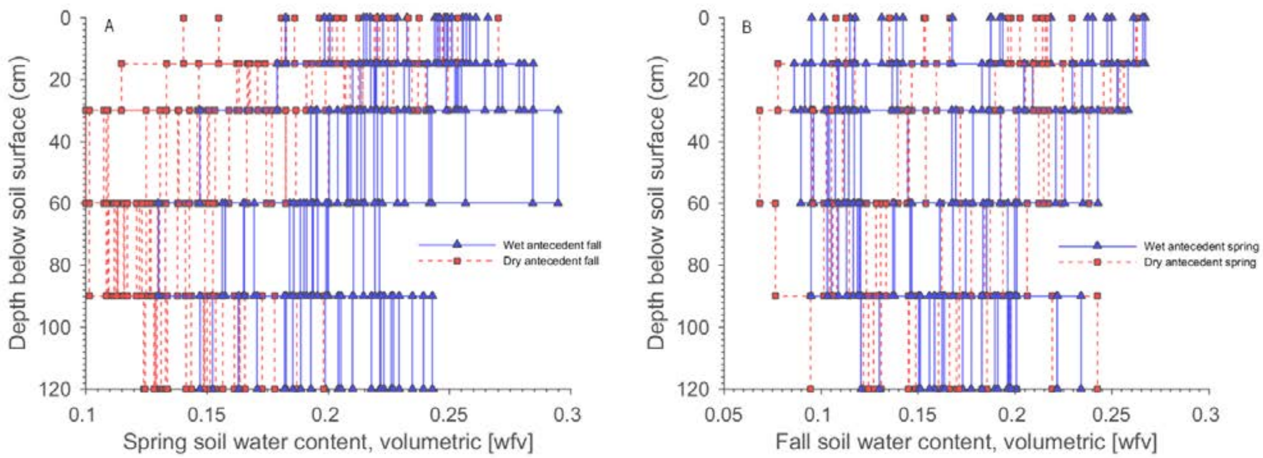

Figure 2.10 Hillslope-averaged volumetric soil water content measurements for five depth intervals in the soil profile for spring (A) and fall (B). The profiles are distinguished by whether, in the antecedent season, all measurement depths were drier (red, dashed line with square markers) or wetter (blue, solid line with triangle markers) than the long-term (1971-2013) mean. For example, in (A) the dry (red) antecedent soil moisture occurrences are predominantly on the left side (dry end) of the figure, which indicates that drier soil in fall remained drier than average come the following spring.

2017). The soil cores were frozen in a walk-in freezer and 'snowmelt' water was applied directly to the soil surface over four days, at delivery rates that simulated observed snowmelt rates at the Swift Current hillslopes.) When frozen, the soil lacks the summer sponge-like function so snowmelt-runoff at and over the soil surface is driven by and significantly related to the precipitation input. Of course, runoff amounts are always smaller than the corresponding snowmelt input amounts largely because there is some infiltration into frozen soil, as shown in these long-term results and in Coles *et al.* (2016) (Chapter 3), and also because of other factors acting at the surface, such as micro-surface depression storage, evaporation, and sublimation.

2.6.3 Hillslope-scale runoff response counter to that of catchment-scale

Overall, our observed changes in hillslope-scale runoff were highly equivocal and largely at odds with existing nearby catchment-scale observations (that have been subject to increasing wetland drainage) on the northern Great Plains (Dumanski *et al.*, 2015). Our decreasing snowmelt-runoff trends at the hillslope scale in response to decreasing snowfall are counter to Dumanski *et al.*'s (2015) catchment-scale findings at Smith Creek Research Basin, with a gross drainage area of 393

km^2 and annual precipitation of 442 mm, which saw a fivefold increase in snowmelt-runoff since 1975 (increase from 2600 dam^3 in 1975 to 17880 dam^3 in 2014), despite decreasing snowfall (decrease by 0.5 mm $year^{-1}$). Further, the lack of a clear change in rainfall-runoff events at the hillslope scale, despite increasing rainfall, is inconsistent with catchment-scale findings, which show a 150-fold increase in rainfall-runoff since 1975 (increase from 80 dam^3 in 1975 to 13350 dam^3 in 2014), in response to increasing rainfall (increase by 0.9 mm $year^{-1}$) (Dumanski *et al.*, 2015). It should be noted that surface runoff is expected to be more prevalent at Smith Creek (Dumanski *et al.*, 2015) than at Swift Current (this study) simply due to the different climatic and soil zones in which they are located. Smith Creek (Black Soil Zone) has higher annual precipitation and lower potential evapotranspiration, while Swift Current (Brown Soil Zone) sees lower annual precipitation and higher potential evapotranspiration (SAMA, 2015). Yet, this does not account for the differences in the observed trends.

At the catchment scale, streamflow generation on the northern Great Plains is strongly related to depressional storage (Shaw *et al.*, 2012; Shook *et al.*, 2015). When depressional storage is satisfied, the hydrological connectivity and contributing area of the catchment increases, resulting in much higher streamflow (Fang *et al.*, 2010; Shook and Pomeroy, 2012). In Dumanski *et al.* (2015), alterations of the landscape affected the catchment results. Drainage channel length increased 8-fold and the surface area of sloughs decreased by one-half. The loss of sloughs and the drainage into lower sloughs would decrease the depressional storage in the catchment and enhance flows by the mechanism described above. In fact, Dumanski *et al.* (2015) noted that some of the largest runoff events were from rainfall falling shortly after the snowmelt season, when sloughs were still relatively full and catchment conditions wet. Although unsatisfied depressional storage will also decrease runoff, increasing artificial drainage minimized this effect over time. Our hillslope scale lacks either the enhancement or damping effect of depressional storage (other than the micro- and meso-topographic, 0-1 m relief in the soil surface). Of the 17 Prairie catchments that Ehsanzadeh *et al.* (2016) analysed for climate-runoff changes, only Smith Creek showed significant changes in streamflow trends beyond that of the climate trends, with they too attributed to wetland drainage. Other changes to the Smith Creek landscape made it difficult to discern the climate signal in runoff in the study of Dumanski *et al.* (2015). The area of unimproved land decreased from 46% to 27% of the catchment. The amount of tillage on the cropland decreased and the proportion of land in summer fallow also decreased dramatically. In contrast, in our hillslope study we were able to relate any hydrologic change to climatic change. Overall, these counteracting findings demonstrate that we cannot linearly scale our hillslope observations up to catchment-scale predictions.

2.6.4 Decrease in snowmelt-runoff ratios

Our results showed that snowmelt-runoff ratio has decreased over time. In other words, the transformation of snowfall into snowmelt-runoff, hampered by processes such as snow redistribution, mid-winter ablation, snowmelt, and frozen soil infiltration (Shook *et al.*, 2015), has become less efficient. A part of this transformation is the nonlinear relationship that we observed between a decreasing trend in the amount of snowfall and a more amplified decreasing trend in the amount of snow cover at the onset of spring snowmelt. This snowfall-snow cover transformation was described by Shook *et al.* (2015) as occurring via snow redistribution and mid-winter ablation. Here, we see that this transformation becomes less efficient – gradually smaller proportions of snowfall are being retained as snow cover. We attribute this to two aspects. The first aspect is the effects of long-term land management changes in the region surrounding the Swift Current hillslopes on reducing the potential for blowing snow deposition on the hillslopes. These small Swift Current hillslopes have been managed differently than the larger surrounding region, which, over the last couple of decades, have seen reduced occurrences of fallow, more continuous cropping, increased chemical fallowing, and reduced tillage (McConkey *et al.*, 2012). All of these have likely enhanced snow-trapping and reduced blowing snow redistribution in the region, thus driving less snow delivery to, and deposition on, the hillslopes.

The second aspect is the effects of snow depth on processes that cause over-winter snow ablation. Any over-winter melting period would be more likely to expose the soil surface, and energy advected from these snow-free areas of lower albedo cause accelerated melting of the surrounding snowpack (O'Neill and Gray, 1973; Colbeck, 1988). There were neither trends in the mean, minimum, or maximum winter temperatures over the time period of study, nor in the number of winter days where temperatures rose above freezing (0°C). There was, however, a trend towards longer periods of cumulative above-freezing days: there were more frequent occurrences of five or more consecutive above-freezing days. There is therefore a feedback effect between reduced snowfall creating smaller snowpacks, which are then nonlinearly smaller because of the enhanced processes of over-winter snowpack melt for small snowpacks.

Also a contributor in the reduction of snowmelt-runoff ratio over time is the nonlinear relationship we observed between a decreasing trend in the amount of snow cover at the onset of snowmelt and the decreasing trend in snowmelt-runoff. In other words, runoff ratios have decreased over time even when we use the SWE of snow cover retained on the hillslopes (rather than the seasonal snowfall total) as the input parameter in the runoff ratio calculation. This is unusual since one might expect that over-winter ablation events (that increased significantly through the record), would create a lower-permeability ice lens at the soil-snow interface or within the snowpack (Gray *et al.*, 2001), and thus also increase runoff ratios. The decrease in runoff ratios does not seem to be due to increased infiltration of spring snowmelt (equivalent to Shook *et al.*'s (2015) second transformation, of snowmelt to runoff, via infiltration processes), since soil water content change between fall and spring has decreased over the long-term. Increased sublimation and evaporation of the snowpack and snowmelt water during the spring snowmelt are potential reasons for decreased runoff ratios.

2.6.5 Relationships between vegetation cover and snowmelt-runoff

Vegetation cover was important for snow accumulation, runoff ratio, and runoff signal in response to the 52-year precipitation signal. Fallow hillslopes showed reduced snow accumulation, compared to the years when the hillslopes had stubble residue on the hillslopes over winter. On average, instances of stubble on the Swift Current hillslopes exhibited 1.6 times as much snow accumulation as instances of fallow. This enhanced accumulation over stubble is supported by previous studies that found that, on the Canadian Prairies, wheat stubble fields had much smaller losses to blowing snow than did fallow fields due to variations that vegetation cover induces in wind speed near the snow surface (Cutforth and McConkey, 1997; Fang and Pomeroy, 2009; Pomeroy *et al.*, 1990; Pomeroy and Gray, 1995). Fang *et al.* (2007) found that, on prairie sites, snow accumulation in stubble fields is approximately 1.1 – 2.1 times greater than snow accumulation in fallow fields.

In the previous section, we described the nonlinear relationship between a decreasing trend in the amount of snowfall and a more amplified decreasing trend in the amount of snow cover at the onset of spring snowmelt. We observed that this long-term trend towards decreasing snow accumulation was strongest for fallow years. Again, this can be explained by reduced blowing snow across the general region, due to changing land management practices, thereby depositing less redistributed snow on the hillslopes. A further explanation is the typically smaller snowpacks that form during fallow conditions, compared to the snow-trapping stubble conditions, and the positive feedback effect that a small snow cover has on over-winter ablation processes. For instances of standing stubble, a strong positive relationship existed between mean temperature of above-freezing winter days and the proportion of SWE that was ablated during that season. For instances of fallow, mean wind speed was strongly correlated with the proportion of ablated SWE during individual seasons, on days where mean wind speed exceeded 7.5 m s^{-1} (the wind speed threshold for transport of fresh snow blowing snow; Li and Pomeroy, 1997). Land covers therefore have different main drivers of ablation; standing wheat stubble reduces surface wind speed (Cutforth and McConkey, 1997) so ablation is more dependent on energy input as indicated by air temperature, compared to fallow, for which wind transport of mass and energy is relatively more important.

For all hillslopes, runoff ratio was greater, but absolute runoff was smaller, under fallow conditions compared to vegetated conditions. Reduced infiltration under fallow conditions would explain these greater runoff ratios. Fang *et al.* (2007) found that the type of vegetation cover affects the soil water content at the time of freeze-up, with fallow fields generally being wetter than stubble fields due to less soil water extraction in the preceding growing season. Our data show higher runoff ratios under fallow conditions, which can be explained by these vegetation cover effects on soil water content. Fallow conditions also exhibited the strongest trend towards decreasing runoff ratios and decreasing runoff over time. Overall, the relative contributions of snowmelt-runoff from vegetated or fallow hillslopes was a combination (and sometimes trade-off) between the snow trapping qualities of stubble fields, and the typically higher soil water contents (albeit with some minor over-winter modifications; De Jong and Kachanoski, 1987; Gray *et al.*, 1985) of fallow fields compared to cropped and/or stubble fields.

2.6.6 Crop type effects on soil water

Not only did the occurrence of fallow have a noticeable effect on soil water storage, but the crop type in non-fallow years also did. The crop type influenced the fall, and often following spring, soil water contents. Pea and lentil crops use less soil water than wheat (Angadi *et al.*, 2008) and so soil water contents following summers when Hillslope 2 was cropped with pulses were greater than soil water contents following summers when wheat was the crop. This is true for soil water contents in both the fall (on average 30% higher) and the following spring (on average 4% higher), for all depths (data not shown). The increased stored soil water in spring following pulse crops compared with wheat is an important benefit of including pulse crops in crop rotations in this semi-arid climate (Gan *et al.*, 2003). Similarly, soil water contents following summers when Hillslope 1 was cropped along with green manure were higher than soil water contents following summers when wheat was the crop: fall soil water contents were on average 32% higher, while spring soil water contents were on average 20% higher (data not shown). This is due to water stored in soil after green manure growth termination in early July. At Swift Current, while soil water is typically high following legume green manure management, and higher than other crops, fallow conditions still are most efficient at storing soil water owing to the use of precipitation during the green manure crop growing period (Zentner *et al.*, 2004).

2.6.7 Outlook for the future of the northern Great Plains

Few studies (Fang and Pomeroy, 2007; Pomeroy *et al.*, 2009) have addressed the effects of future climate change on the hydrology, runoff generation processes, and agricultural productivity of the northern Great Plains. Climate change scenarios for the region project warming by between 0.5 and 3°C for the 2020s, and between 2 and 6.5°C for the 2080s, above baseline (1961-1999) temperatures (Barrow, 2009). The largest range of temperatures (and also the biggest rise in temperatures) are expected in the winter months (Barrow, 2009). Precipitation changes are uncertain: decreases by as much as 30% are projected by some scenarios into the 2080s, while increases are more likely to occur (Barrow, 2009). Further warming, therefore, will inevitably lower the influence of snow on hydrological systems, with cascading impacts on the streamflow regime and the magnitude and timing of runoff (Tetzlaff *et al.*, 2013).

Our 52-year analysis shows that the partitioning between surface, near-surface and deeper water sources is shifting. Over the last half-century, decreases in snowfall and snowpack depth have driven decreases in spring soil water content and spring snowmelt-runoff. These decreases seem to be damped if the previous growing season was cropped with wheat and had vegetation residue (stubble) on the fields over winter. Whether trends will continue in the same direction and to the same magnitude as those observed here is unclear, and depends upon the balance between runoff-enhancing and runoff-damping factors (*e.g.* less snowfall *vs*. more fall rainfall and wetter soils).

The amount of stored soil water is an important determinant of crop yield in this semi-arid climate where growing season moisture deficit is a certainty. Stored soil water is as important as growing season precipitation for crop yield (Campbell *et al.*, 1997) and the yield of crops grown on stubble is particularly sensitive to the amount of stored soil water (Kröbel *et al.*, 2014). Therefore, the reduction in soil water in the spring makes crop production, especially that grown on stubble, increasingly dependent on growing season precipitation. The trends towards increasing rainfall and increasing multi-day rain events are beneficial for crop production. However, in the semi-arid climate, drought is a continual risk. Multi-year droughts, such as that in 2000-2002, where multi-day rain events are in short supply, are a likely feature of future climate change in this region (Masud *et al.*, 2016). Such events result typically in poor crop yields, such as was seen in 2001 (Masud *et al.*, 2016). Decreasing soil water reserves in the spring will accentuate the impact of droughts. Minimum tillage to promote infiltration into frozen soils through macropores, and continuous cropping systems to reduce blowing snow sublimation may help promote higher post-melt soil water contents (Elliot *et al.*, 2001). These might ameliorate the detrimental effects on agriculture from low rainfall in the growing season, and earlier spring runoff with greater evaporation losses before the growing season (Cutforth *et al.*, 1999).

The reduction in spring freshet volumes has important ramifications for on-farm water supplies. Investments for additional water collection and storage and/or for transporting water will be needed to meet water demands at farmsteads. Such investments may not be warranted for pastures, so some pastures may have to be left ungrazed when there is insufficient volumes of water in dugouts to meet the needs of livestock. As a result, lower runoff increases the costs of agriculture.

2.7 Conclusions

Our analysis of a 52-year, hillslope-scale, climate-runoff record from the northern Great Plains shows that snowmelt-runoff and spring soil water amounts have decreased in response to winter snowfall decreases, but that rainfall-runoff has shown no response to increases in rainfall or shifts to more multi-day rain events. We attribute these seasonal differences to soil infiltrability, soil storage modulation, and differences in evapotranspiration between the summer and winter months. In the summer, thawed, deep, high-infiltrability soils with high evapotranspiration demands act to buffer the long-term runoff response to rainfall. In the winter and spring freshet, frozen ground limits infiltration and means runoff responses more closely mirror the snowfall and snowmelt trends (albeit with some nonlinear trend relationships between snowfall and runoff, which could be explained in part by enhanced over-winter ablation of smaller snowpacks). These findings are different from climate-runoff relationships observed at the catchment scale on the northern Great Plains. This is likely due to the confounding effect of landscape alteration, especially drainage. These long-term findings have clear implications for agriculture on the northern Great Plains. The hydrology of hillslopes is important for dryland crop production and for on-farm water supplies. Meeting water needs in a situation of declining runoff, declining spring soil water, and resultant accentuation of summer drought impacts will increase costs to agriculture.

2.8 Transition statement

Chapter 2 has established the general trends and factors in precipitation-runoff responses over the 52-year period. However, there was significant year-to-year variation in these runoff trends and a nonlinear decrease in snowmelt-runoff ratio over time. The year-to-year variation and nonlinearities were likely driven by nuances, interactions, and feedbacks between controls. These were disentangled using data mining, which is the subject of Chapter 3. Data mining also was an ideal tool with which to analyze in greater detail the effects of factors that arose in the discussion of Chapter 2, such as crop type and land cover, tillage, and over-winter ablation via melt or blowing snow. Overall, the insights and conclusions drawn from Chapter 2 on climate-runoff trends are a useful backdrop to the work in Chapter 3, which sought to understand the factors behind the variations in these trends.

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2.10 Author contributions

JJM and AEC conceived the rationale for the study. BGM provided the long-term dataset. AEC carried out the analyses. AEC wrote the paper. BGM and JJM commented on the manuscript and contributed to the text in later iterations.

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CHAPTER 3

THE HIERARCHY OF CONTROLS ON SNOWMELT-RUNOFF GENERATION OVER SEASONALLY-FROZEN HILLSLOPES

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3.1 Abstract

Understanding and modeling snowmelt-runoff generation in seasonally-frozen regions is a major challenge in hydrology. Partly, this is because the controls on hillslope-scale snowmelt-runoff generation are potentially extensive and their hierarchy is poorly understood. Understanding the relative importance of controls (*e.g.* topography, vegetation, land use, soil characteristics, and precipitation dynamics) on runoff response is necessary for model development, spatial extrapolation, and runoff classification schemes. Multiple interacting process controls, the nonlinearities between them, and the resultant threshold-like activation of runoff, typically are not observable in short-term experiments or single-season field studies. Therefore, long-term datasets and analyses are needed. Here, we use a 52-year dataset of runoff, precipitation, soil water content, snow cover, and meteorological data from three monitored *c.*5 ha hillslopes on the Canadian Prairies to determine the controls on snowmelt-runoff, their time-varying hierarchy, and the interactions between the controls. We use decision tree learning to extract information from the dataset on the controls on runoff ratio. Our analysis shows that there was a variable relationship between total spring runoff amount and either winter snowfall amount or snow cover water equivalent. Other factors came into play to control the fraction of precipitated water that infiltrated into the frozen ground. In descending order of importance, these were: total snowfall, snow cover, fall soil surface water content, melt rate, melt season length, and fall soil profile water content. While mid-winter warm periods in some years likely increased soil water content and/or led to development of impermeable ice lenses that affected the runoff response, hillslope memory of fall soil moisture conditions played a strong role in the spring runoff response. The hierarchy of these

controls was condition-dependent, with the biggest differences between high and low snow cover seasons, and wet and dry fall soil moisture conditions. For example, when snow cover was high, the top three controls on runoff ratio matched the overall hierarchy of controls, with fall soil surface water content being the most important of these. By comparison, when snow cover was low, fall soil surface content was relatively unimportant and superseded by four other controls. Existing empirical methods for predicting infiltration into frozen ground failed to adequately predict runoff response at our site. Our analysis of the hierarchy of controls on meltwater runoff will aid in focusing new model approaches and understanding what to focus future measurement campaigns on in snowmelt-dominated, seasonally-frozen regions.

3.2 Introduction

Understanding the hierarchies and the condition-dependent relative importance of controls (*e.g.* topography, vegetation, land use, soil characteristics, and precipitation dynamics) on runoff response is a major challenge in hydrology (Jencso and McGlynn, 2011). Formulating a hierarchy of controls for runoff is necessary for model development (Uchida *et al.*, 2005), a key component of spatial extrapolation (Cammeraat, 2002), and a necessary building block for runoff classification schemes (Barthold and Woods, 2015).

On the Canadian Prairies, spring snowmelt is the dominant runoff-producing event of the year, driving typically 80% or more of annual runoff (Granger *et al.*, 1984). While some summer runoff is generated by intense rain storms where high rainfall intensities drive infiltration-excess overland flow over localized areas (Shook and Pomeroy, 2012), the controls on these types of events are few in number: rainfall intensity, rainfall magnitude, and antecedent soil moisture conditions (Shook and Pomeroy, 2012). By comparison, the hydrologically more important snowmelt events are much more complicated and affected by multiple interacting factors including snow accumulation, distributed melt inputs, seasonally-frozen ground, ice lenses, and variable pre-melt soil moistures, which combine to produce highly nonlinear runoff responses (Fang *et al.*, 2007; DeBeer and Pomeroy, 2010; Ireson *et al.*, 2013). Consequently, understanding and modeling snowmelt-runoff generation remains problematic throughout many areas of North America and northern Eurasia where snowmelt-influenced, seasonally-frozen ground dominates runoff generation. Nevertheless, in these areas, there is a need to understand snowmelt-runoff generation as it is a critical source of water for human activities and aquatic ecosystems, and snowmelt can cause serious flooding.

In western Canada, snowmelt-runoff has been the subject of many experimental and modeling studies aimed at understanding individual controls: the effects of snow accumulation and redistribution (*e.g.* Pomeroy and Gray, 1995; Fang and Pomeroy, 2009), snowmelt processes (*e.g.* Gray and Landine, 1988), land use and land cover effects (*e.g.* Elliot and Efetha, 1999; Van der Kamp *et al.*, 2003), topography (*e.g.* Shaw *et al.*, 2012), and seasonally-frozen soil (*e.g.* Granger *et al.*, 1984; Gray *et al.*, 2001). However, these studies, and our resultant understanding, are based upon mostly short-term experiments and single-season runoff events. Temporally- and spatially-unstable activation of runoff is the product of nonlinearities and interactions between the various process controls that are not observable in short-term field studies. Much longer records are needed to witness these combinations and interactions of process factors. However, such datasets are rare.

Recently, Coles *et al.* (2017) (Chapter 2 of this thesis) presented a 52-year dataset of snowmelt-runoff from three adjacent monitored hillslopes in southern Saskatchewan, Canada. That work showed that long-term snowmelt-runoff and spring soil water content have decreased in response to winter snowfall decreases, while rainfall-runoff has shown no response to changes in rainfall regimes (Coles *et al.*, 2017; Chapter 2). They attributed this to the seasonal differences in soil infiltrability, indicating that the controls on infiltration are likely to be most important for snowmelt-runoff amount, as others have shown (Fang *et al.*, 2007; Ireson *et al.*, 2013). However, we still do not know about the hierarchies, interactions, and feedbacks between these controls, and any condition-dependent differences in their behaviour. Here, we use the same 52-year dataset to explore these aspects, and contribute for the first time new understanding of the hierarchical importance of runoff controls. And, over a multi-decadal time period, if and how such controls on meltwater runoff interact.

We use decision tree learning (De'ath and Fabricius, 2000) as an investigative tool to extract information from the long-term dataset about the hierarchical controls on runoff generation. Decision tree learning (including classification or regression trees) is an established data mining tool in ecological studies (*e.g.* Spear *et al.*, 1994; Rejwan *et al.*, 1999; De'ath and Fabricius, 2000). It has more recently been incorporated into hydrological studies to leverage process understanding from long-term datasets in temperate regions (*e.g.* Iorgulescu and Beven, 2004; Tighe *et al.* 2012; Scholefield *et al.*, 2013; Galelli and Castelletti, 2013). To our knowledge, no studies have used decision trees to explore snowmelt-runoff generation. Decision tree learning is fast, conceptually simple, data-based, nonlinear, and non-parametric. Importantly, it allows insights into complexities, nonlinearities, equifinalities, interactions, and feedbacks in the data, which are illustrated clearly in resultant tree-like diagrams (Rejwan *et al.*, 1999; Iorgulescu and Beven, 2004; Michaelides *et al.*, 2009). Here we use the decision tree approach to determine the hierarchies of controls on snowmelt-runoff generation in a seasonally-frozen, snowmelt-dominated region, and any interactions and feedbacks between those controls. Specifically, we focus on the following research questions:

i) What is the relationship between annual snow input and snowmelt-runoff output over the 52 years of data?
ii) What is the hierarchy of controls on snowmelt-runoff amount?
iii) Does the hierarchy vary under different conditions?
iv) What are the interactions and feedbacks between the hierarchical process controls?

3.3 Study site and dataset

The study site, known as the Swift Current hillslopes, at South Farm of Swift Current Research and Development Centre of Agriculture and Agri-Food Canada, Swift Current, Saskatchewan, Canada (50°15'53"N 107°43'53"W) on the Canadian Prairies is a set of three adjacent agricultural hillslopes between 4.25 and 4.86 ha in size (Figure 3.1). Coles *et al.* (2017) (Chapter 2) provided a brief description of the study site, which has undulating topography and shallow north-facing slopes with gradients of 1-4%. Grassed berms around the perimeters of all three hillslopes prevent runoff from moving between the hillslopes or entering from adjacent land. The only outlet from each hillslope is through a 0.61 m H-flume at the northwest corner of each hillslope. The soil is a

Figure 3.1 Aerial photograph (facing south) of the Swift Current hillslopes (from right to left: Hillslope 1, Hillslope 2, Hillslope 3), taken in a year when wheat was grown. The three small heated huts at the northwest corners of the hillslopes, which house the runoff-monitoring equipment, are visible. Photograph reproduced, with permission, from Cessna *et al.* (2013).

Swinton silt loam and classified as an Orthic Brown Chernozem (Cessna *et al*., 2013). The hillslopes typically are under an annual rotation of wheat (*Triticum aestivum* L.) and fallow. Exceptions to this are: a period (1977-1980) of grass (*Psathyrostachys juncea* (Fisch.) Nevski) and a period (1982-1985) of annual wheat on Hillslopes 1 and 2; an annual rotation (1994-2010) of wheat and legume green manure (*Lathyrus sativus* L.) on Hillslope 1; and an annual rotation (2004-2011) of wheat and pulses (lentils and peas; *Lens culinaris* L. and *Pisum sativum* L., respectively) on Hillslope 2. Hillslope 3 is the only hillslope that has a consistent two-crop rotation and consistent tillage management throughout the 52 years. The hillslopes have largely been under conventional tillage practice, with the exception of the period 1993-2011 when Hillslope 2 was switched to zero tillage practice. During the period 1993-2004 on Hillslope 2, when the wheat-fallow rotation coincided with the zero tillage period, there was constant standing stubble or standing crop.

From 1962-2013, runoff, snow cover, and soil water content were monitored on the hillslopes. This rich dataset is coupled with long-term meteorological data recorded at a nearby (*c.* 700 m to the south-southeast) Environment and Climate Change Canada standard meteorological station. Data have been used primarily for studies on the effects of agricultural land management practices on runoff water quality, chemical transport, and soil erodibility (Nicholaichuk and Read, 1978; McConkey *et al.*, 1997; Cessna *et al.*, 2013). More recently, data were used to study the effects of changing precipitation form and amounts on rainfall- and snowmelt-runoff generation (Coles *et al.*, 2017; Chapter 2). Over the 52 years of record, runoff in 22 years was generated exclusively during snowmelt on all three hillslopes (*i.e.* no rainfall-driven contribution to annual runoff on any hillslope), and runoff in 27 years was generated by both snowmelt and rainfall on one or more hillslopes (with an average of 75% annual runoff from snowmelt). The long-term mean annual snowmelt-runoff depth is 29 mm (Coles *et al.*, 2017; Chapter 2). Snowmelt-runoff at this site, and on the Prairies as a whole, is generated as infiltration-excess overland flow when a rapid release of relatively large volumes of water from the snow cover (usually in a short, one to three week long snowmelt season) occurs over frozen ground of limited infiltration capacity (Granger *et al.*, 1984; Coles *et al.*, 2017; Chapter 2).

3.3.1 Meteorological data

Daily (1962-1995) and hourly (1995-present) meteorological data are available from the Environment and Climate Change Canada meteorological station. The data used here include: precipitation (measured using a Belfort weighing gauge), air temperature (daily maximum, minimum, and mean measured inside a Stevenson Screen, and then hourly data measured using a Campbell Scientific HMP35C Temperature and Relative Humidity Probe), wind speed (measured at 2 m and 10 m above ground surface using an RM Young Anemometer Model 05103), and soil temperature (measured at 5, 10, 20, 50, 100, 150 and 300 cm depths using 107B Campbell Scientific Temperature Probes).

3.3.2 Runoff data

Runoff was measured from 1962-2013 with a heated H-flume, stilling well, and a Stevens water level chart recorder at the outflow of each hillslope (Figure 3.1). Rating curves for each flume/hillslope were used to determine runoff depths (mm) from the stilling well water levels. Runoff depths on hourly, daily, and seasonal timescales (mm) were calculated using a rating curve for the flumes (following Cessna *et al.*, 2013).

3.3.3 Soil water content data

Gravimetric soil water content was measured twice per year from 1971-2013 on each hillslope. In October (prior to freeze-up) and April (following spring snowmelt) each year, gravimetric soil water content was measured at five increments over the soil profile (0-15, 15-30, 30-60, 60-90, and 90-120 cm). This was done on a nine-point grid on each hillslope, with each of the nine points being in approximately the same location as in previous years (within *c.* 10 m). Hillslope-averaged soil water content at each depth was calculated from the point-scale data. Both hillslope-averaged and point-scale data were recorded from 1980-2013, and from 1971-1979 only hillslope-averaged data were recorded. We converted all soil water content data from gravimetric to volumetric (vwc) using bulk density data for each depth interval (which ranges from 1.22 g cm^{-3} at the soil surface to 1.51g cm^{-3} at a depth of 100 cm).

3.3.4 Snow cover data

Snow cover depth and density were measured, and SWE calculated (hereafter referred to as ScWE for snow cover SWE), for each hillslope during manual snow surveys each year from 1965-2013 on the same nine-point grid as that used for soil water measurements. The means of the nine points were calculated to give three hillslope averages. These were repeated several times from January to March. Measurements from the most recent snow survey before snowmelt were used to predict the ScWE on each hillslope at the onset of spring snowmelt. The timing of the most recent snow survey before snowmelt would have been difficult, and the snow cover might have changed significantly before snowmelt started. The ScWE calculated from this snow survey therefore might not be an accurate representation of the actual ScWE at the onset of melt. This is indicated by the fact that, for many years, the runoff ratios from the hillslopes (where, here, runoff ratio is the total seasonal runoff (mm) divided by the ScWE (mm)) exceed 1.

3.3.5 Quality control

The data were checked and corrected for missing or unrealistic data. During the snowmelt seasons of 1982 and 1985, researchers observed high volumes of snowmelt overwhelming the raised berms causing flow onto Hillslope 2 and Hillslopes 1 and 2, respectively, from adjacent land to the south. For these three occurrences, the total seasonal runoff depth from the hillslope exceeded the depth of total winter snowfall (*i.e.* the runoff ratio exceeded 1). These runoff data were omitted from our analysis and instead given a missing data notation.

3.4 Methods

Decision trees determine a set of ‘if-then’ conditions between the response and predictor variables and split the dataset according to the largest deviance produced (Rejwan *et al.*, 1999; Michaelides *et al.*, 2009). The result is illustrated in a simple tree-like diagram, with branches, nodes, and leaves. Each final partition (branch) is associated with a certain set of conditions. Branches are composed of nodes. At each node, the dataset is split according to agreement with a single rule (*e.g.* total seasonal precipitation > 50 mm). Splitting continues until the dataset is divided as much as possible. Branches end in a terminal node (leaf), which represents the final partitioning of the data, and the predicted response given the set of conditions dictated by the nodes of the branch. Finally, the tree can be pruned, which removes leaves and nodes with little predictive power, reduces overfitting, and therefore improves predictive accuracy.

To construct the decision trees, we used the ‘classregtree’ CART algorithm of MATLAB (MathWorks, Inc.). We calculated the runoff ratio, defined as total runoff divided by total seasonal snowfall (in SWE; hereafter referred to as SfWE for snowfall SWE) measured from the start of the hydrological year to the end of snowmelt-runoff, for each hillslope and for each snowmelt season. This resulted in 140 runoff ratios ranging from 0 to 1. We classified the runoff ratios into five equally-sized classes, separated at the 20th, 40th, 60th and 80th percentiles of the runoff ratios. These runoff ratio classes then formed the response (Y) variables for the CART function (Table 3.1). We decided upon five classes so that one class could represent the median runoff ratio (40th – 40th percentile class), two could represent the extreme high and the extreme low runoff ratios, and two could represent the medium-high and the medium-low runoff ratios, thereby covering a reasonable spread of the observations. Further, five classes allowed for 28 observations per class, which was an acceptable number of observations for the classification method of CART. We used the classification method of CART so that the predicted outcome is the class to which the data

Table 3.1 Response (Y) variable classes, dependent on runoff ratio.

Response (Y) variable (runoff ratio class)	Runoff ratio (RR)
1	$0 < RR \leq 0.032$
2	$0.032 < RR \leq 0.12$
3	$0.12 < RR \leq 0.28$
4	$0.28 < RR \leq 0.48$
5	$0.46 < RR \leq 1$

belongs (one of the five runoff ratio classes), instead of the regression method, which is when the predicted outcome is considered to be a real number. The predictor (X) variables for each hillslope and each season were derived from the long-term dataset (Table 3.2). We used SfWE, rather than ScWE, to calculate the runoff ratios used in the response variables and also to calculate the predictor variable 'melt rate'. This was to avoid introducing additional error due to the uncertainties associated with the snow cover data.

An advantageous feature of decision tree construction is that variables can consist of both numerical and categorical data (*e.g.* "fallow" or "wheat" crop types). After the algorithm had divided the dataset as much as possible, we then pruned the tree to a tree size that maximised predictive accuracy and ensured that all leaves were left with response variable datasets of size N>1. The runoff ratio class at each leaf was the mode of the classes predicted by that branch of the tree.

We first constructed one decision tree (the 'primary' decision tree) using the entire dataset. The controls on runoff ratio were the variables that the decision tree used in its construction. In a second round of decision tree construction, we then took each of those variables and used them to divide the dataset into two evenly sized halves of the dataset, split by the median of that variable. We constructed two decision trees using these two halves of the dataset. This approach was to explore directly the reasons for why high or low runoff and runoff ratios might be found under opposite conditions (*e.g.* under both high snowfall and low snowfall conditions). It was also useful for determining the condition-dependent hierarchy of controls.

Table 3.2 Predictor (X) variables, their descriptions including how they were derived from the long-term dataset, and their minimum, mean, and maximum values.

Predictor (X) variable name	Units	Description	Mean (Min-Max)
Surface depression storage	mm	Calculated using a 2-meter resolution digital elevation model (DEM) of each hillslope. Hillslope 1 = 2.60 mm; Hillslope 2 = 0.70 mm; Hillslope 3 = 1 mm	1.43 (0.70-2.60)
Topographic wetness index (TWI)	-	Mean hillslope TWI calculated using a 2-meter DEM of each hillslope, following Beven and Kirkby (1979). Hillslope 1 = 5.54; Hillslope 2 = 5.70; Hillslope 3 = 6.03	5.76 (5.54-6.03)
Land cover	-	Classified as fallow and grass (1) or wheat (2) for the previous summer's crop.	1.49 (1-2)
Fall soil surface water content	(fraction)	Mean volumetric hillslope soil water content (θ_v) in October at the surface (0-15 cm).	0.154 (0.0713-0.222)
Fall soil profile water content	(fraction)	Mean volumetric hillslope soil water content (θ_v) in October for the soil profile (0-120 cm).	0.123 (0.069- 0.178)
Total seasonal snowfall	mm SfWE	Total snowfall depth from Oct 1st to the end of the runoff period in the following spring, measured at the meteorological station.	80.5 (37.0-152)
Number of warm winter days	-	Number of days each year between Oct 1st and the season's last snow survey that had snow cover (at the meteorological station) and mean air temperature > 0 °C.	4.53 (0-12.0)
Mean temperature on warm winter days	°C	Mean air temperature for days between Oct 1st and the season's last snow survey that had snow cover (at the meteorological station) and mean air temperature > 0 °C.	2.49 (1.05-5.08)
Mean daily wind speed above blowing snow threshold	m s^{-1}	Mean daily wind speed for days when mean wind speed > 7.5 m s^{-1} (minimum threshold for blowing snow redistribution of fresh dry snow on northern prairies; Li and Pomeroy, 1997).	8.85 (8.25-9.59)
Snow cover	mm ScWE	Mean snow cover water equivalent (SWE_C) on each hillslope in the last snow survey before the start of the snowmelt season.	33.1 (0-121)
Spring temperature gradient	°C	Temperature gradient over the seven days prior to the date of peak runoff on Hillslope 2.	1.66 (-0.0655-5.12)
Melt season length	days	Number of days after the season's last snow survey that had snow cover (at the meteorological station) and mean temperature > 0 °C.	3.80 (1-8)
Melt rate	mm d^{-1}	Calculated as: total seasonal snowfall / melt season length.	26.4 (9.45-111)
Date of peak runoff	-	The date when maximum runoff depth occurred, for each hillslope.	Mar 14 (Jan 10-Apr 24)
Thawed layer depth	cm	The depth between the soil surface and the top of the frozen layer, at 4pm on the date of maximum runoff, determined using soil temperature data at the meteorological station.	4.62 (0-147)

We quantified the predictive accuracy (or amount of variance successfully explained) of the trees at each leaf using the resubstitution method (Spear *et al.*, 1994). For this, we calculated the percentage of the runoff ratio class predicted correctly at that leaf. The overall predictive accuracy of the tree was the mean of the predictive accuracies at each leaf. While cross-validation is generally the preferred method of estimating accuracy, our dataset was too small to use this method. We quantified the hierarchy of controls by ranking the controls' positions in each decision tree and weighting that rank by the number of nodes in the tree. If a variable appeared more than once in the tree, we summed the ranking position of each of the nodes at which it occurred, prior to weighting.

3.5 Results

3.5.1 The hierarchy of controls on snowmelt-runoff

Figure 3.2 shows that there was no unique relationship between precipitation input and total seasonal runoff output, where inputs were the SfWE (Figure 3.2a), and ScWE (Figure 3.2b). Six predictor variables were identified by the primary decision tree (Figure 3.3, Table 3.3) to explain runoff response: total snowfall (SfWE), snow cover (ScWE), fall soil surface water content (0-15 cm), melt rate, melt season length, and fall soil profile water content (0-120 cm) (Table 3.4). Therefore, 13 decision trees were constructed in total (the primary decision tree using the entire

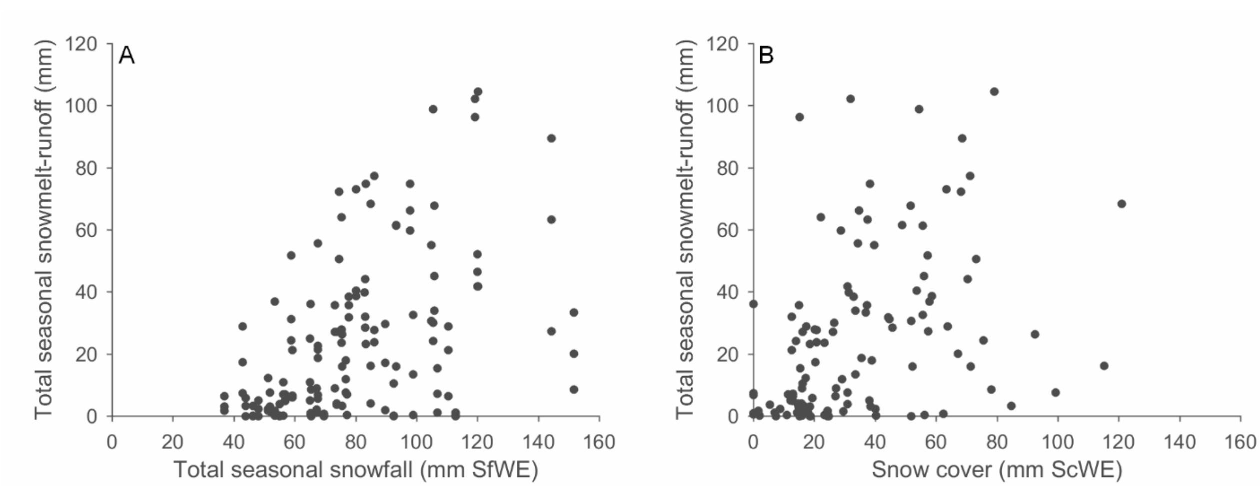

Figure 3.2 Spring snowmelt-runoff events on the Swift Current hillslopes, showing A) the relationship between total seasonal snowfall (mm SfWE) and total seasonal snowmelt-runoff (mm); and B) the relationship between snow cover (mm ScWE) and total seasonal snowmelt-runoff (mm).

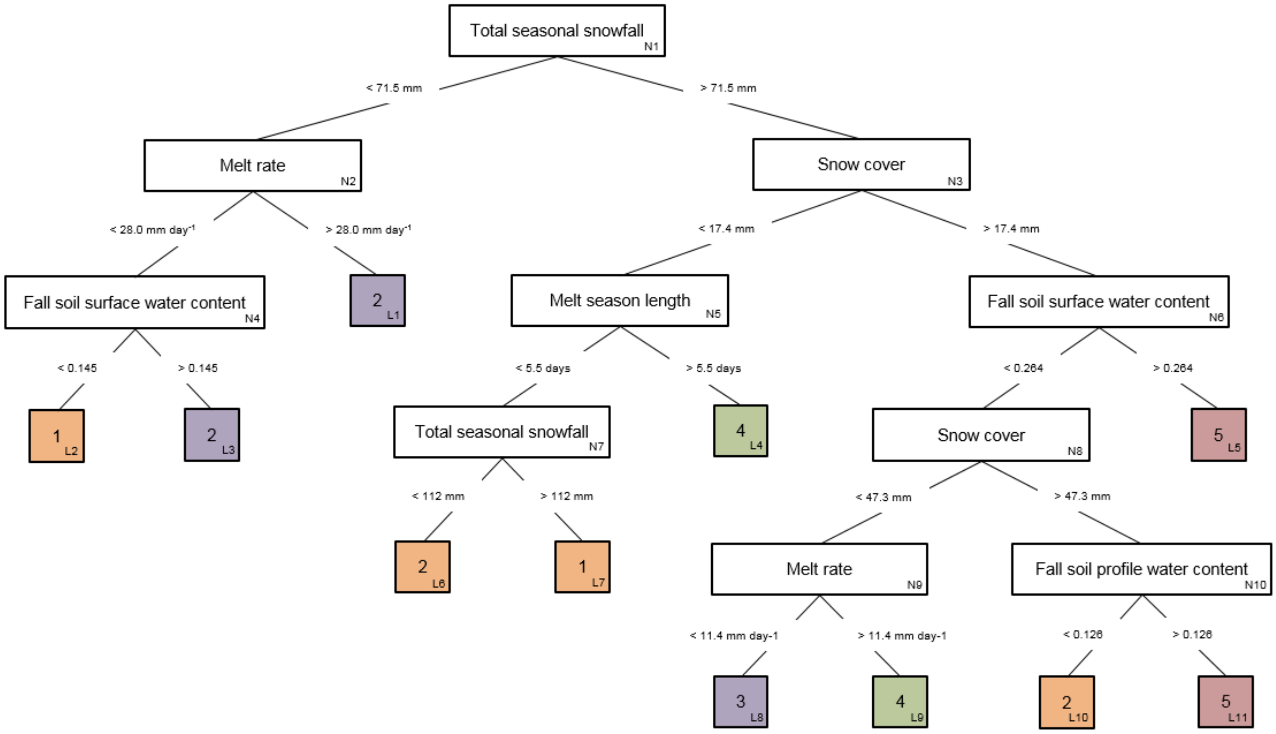

Figure 3.3 Primary decision tree for predicting runoff ratios. The tree shows the variables used to explain runoff ratio, located at the nodes (N1, N2, etc.; white boxes) and the resultant predicted runoff ratio class at the leaves at the ends of each branch (L1, L2, etc.; coloured boxes). Table 3.3 gives the runoff ratio class data at each node and leaf, and the predictive accuracy at each leaf. The leaf colours are in reference to Figure 3.5.

dataset, and six pairs of smaller decision trees). The primary decision tree (Figure 3.3; Table 3.3) contained 10 nodes with six predictor variables and 11 leaves. The primary decision tree explained 70% of the variance of the runoff ratio classes (Figure 3.4). The 12 secondary decision trees, constructed by splitting the dataset at the medians of each of those variables, identified fewer predictor variables and had fewer nodes and leaves. Their identified controls and hierarchies are given in Table 3.5. Eight of those decision trees explained more of the variance of the runoff ratio classes than the primary decision tree (Table 3.6). The decision trees using datasets characterised by high total snowfall, high snow cover, high soil surface water content, and high soil profile water content all were better at predicting high runoff ratios (class 4-5) than their low counterparts, and vice versa for low runoff ratios (class 1-2).

Of the (15) predictor variables in Table 3.2, 12 appeared at least once in any of the decision trees. Snow cover, total snowfall, and fall soil surface water content appeared in the most trees (10, eight,

Table 3.3 Runoff ratio class data at each node and leaf (underlined), and the predictive accuracy at each leaf, for the primary decision tree.

Node number	Parent node	Sample size						Predicted RR class	Predictive accuracy
		RR class 1	RR class 2	RR class 3	RR class 4	RR class 5	Total		
N1	-	28	28	28	28	28	140	-	-
N2	N1	18	18	12	5	4	57	-	-
N3	N1	10	10	16	23	24	83	-	-
N4	N2	11	6	2	0	2	21	-	-
L1	N2	0	6	4	1	1	12	2	50
N5	N3	6	6	4	2	2	20	-	-
N6	N3	1	3	11	17	21	53	-	-
L2	N4	9	3	1	0	0	13	1	69
L3	N4	0	3	1	0	1	5	2	60
N7	N5	6	6	3	0	0	15	-	-
L4	N5	0	0	1	2	2	5	4	40
N8	N6	0	2	8	15	14	39	-	-
L5	N6	0	0	0	0	5	5	5	100
L6	N7	3	6	3	0	0	12	2	50
L7	N7	3	0	0	0	0	3	1	100
N9	N8	0	0	3	13	2	18	-	-
N10	N8	0	2	5	2	12	21	-	-
L8	N9	0	0	2	0	0	2	3	100
L9	N9	0	0	1	13	2	16	4	81
L10	N10	0	2	1	1	0	4	2	50
L11	N10	0	0	4	1	12	17	5	71

and eight trees, respectively), often with multiple occurrences in any one tree. This confirms that they were important controls on snowmelt-runoff ratio. Surface depression storage, date of peak runoff, and mean daily wind speed appeared in four trees each. Melt rate, melt season length, and fall soil profile water content appeared in three trees each. Lastly, TWI, spring temperature gradient, and land cover appeared in one tree each.

We performed a second iteration of the decision tree construction during which we removed snow cover ScWE as a possible predictor variable. We hypothesized that this would force the inclusion of any variables that influenced the loss or accumulation of ScWE on the hillslopes over winter and also after the last snow survey from which ScWE was calculated (*i.e.* those variables that controlled the transformation of total snowfall to snow cover). Two variables that were neglected

Table 3.4 Hierarchy of controls (ranked 1-6) on snowmelt-runoff generation, for the primary decision tree.

Hierarchy	Control
1	Total seasonal snowfall
2	Snow cover
3	Fall soil surface water content
4	Melt rate
5	Melt season length
6	Fall soil profile water content

in the original decision trees then became important: number of warm winter days, and mean temperature on warm winter days. One variable, thawed layer depth, did not appear in any version of the decision trees, which suggested that either it was not an important control on snowmelt-runoff ratio or it covaried with other variables, such as snow cover.

3.5.2 Condition-dependent hierarchy of controls

The primary decision tree showed that the overall hierarchy of controls was (in descending order of importance): total snowfall, snow cover, fall soil surface water content, melt rate, melt season length, and fall soil profile water content (Table 3.4). The secondary decision trees showed that the selection of controls and their hierarchy vary when the dataset is split into high or low expressions of those six key variables (Table 3.5).

For high and low snow cover years, the hierarchies of controls differed significantly from one another. When snow cover was low, the controls on runoff ratios were largely spring-related (melt season length, date of peak runoff, melt rate, and spring temperature gradient). By comparison, when snow cover was high, the top three controls on runoff ratio matched the overall hierarchy of controls, albeit with differing orders of importance. Further, under high snow cover conditions, fall soil surface water content played the most important role in controlling runoff ratios; whereas

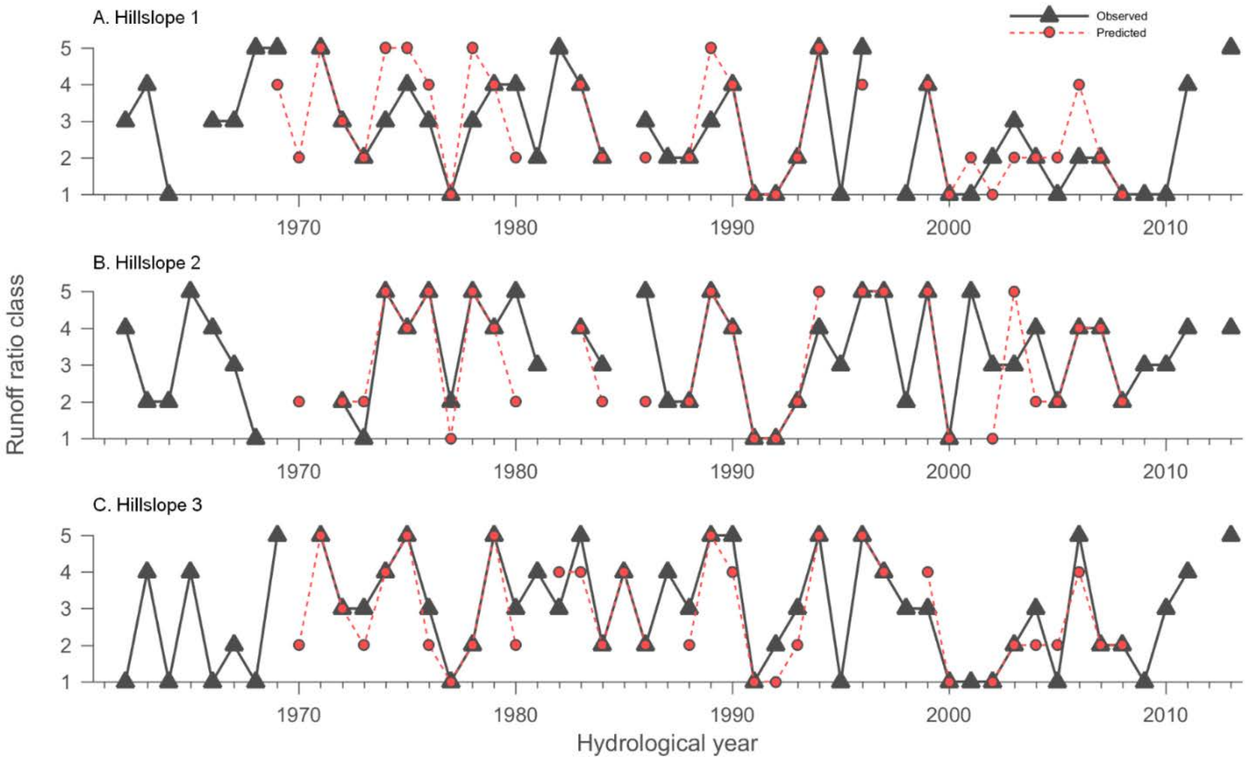

Figure 3.4 Time series of observed runoff ratio classes (response variable) and predicted runoff ratio classes (predicted by the primary decision tree) for A) Hillslope 1, B) Hillslope 2, and C) Hillslope 3.

under low snow cover conditions, fall soil surface water content was superseded by four other variables in controlling runoff ratios. The hierarchies of controls for high and low instances of total snowfall were relatively similar and retained three of the original six controls from the primary decision tree. They both had snow cover and fall soil surface water content as the top controls, and introduced surface depression storage as a main control.

For instances of low fall soil surface water content, the controls on runoff ratio were quite dissimilar from those in the overall hierarchy of controls. Snow cover, usually an important control, was not important here. Meanwhile, mean daily wind speed did exert a large influence on the prediction of runoff ratios. Further, neither total snowfall nor snow cover controlled runoff ratios under instances of low fall soil profile water content. This indicated that when soil water content was low throughout the soil profile, runoff ratio was not at all predictable based on precipitation amounts. On the other hand, the controls on runoff ratio for instances of high fall soil

Table 3.5 Condition-dependent hierarchy of controls (ranked 1-6) on snowmelt-runoff generation, for all secondary decision trees.

Total seasonal snowfall		Fall soil surface water content		Melt rate	
High	Low	High	Low	High	Low
1. Snow cover 2. Fall soil surface water content 3. Melt rate 4. Surface depression storage	1. Snow cover 2. Fall soil surface water content 3. Total seasonal snowfall 4. Date of peak runoff & Surface depression storage	1. Snow cover 2. Total seasonal snowfall 3. Fall soil surface water content	1. Total seasonal snowfall 2. Mean daily wind speed & Date of peak runoff 4. Surface depression storage	1. Total seasonal snowfall 2. Snow cover	1. Melt season length 2. Mean daily wind speed 3. Fall soil surface water content 4. Total seasonal snowfall
Snow cover		**Fall soil profile water content**		**Melt season length**	
High	Low	High	Low	High	Low
1. Fall soil surface water content 2. Total seasonal snowfall & Snow cover 4. TWI	1. Melt season length 2. Date of peak runoff 3. Snow cover 4. Melt rate 5. Fall soil surface water content 6. Spring temperature gradient	1. Snow cover 2. Total seasonal snowfall 3. Land cover	1. Mean daily wind speed 2. Fall soil profile water content	1. Fall soil profile water content 2. Fall soil surface water content 3. Snow cover 4. Surface depression storage 5. Date of peak runoff	1. Snow cover 2. Mean daily wind speed

surface water content and high fall soil profile water content were similar to one another and to those in the overall hierarchy of controls: total snowfall and snow cover were the top two. This indicated that when the ground was wetter than average at the surface and/or throughout the entire soil profile, then the runoff ratio was controlled by precipitation inputs.

When only high melt rate events were analysed, the identified controls on runoff ratio were both precipitation-related (total snowfall and snow cover). Even if a high melt rate was observed, low runoff ratios were still possible if total snowfall or snow cover were low. The highest runoff ratios

Table 3.6 Predictive accuracies for each runoff ratio (RR) class and for the overall tree, for the primary decision tree and each of the 12 secondary decision trees.

Dataset type	Predictive accuracies					
	RR class 1	RR class 2	RR class 3	RR class 4	RR class 5	Overall
Primary	84.6	52.5	100	60.6	85.3	70.1
High total seasonal snowfall	60.0	-	63.0	82.0	95.0	77.0
Low total seasonal snowfall	66.7	75.0	56.3	66.7	100	70.3
High snow cover	83.3	75.0	80.0	87.5	90.6	83.6
Low snow cover	67.5	83.3	62.5	65.0	-	72.2
High fall soil surface water content	100	50.0	-	60.0	91.7	78.7
Low fall soil surface water content	72.2	-	50.0	57.1	62.5	62.8
High fall soil profile water content	100	52.6	-	63.6	70.8	71.6
Low fall soil profile water content	77.8	75.0	100	-	-	82.6
High melt rate	40.0	45.8	-	-	50.0	45.3
Low melt rate	100	58.3	50.0	-	69.2	67.2
High melt season length	100	66.7	50.0	58.3	87.5	70.1
Low melt season length	66.7	46.2	-	-	71.4	61.4

occurred when there was a high snow cover and a high melt rate. By comparison, the predicted runoff ratios under instances of low melt rates were determined by a combination of different controls, with total snowfall at the bottom of the hierarchy. Even given low melt rates, high runoff ratios could still occur when there was a long melt period and high fall soil surface water content.

When we analysed those years which had short melt seasons, the runoff ratios were strongly controlled by snow cover. A short melt season coupled with high snow cover produced the highest runoff ratio. Otherwise, runoff ratio appears to have been controlled by mid-winter mean wind speeds. By comparison, the runoff ratios in those years with long melt seasons were controlled predominantly by soil water content (both profile and surface), followed by snow cover, surface depression storage and finally date of peak runoff. A long melt season coupled with either low soil water content or little snow cover typically led to very low runoff ratios.

3.5.3 Interactions between controls on runoff response

We have so far identified the controls, their hierarchy, and how that hierarchy varied under different conditions. This section outlines interactions between controls, *i.e.* whether one variable offset or promoted another variable in the determination of runoff response. While high fall soil water content was associated typically with high runoff ratios, and low soil water content was associated typically with low runoff ratios, these were sometimes mediated by other controls. For example, low runoff ratios could still occur with high fall soil water content if there was little snow cover on a hillslope with high surface depression storage. For winter-time variables, the amount of snowfall and snow cover both had clear effects on the runoff ratio result: typically, high amounts of SfWE or ScWE resulted in high runoff ratios, and low amounts of SfWE or ScWE resulted in low runoff ratios. However, these were mediated by other controls such that, in eight years, very low runoff ratios (runoff ratio class 1 and 2) resulted despite high snow cover. Finally, for spring-time variables, high runoff ratios occurred when melt rates were fast, and when the melt period was prolonged and late in the spring. This was especially apparent for high snowfall amounts in the primary decision tree (Figure 3.3).

For all decision trees, the lowest runoff ratios (class 1) occurred typically when there was either little snowfall or little snow cover on the hillslopes. In some rare occasions, these factors alone triggered low runoff ratios, despite competition from opposing factors that would typically promote high runoff ratios (*e.g.* high soil water content or high melt rate). In other instances, for low seasonal snowfall or snow cover to trigger low runoff ratios, they had to be associated with one or more factors that would also limit runoff ratios. These factors were: a) low fall soil surface water content; b) slow melt rates; or c) high surface depression storage. A very dry fall soil surface water content (< 0.11) was always associated with the lowest runoff ratio, regardless of other conditions.

The highest runoff ratios (class 5) always occurred when there was high snowfall or large snow covers. Typically, wet antecedent soil surface water content was also required. If soil surface water contents were low, the highest runoff ratios could still be generated when high snowfall and large snow covers occurred either: on a hillslope with little surface depression storage; or if the entire soil profile (not just the surface) was wet; or if the peak runoff occurred late in the season. For years when winter snowfall amounts were less than the long-term average, the highest runoff ratios could still be generated if soil surface water contents were not low, if peak runoff occurred late (at the end of March or into April), and, finally, if a large proportion of that SfWE was not retained in the snow cover during the winter (*i.e.* if a large proportion of SfWE was ablated over winter). This was echoed in a second iteration of the decision tree construction when snow cover was removed from the response variables, in which a condition for the highest runoff ratio to occur was when there were many mid-winter warm days. This indicated that the highest runoff ratios could only occur in low snowfall years when there were mid-winter melts that perhaps raised the soil surface water content or created an ice lens at the soil surface, thus reducing the infiltrability of the soil come spring melt.

Finally, the removal of snow cover from the possible response variables also triggered the inclusion of the land cover variable as an important predictor of runoff ratios; in that case, the highest runoff ratio occurred if the hillslopes were in fallow in the previous growing season. We also analyzed the period of continuous standing stubble on Hillslope 2 (1993-2004) separate from the remainder of the dataset (not, however, using the decision tree approach due to there being only 12 data points of continuous standing stubble). The proportion of the season's snowfall that was retained on the hillslope as snow cover (ScWE / SfWE) was significantly greater ($p < 0.01$) during the period of standing stubble on fallow (on average, a 0.65 retention) than not (on average, a 0.40 retention), likely due to the snow-trapping qualities of standing stubble. While the period of continuous standing stubble also was associated with, on average, wetter fall soil surface and soil profile water content and higher runoff ratios (0.391 compared to 0.254 for the non-standing stubble on fallow instances) there was no significant ($p > 0.05$) difference between the two groups of data.

3.6 Discussion

Our 52-year analysis of runoff, precipitation, soil water content, snow cover, and meteorological data showed little relationship between precipitation input and total seasonal runoff output. This highlights the extreme nonlinear relationship between precipitation inputs and runoff outputs on frozen ground at the hillslope scale. The additional factors that controlled runoff partitioning included total snowfall, snow cover, fall soil water content at the surface and through the soil profile, melt rate, and melt season length. Together, these explained 70% of the runoff ratio variance (Figure 3.4). We used the combinations of these factors, as dictated by the decision tree results, to account for the scatter in the relationship between total seasonal snowfall (SfWE) and total seasonal runoff (originally shown in Figure 3.2a, and now updated with colour-coded partitioning in Figure 3.5). Further, these factors were hierarchical and condition-dependent. For example, in years when snow cover water equivalent (ScWE) was high, fall soil surface water content played the most important role in controlling runoff ratio. In years when snow cover was low, fall soil surface water content was relatively unimportant and was trumped by four other variables.

3.6.1 Infiltration into frozen soil controls hillslope runoff ratio

Three groups of variables controlled snowmelt-runoff ratio: precipitation amount (represented by total snowfall and snow cover), antecedent wetness condition (represented by fall soil surface water content and fall soil profile water content), and melt intensity (represented by melt rate and

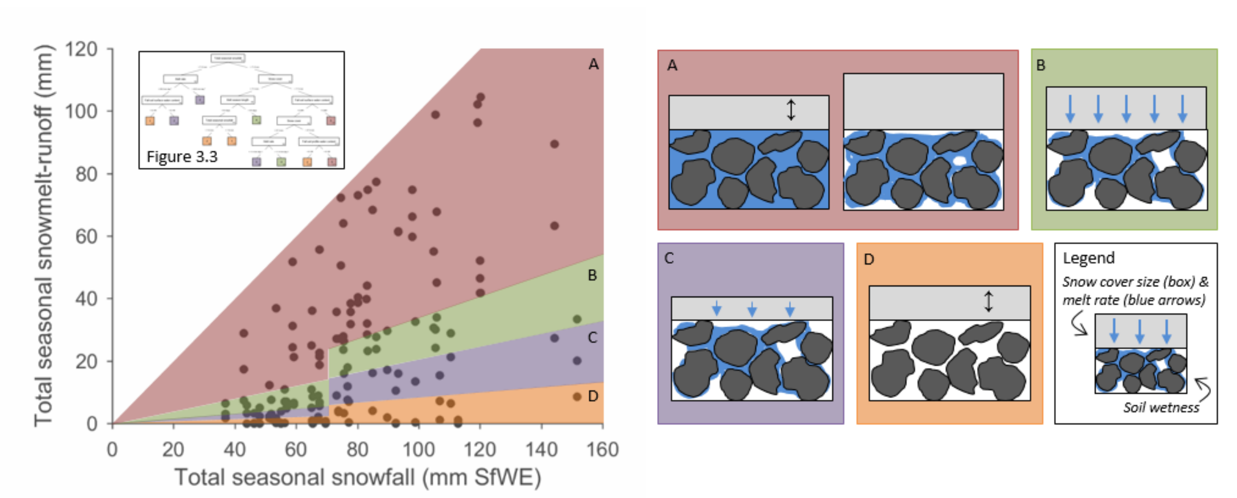

Figure 3.5 Partitioning of the relationship between total seasonal snowfall (mm SfWE) and total seasonal snowmelt-runoff (mm) (Figure 3.2a), showing typical characteristics of the snowmelt-runoff conditions in different parts of the plot. Different runoff ratios are colour-coded according to the leaf colours in Figure 3.3. Typical relative soil water contents for each partition are shown by the granular soil boxes (typically saturated or very wet in A, to very dry in D). Typical relative snow cover water equivalent (ScWE) are shown by the height of the overlying grey box (*e.g.* low ScWE in C, to high ScWE in A). A double-ended black arrow (↕) indicates that the ScWE depths were unimportant for that scenario (*e.g.* in D, runoff ratios are typically low regardless of the ScWE). Typical melt rates are shown by the height of the blue downward arrows (typically high melt rate in B, and typically low melt rate in C).

melt season length). Together, they determined collectively the balance between the fraction of precipitated water that infiltrated and the fraction that ran off and was delivered to the hillslope outlet. These six key variables controlled infiltration. While others have shown the importance of infiltration for snowmelt-runoff generation (Granger *et al.*, 1984; Fang *et al.*, 2007; Ireson *et al.*, 2013), our long-term analysis is the first to show the hierarchical importance of factors in controlling infiltration.

Infiltration capacity is known to change through an event (Horton, 1933). During a water input event where soil is saturated from above, infiltration capacity typically is high at the beginning of the event, followed by a rapid decline and asymptotic reduction (over minutes, hours, or days) to a near-constant value and quasi-steady-state (Zhao and Gray, 1997; Dingman, 2008). The decline in infiltration capacity is due to sorptivity – the potential of the soil to absorb and transmit water through capillarity. This is higher for dry than for wet soil. For water input to frozen soil, an additional factor linked to infiltration capacity decline is the re-freezing of meltwater in soil pores causing blockages (Ireson *et al.*, 2013). Nevertheless, the shape of frozen soil infiltration curves is similar to that of unfrozen soil (Kane and Stein, 1983).

Figure 3.6 shows a conceptual model of how each control in turn influences runoff ratio via the process of infiltration, where a constant melt rate that exceeds the infiltration capacity of the soil is assumed. The greater the precipitation event amount (where the event is the melt season and the precipitation amount is the depth of snow cover or the total seasonal snowfall) or the longer the melt season, the more of the declining infiltration capacity curve is traversed with time (along the x-axis). Figure 3.6b shows a scenario where a large melt event generates greater amounts of runoff, and greater runoff ratios because of the relative volume of meltwater that infiltrates (the integral of the infiltration curve) versus that which does not. Similarly, melt rate determines how much precipitation, at any point in time, exceeds the infiltration rate of the soil (y-axis of the curve). Figure 3.6c shows that a higher melt intensity exceeds the infiltration rate for a longer period of time, and produces greater runoff ratios. Finally, antecedent soil water content (and other soil characteristics) control the shape of the infiltration rate curve. Figure 3.6d shows that wet antecedent soil conditions cause reduced initial infiltration rates, which are more readily exceeded

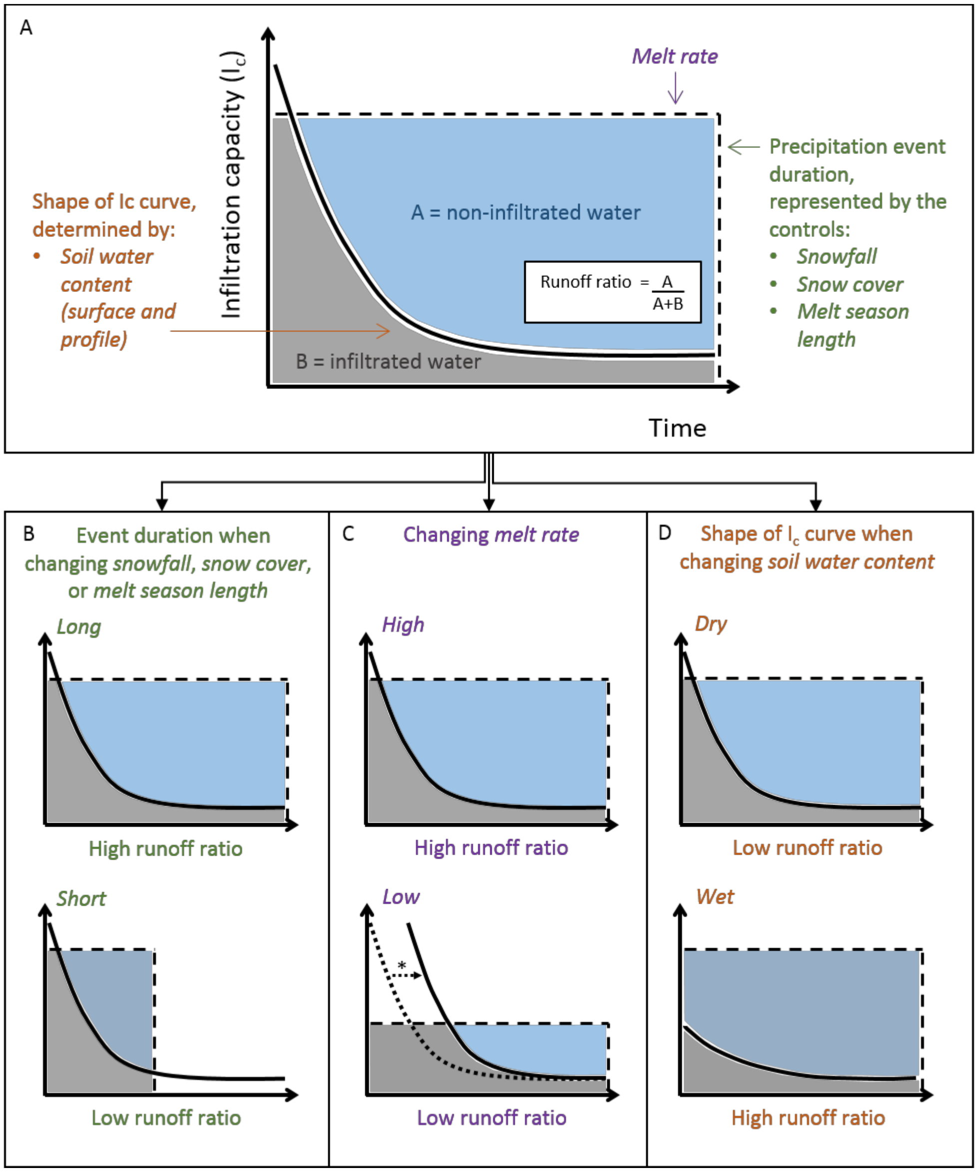

Figure 3.6 Conceptual figure showing how the key controls on runoff ratio affect runoff ratio via infiltration. A) Features of a typical infiltration capacity curve; B) snowfall, snow cover, and melt season length control event duration (green), with all other factors being equal; C) melt rate (purple) controls the incoming water flux; and D) soil moisture affects the shape of the infiltration curve (orange). When just one of these is varied (and the other are held constant), the volume of water that infiltrates (brown) and the volume that is in infiltration-excess (blue) changes, and it is the balance between these two that dictates the runoff ratio. The curve-shift (*) is a conceptual illustration of the moving curve method for Horton's infiltration model for instances where the volume of precipitation input is less than the infiltration volume.

by melt rates and thus lead to higher runoff ratios. This importance of antecedent soil water content for frozen ground infiltration and snowmelt-runoff response further demonstrates, beyond the findings of Coles *et al.* (2017) (Chapter 2), that the runoff response is not just about whether the ground was frozen or unfrozen (*i.e.* the soil temperature), but also is strongly related to the soil moisture conditions when frozen or unfrozen. The simple conceptual model of infiltration explains why these six variables – snowfall, snow cover, melt season length, melt rate, fall soil water content at the surface and through the soil profile – exert the greatest control on runoff ratio.

Varying snow cover and antecedent soil moisture conditions seemed particularly influential in causing shifts in the hierarchical ordering of controls. Under high snow cover conditions, fall soil surface water content played the most important role in controlling runoff ratios. By comparison, under low snow cover conditions, fall soil surface water content was superseded as a control on runoff ratio by four other, largely spring seasonally-related, variables. The infiltration rate curve helps explain why this might be the case (Figure 3.6). Given a decline in infiltration over time, for a small amount of snow cover, the infiltration curve is only traversed at the start (where infiltration rate changes quickly over time), so any change in melt rate exerts a large effect on the resultant runoff ratio. By comparison, for a large amount of snow cover, runoff ratios are less sensitive to changes in melt rate. This would explain why melt rate was identified as being a stronger control on runoff ratios under conditions of lower snow cover.

Our results showed that when fall soil water contents were low, runoff ratios were not predictable based upon the usual controls on runoff ratio. For example, runoff ratios when the entire soil profile was dry were not predictable based on precipitation amounts. Runoff ratios when the soil surface was dry were not predictable based on snow cover or the actual soil surface water content. This further indicates that infiltration into frozen soil, already a difficult flux to predict, was especially variable when the soil was dry in the fall. Further, while soil water content was the most important control on runoff ratios under conditions of high snow cover, it was a relatively unimportant control under conditions of low snow cover. Low snow cover at the end of winter could imply either little seasonal snowfall or significant ablation over the winter. Ablation might have caused over-winter changes in soil water content, thus rendering the measurement of the fall soil water content an imprecise or misleading representation of the pre-melt water content. This would in turn lead to its reduced importance in the decision tree. The snow cover SWE on the hillslopes was typically significantly less than the total seasonal snowfall, illustrating the importance of mid-winter ablation events (*e.g.* by melt, sublimation, or snow redistribution by wind) and supporting the findings of Gray *et al.* (1983; 2001) and Pomeroy *et al.* (2007a).

3.6.2 Soil moisture memory

Notwithstanding these likely mid-winter changes to soil water content, soil moisture memory from fall freeze-up was important. We observed that fall soil surface water content, measured immediately prior to temperatures falling below freezing and before the onset of snowfall, was the third key control on spring (four to six months later) snowmelt-runoff response. This is consistent with observations from more humid regions, where soil moisture memory has been analysed and described in the context of the persistence in the soil of anomalous wet or dry conditions that have long since been forgotten by the atmosphere (*e.g.* Entin *et al.*, 2000; Mahanama and Koster, 2003; Orth and Seneviratne, 2013). These studies, largely based on modeling approaches or long-term data analysis, have shown that memory in mid-latitude regions is strongest under extreme (particularly extreme dry) conditions (Wu and Dickinson, 2004). Memory timescales have been reported up to 2-3 months (Mahanama and Koster, 2003; Vinnikov *et al.*, 1996; Entin *et al.*, 2000). Also, soil and vegetation characteristics have been shown to be more important than the climate regime in determining the soil moisture memory strength (Orth and Seneviratne, 2013).

That fall soil water content in cold, snow-dominated seasonally- or permanently-frozen ground locations was a key control on spring runoff is not new; a common assumption of hydrological modeling approaches is that the soil water content at the start of snowmelt (typically March-April) equals the soil water content at the time of freeze-up (October-November). In other words, the water content is believed to be ‘locked in’ through the winter and remain constant until spring thaw, and thus exhibits soil moisture memory. Of course, we know this to not be necessarily true due to the likelihood of mid-winter melt events having caused increases in soil water content.

Further, vapor transfer across a soil-air or soil-snow interface, and mid-winter vertical redistribution of water within the soil towards the downward-advancing freezing front all challenge this memory assumption (Kane and Stein, 1983; Gray *et al.*, 1985; Quinton and Hayashi, 2008; Nagare *et al.*, 2012). Gray *et al.* (1985) suggested that in the absence of mid-winter melt events, soil water content in the 0-30 cm surface layer decreases over winter, with the greatest losses (only a 3-4% decrease, typically) seen for fallow (as opposed to stubble) lands. We hypothesize that this is due to further infiltration to deeper parts of the soil profile. We found that there was a strong positive relationship between the length of time during which the soil was frozen over winter and the runoff ratio in the following spring, with the relationship strongest for 50 cm and 100 cm depths (data not shown). This might indicate that a longer winter period of deep soil freezing prevented the soil surface water content from dissipating vertically, thus retaining the soil moisture at the surface, and driving a high runoff ratio come the spring. It might also indicate that faster infiltration rates and reduced runoff ratios – when the wetting front reached the thawed layer below the frozen layer (Watanabe *et al.*, 2012) – were not reached because the wetting front did not reach thawed soil when the soil was frozen to greater depths. The longer the soil was frozen and the deeper it was frozen, the stronger the soil moisture memory was. If the fall soil water content was high, this then drove higher runoff ratios in the spring.

Despite mid-winter melt events and land cover causing deviations between fall and pre-thaw soil water content, fall soil water content remained an important control on spring runoff ratios. This indicates long soil moisture memory in the system, which was heightened the longer the soil profile was frozen down to 100 cm. Soil moisture memory propagated through to runoff response. However, the soil moisture memory observed here, we believe, ought to be distinguished from existing descriptions of soil moisture memory from elsewhere (*e.g.* Entin *et al.*, 2000; Mahanama and Koster, 2003; Orth and Seneviratne, 2013). This is because those could be thought of as 'active' systems, while our observations are from a more 'dormant', frozen system. For this cold, seasonally-frozen region, soil moisture memory was less about the persistence of an anomaly, and more about dormancy. Hence, memory of the system was due to climatic conditions, more so than the soil and vegetation characteristics.

3.6.3 Implications for modeling and future field campaigns

Our findings have implications for existing approaches for predicting runoff responses to snowmelt events, especially regarding the ways existing models deal with infiltration and regarding what (and when) experimentalists ought to focus their observations on in the field. We do not believe that the nonlinearity and condition-dependent nature of these controls defies our ability to model, since we have shown here that it can largely be explained in the context of infiltration. Several physically-based approaches exist for determining infiltration into frozen soil, including models that solve heat and water transfers through porous media such as GeoStudio (GeoSlope International, 2015), SUTRA (Voss and Provost, 2002), SHAW (Flerchinger and Saxton, 1989), SOIL (Stähli *et al.*, 1999), and HYDRUS 1D (Hansson *et al.*, 2004), and models that use the pore size distributions and other physical aspects of the soil such as capillary bundle models (Watanabe and Flury, 2008). These approaches require considerable data to drive the energy and meteorological inputs, and to model the soil domain; much of which is not available here.

In hydrological studies on the Canadian Prairies, commonly implemented models include empirical equations for determining infiltration (INF), such as that of Granger *et al.* (1984):

$$INF = 5(1 - \theta_a)\, ScWE^{0.584} \qquad \text{(Equation 3.1)}$$

where θ_a is the pre-melt volumetric soil water content (-) in the 0-30 cm soil layer, and INF and snow cover water equivalent ($ScWE$) are in millimeters. This equation drives the infiltration module of the Cold Regions Hydrological Model (CRHM; Pomeroy *et al.*, 2007b), a widely implemented model for snow- and snowmelt-dominated regions (*e.g.* in Canada: Ellis *et al.*, 2010; Quinton and Baltzer, 2013; Fang and Pomeroy, 2008; in China: Zhou *et al.*, 2014; in Europe: Lopez-Moreno *et al.*, 2014). If we test this equation with our data to calculate its efficacy for determining the component of the water balance that infiltrates into frozen soil we need to rearrange Equation 3.1 to calculate runoff ratio (RR) as we defined it in our analyses above, using the total seasonal snowfall ($SfWE$):

$$RR = 1 - \frac{5(1 - \theta_a)\, ScWE^{0.584}}{SfWE} \qquad \text{(Equation 3.2)}$$

According to Granger *et al.* (1984), two controls determine snowmelt-driven infiltration and runoff ratio: snow cover and soil water content. Our results support that these two variables are indeed key controls on runoff ratio (in our case, these controls were ranked second and third, respectively, in the hierarchy of controls). However, using Equation 3.2 to predict our observed runoff ratios over the 1971-2013 period explained only 13.6% of the variance of the runoff ratio classes (Figure 3.7). It overestimated low runoff ratios, and underestimated high runoff ratios.

Therefore, this frequently-used equation for infiltration into frozen soil is of limited use, despite the fact that our decision tree findings show that infiltration into frozen soil is the main control on hillslope runoff ratio. This is not surprising since, based on our results, the determinants of runoff ratio are more complex. While a more mechanistic model is certainly needed to bring in these elements, doing this in a deterministic way would be the basis for future work. The results of this paper's decision tree learning – the key controls the decision tree has identified, the condition(s) under which each control are important, and the ways in which they interact and feedback between one another – could be a way to frame a model structure for snowmelt-runoff over seasonally-frozen hillslopes.

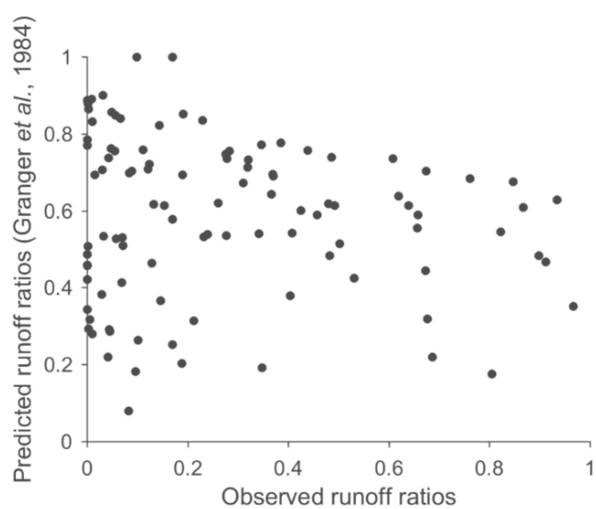

Figure 3.7 Observed and predicted runoff ratios over the 1971-2013 period. The predicted runoff ratios were calculated using Equation 3.2, based on Granger's *et al.* (1984) empirical equation for infiltration into frozen ground.

In terms of what to measure to parameterize new models, we can use our hierarchy of controls to guide cost-effective and useful field measurements. While seemingly obvious, this study has reinforced the need for reliable snowfall and snow cover data. It also has emphasized the critical need for measuring pre-freeze soil water content in the fall: surface observations (0-15 cm) are most important, but total soil profile water content would also be beneficial. The influence that mid-winter melt events appeared to have on soil water content and ice lens creation means that pre-melt soil water content should also be measured, if possible. This measurement is most important when the soil is relatively dry in the fall, when mid-winter melt events occur, or when there is a large amount of snow cover. However, accurate measurement of unfrozen and frozen water content in frozen soils in field conditions remains a problem, with probes (for example, dielectric instruments or gamma probes) requiring significant and complex calibration (Ireson *et al.*, 2013), and manual, sample-extraction methods proving very difficult given the frozen nature of the ground and the overlying snow. With any advances in instrumentation and methods for more reliable measuring of soil water content in frozen ground, these should be deployed to track antecedent moisture conditions through the winter and aid in the prediction of snowmelt-runoff response. While these fall-based and winter-based observations are most important for the prediction of runoff response, the spring conditions are of course also key to the response, especially when there is a small amount of snow cover. For predictive purposes, the melt rate and duration of the melt season need to be estimated in advance. For development, calibration, and/or validation purposes, these need to be documented.

3.7 Conclusions

We examined a 52-year dataset of runoff, precipitation, soil water content, snow cover, and meteorological data to determine the hierarchy of controls on snowmelt-runoff generation. Our decision tree analysis showed that the most important controls on snowmelt-runoff generation were, in descending hierarchical order of importance: total snowfall, snow cover amount, fall soil surface water content, melt rate, melt season length, and fall soil profile water content. Together, these were able to account for the scatter in the precipitation-runoff relationship. The hierarchy of these controls was controlled by actual conditions, with the biggest hierarchical differences between high and low snow cover seasons, and wet and dry antecedent conditions.

The key variables determining the runoff ratio collectively reflected the controls on the fraction of precipitated water that infiltrated. Despite the possibility of mid-winter changes in soil water content, the system tended to show significant memory in that the soil water content in the fall was a strong control on runoff in the spring. Here soil moisture memory was mostly determined by system dormancy and less so by the persistence of an anomaly. This distinguishes soil moisture memory in our system from that in more humid regions. An existing commonly-used two-parameter method for predicting infiltration into frozen soil (Granger *et al.*, 1984) predicted just 13.6% of runoff ratio variance, compared to 70% predicted by the five-parameter decision tree approach here. This would suggest that there is potential for a new or amended empirical model with improved predictability. Our results showed field-based measurements for estimating snowmelt-runoff response must include pre-freeze soil water content (primarily at the surface but also through the entire soil profile, if possible), snowfall and snow cover water equivalents, pre-melt soil water content to account for any over-winter changes in the soil water content, and, through the spring snowmelt season, melt rate and melt season duration.

3.8 Transition statement

Chapter 3 built upon the long-term trends analysis of Chapter 2, and found that the controls on long-term runoff were hierarchical and condition-dependent. Both studies observed that factors affecting infiltration into frozen soil were key determinants of the long-term runoff response and condition-dependent variability. These hinted at the importance that patterns of infiltration-excess surface water might have for hillslope-scale connectivity dynamics within a single snowmelt season. Therefore, these findings provided a multi-decadal context to the intensive field campaign, the focus of Chapter 4, in which I measured the spatial patterns of the controls on runoff. This was to determine the mechanisms behind hillslope-scale connectivity.

3.9 Acknowledgements

We thank the many Agriculture and Agri-Food Canada researchers, technicians, and students in Swift Current who have collected these data over the last 52 years. In particular, we acknowledge

Don Reimer and Marty Peru who were the responsible technicians for 1971-1992 and 1993-2011, respectively. The data analysis work was supported by an NSERC Discovery Grant awarded to JJM.

3.10 Author contributions

BGM provided the long-term dataset. AEC, WMA, and JJM brainstormed on data analyses and data mining methods. AEC designed and carried out the decision tree approach and carried out the analyses. AEC wrote the paper. WMA, BGM, and JJM commented on the manuscript and contributed to the text in later iterations.

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CHAPTER 4

FILL AND SPILL DRIVES RUNOFF CONNECTIVITY OVER FROZEN GROUND

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4.1 Abstract

Snowmelt-runoff processes on frozen ground are poorly understood at the hillslope scale. This is especially true for hillslopes on the northern Great Plains of North America where long periods of snow covered frozen ground with very shallow slopes mask any spatial patterns and process controls on connectivity and hillslope runoff generation. Here, we examine a 5 ha hillslope on the northern Great Plains during the 2014 spring snowmelt season to explore runoff processes at the hillslope scale. Specifically we explore the spatial patterns of runoff production source areas and examine how patterns of soil water content and thawed layer depth affect melt water partitioning and lateral delivery to the hillslope base. We explore if the controls on connectivity – where 'connectivity' is conceptualised as the generation of continuous flow fields across a hillslope – are consistent with the fill and spill mechanism found elsewhere in rain-dominated and unfrozen soil domains. We measured soil water content, thawed layer depth, snow cover, and snow water equivalent on a 10 m grid on our 5 ha hillslope. We also measured snow, soil water, ponded water, and hillslope runoff stable isotope composition during the spring snowmelt season. The contrast between the slow infiltration rates into frozen soil and the relatively fast rates of snowmelt delivery to the soil surface resulted in water accumulation in small depressions under the snowpack. Consequently, infiltration was minimal over the 12 day melt period. Instead nested filling of micro- and meso-depressions was followed by macro-scale spilling. This spilling occurred when large patches of ponded water exceeded the storage capacity behind downslope micro barriers in the surface topography, and flows from them coalesced to drive a threshold-like increase in runoff at

the hillslope outlet. These observations of ponded water and flowpaths followed mapable fill and spill locations based on 2 m resolution digital topographic analysis. Interestingly, while surface topography is relatively unimportant under unfrozen conditions at our site because of low relief and high infiltrability, surface topography shows episodically critical importance for connectivity and threshold-like runoff generation when the ground is frozen.

4.2 Introduction

Understanding snowmelt-runoff generation in cold, snowmelt-dominated regions is critical for predicting water delivery and water availability as the climate changes and these cold regions lose their cold. While point-scale (*e.g.* Granger *et al.*, 1984; Zhao and Gray, 1999) and hillslope-scale (*e.g.* Kane *et al.*, 1981; Quinton and Marsh, 1999; Carey and Woo, 2001; Quinton *et al.*, 2004; Suzuki *et al.*, 2005; Woo *et al.*, 2008) melt and runoff processes have been well studied, we still do not fully understand process controls on hillslope snowmelt-runoff connectivity and threshold-like runoff. Indeed, connecting point-scale runoff generation elements across hillslopes and catchments is now seen as a fundamental challenge for assessing the nonlinearities in runoff relations. Several studies have now shown how key nonlinearities like thresholds and feedbacks can produce emergent behaviour that is not explainable by traditional point-scale concepts (Grayson and Blöschl, 2001; Sivapalan, 2005; Bracken and Croke, 2007; James and Roulet, 2007; Troch *et al.*, 2008; Ali and Roy, 2009; Bracken *et al.*, 2013; McDonnell, 2013). In rainfall-runoff studies, pattern-based or spatially-distributed measurements have enhanced our understanding of hydrological connectivity and associated thresholds as linked to surface or bedrock topography (Darboux *et al.*, 2002; Tromp-van Meerveld and McDonnell, 2006) and soil moisture (Western *et al.*, 2001; Penna *et al.*, 2011).

Spatially-distributed approaches have led to or supported the concept of fill and spill (Spence and Woo, 2003; Tromp-van Meerveld and McDonnell, 2006) as a potential underlying mechanism for emergent threshold behaviour in runoff generation (McDonnell, 2013). The fill and spill mechanism posits that storage capacities (*e.g.* depressions) in subsurface or surface topography must fill up to a certain threshold (*e.g.* the downslope sill of the depression) before it can spill downslope. Fill and spill has now been used to account for: runoff from soil-filled valleys in which valley physiography has created various segments of varying storage conditions (Spence and Woo, 2003); along the bedrock of upland humid forested hillslopes (Tromp-van Meerveld and McDonnell, 2006; Hopp and McDonnell, 2009); and along an impeding layer of shallow humid forested hillslopes (Du *et al.*, 2016; Jackson *et al.*, 2016). In peat-dominated, permafrost environments, filling and spilling of spatially variable storage above a frost table has been shown to generate hillslope subsurface flow (Wright *et al.*, 2009) and surface runoff connectivity (Williams *et al.*, 2013). Catchment-scale overland flow generation analogous to fill and spill have been observed in lake- and wetland-dominated landscapes (Leibowitz and Vining, 2003; Shaw *et al.*, 2012; Leibowitz *et al.*, 2016). Surface overland flow studies at the hillslope scale have shown that runoff is modulated by micro-topography and surface roughness (Darboux *et al.*, 2002; Appels *et al.*, 2011; Chu *et al.*, 2013). While not labelled *sensu stricto* as 'fill and spill', they too are examples of overland flow being driven by the filling and spilling of depressions at a partitioning surface with loss along the flowpath and threshold behaviour at the larger scale (Ameli *et al.*, 2015).

The fill and spill mechanism fits within a storage-excess framework of water delivery (Spence, 2010; McDonnell, 2013). Existing runoff concepts are somewhat limited in geographic relevance; for example, the variable source area theory (Hewlett and Hibbert, 1967) typically only explains runoff generation in humid, vegetated sites, while the partial area concept (Betson, 1964) is restricted to more arid, infiltration-excess overland flow systems (McDonnell, 2013). While fill and spill is not a theory *per se,* McDonnell (2013) suggested that it represents a framework that could guide field measurements that map and describe the storages, connectivity, and thresholds relationships for a given site, and lead to new theory linking the similarities of runoff processes – one that is related to storage, storage thresholds, and connectivity (Spence, 2010).

While evidence now abounds linking storage exceedance and emergent threshold behaviour with the fill and spill mechanism, relatively few studies have observed such a mechanism in frozen environments (Spence and Woo, 2003; Wright *et al.*, 2009; Williams *et al.*, 2013). No studies that we are aware of have examined whether or not such a mechanism operates over seasonally-frozen ground on the well-drained, glacial deposits of the northern Great Plains of North America. Melt onto frozen ground is notoriously difficult to model (Gupta and Sorooshian, 1997; Pomeroy *et al.*, 2007). On the northern Great Plains, the focus of this study, this is especially difficult due to minimal topographic slope and deep, permeable soils. Upscaling point-scale frozen ground runoff measurements (*e.g.* Granger *et al.*, 1984; Zhao and Gray, 1999) to the hillslope scale has been difficult. For instance, Coles *et al.* (2016) (Chapter 3) tested the widely-used infiltration model of Granger *et al.* (1984) over 52 years of snowmelt-runoff recorded at the Swift Current hillslopes in Saskatchewan and found that the point-scale model was able to explain only 13.6% of the hillslope-scale meltwater runoff ratio.

Runoff in the melt season on the northern Great Plains is typically infiltration-excess overland flow over frozen ground (Fang *et al.*, 2007). Natural drainage systems at the landscape scale in the region are poorly developed, disconnected and sparse, due to the aridity and exceptionally low angled topography (Fang *et al.*, 2007). The snowmelt season sees, on average, a third of the annual precipitation melt within 1-2 weeks to generate *c.* 80% of the annual runoff. At the hillslope scale we might expect that these factors would encourage sheet-like overland flow across the soil surface. At larger basin scales in these regions, many non-contributing areas exist from which there is no water routing to a main drainage system even under extremely wet conditions (Stichling and Blackwell, 1957; Martin *et al.*, 1983). At the hillslope scale, shallow slopes and a lack of defined drainage system can lead to a large non-contributing proportion of the hillslope.

While undulations in the frozen soil surface could be enough for some spatial flowpath organization or generation of non-contributing areas, these need to be mapped and addressed. Critically too, the contrast between frozen ground infiltrability and snowmelt input rates dictate whether overland flow is generated or not – these are rarely mapped or reported. If the contrast is large enough it may enable widespread filling and spilling and whole-hillslope connectivity. If the contrast is too small (less than 10^1 as noted in modeling studies by Hopp and McDonnell (2009) and James *et al.* (2010)) then it would encourage infiltration, loss along a flowpath, and diminished or negated connectivity.

While frozen, a soil's infiltrability is usually less than its unfrozen state (Granger *et al.*, 1984), but these frozen infiltrabilities are varied and can sometimes still be significant (Burt and Williams, 1976), especially if the soil had a low pre-melt water content. For example, Spence and Woo (2003) observed infiltration rates of 41 mm hr^{-1} on the subarctic Canadian shield regardless of whether the unsaturated soil was frozen or not. Fang *et al.* (2007) observed greater infiltration than runoff on frozen, agricultural fields in southern Saskatchewan due to dry soils from the previous year's cropping. Zheng *et al.* (2001) measured cumulative infiltration over 90 minutes into thawed soil and soils frozen to various depths, and found that thawed soil infiltration (65.6 mm) was only 19.1% greater than infiltration into shallow frozen soils (55.1 mm). Snowmelt rates, too, are highly variable. Spence and Woo (2003) noted that melt input intensity averaged 0.11 mm hr^{-1} and melt water readily infiltrated their relatively high infiltrability frozen soils. At the Swift Current hillslopes, Coles *et al.* (2016) (Chapter 3) found that season-averaged snowmelt rates over the last 52 years of record have varied between 0.39 and 4.63 mm hr^{-1}.

Here we explore the factors controlling the patterns and mechanisms of hillslope meltwater runoff on seasonally-frozen ground of the northern Great Plains, specifically at the Swift Current hillslopes. We build upon long term analysis at this site (Coles *et al*, 2016, 2017; Chapters 2 and 3) but focus on the 2014 melt season. We measure the spatial patterns of snow cover, snow water equivalent, soil water content, frozen ground, and topography to understand the primary controls and processes behind hillslope-scale runoff activation. We seek to understand the role of micro-, meso-, and macro-topography in controlling the snowmelt-runoff response and explore the similarities and differences in comparison to processes observed in warmer and/or more sloping regions (*e.g.* Darboux *et al.*, 2002; Appels *et al.*, 2011; Chu *et al.*, 2013; Tromp-van Meerveld and McDonnell, 2006). We combine these hydrometric observations and mapping of spatial patterns with isotope analysis of snowmelt inputs and runoff outputs to quantify the 'newness' of snowmelt-runoff over frozen ground. Specifically, we address the questions:

i) What are the spatial patterns and process controls on connectivity and hillslope runoff generation over frozen ground?
ii) How do patterns of soil water content and thawed layer depth affect melt water partitioning and lateral delivery to the hillslope outlet?

iii) Are the controls on connectivity consistent with the fill and spill mechanism found elsewhere?

4.3 Study site

The Swift Current hillslopes are three adjacent agricultural hillslopes located at South Farm, Swift Current in southern Saskatchewan in the Canadian Prairie region of the northern Great Plains. The study site has been described previously in Coles *et al.* (2016, 2017) (Chapters 2 and 3), which undertook long-term analyses on all three hillslopes. Here, we focus our high resolution spatial analysis on Hillslope 2 (Figure 4.1), the central of the three hillslopes, with an area of 4.66 ha. A raised, grassed berm around the perimeter of the hillslope prevents flowing water from entering from adjacent land and ensures that the only outlet for runoff is an instrumented H-flume at the hillslope's northwest corner. The hillslope is relatively concave in shape and has a gradient of 1% in the upper two-thirds of the hillslope and a gradient of 2.5% in the lower one-third, sloping towards the northwest. A digital elevation model (DEM), obtained using a Leica Viva GS15, is available for the hillslope at a 2 m horizontal resolution. At a finer scale than the 2 m resolution of

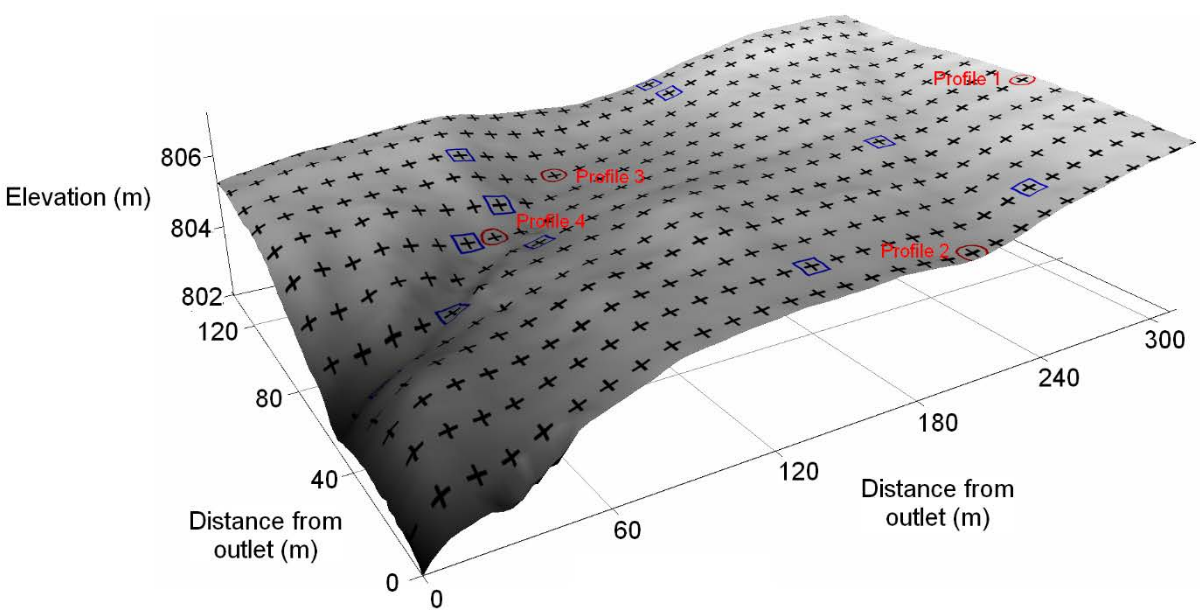

Figure 4.1 Digital elevation model of the surface topography of Hillslope 2, with vertical exaggeration. Also shown: 10 x 10 m measurement grid (plus signs), locations of the four soil moisture and soil temperature profiles (red circles), and locations of the 11 snowmelt lysimeters (blue squares).

the DEM are micro-topographic features – ridges and furrows – from seeding in the growing season. If the ground is not tilled following harvest, these features remain through the winter and into the spring. This was the case in the spring of 2014, the season studied here. The hillslope was ploughed and seeded in early summer 2013 in a north-south direction, with wheat planted in the raised ridges (*c.* 5-15 cm in elevation above the neighbouring furrow, and the cross-section of one ridge and one furrow being approximately 30-50 cm wide).

The soil is a Swinton silt loam and classified as an Orthic Brown Chernozem (Cessna *et al*., 2013). Archived data from a one-location soil profile investigation, with data for 15 cm intervals from the soil surface down to 180 cm, were provided by Agriculture and Agri-food Canada for the site. They show that silt content decreases with depth from 50.4% in the 0-15 cm surface layer to 27.9% at 150-165 cm, clay content increases with depth from 18.2% to 30.9%, and sand content fluctuates between 24.4% and 42.4% through the profile. There is a clay layer (48.4% clay) observed at a depth of 165-180 cm (which presumably prevented deeper investigations). Bulk density increases with depth, from 1.22 g cm^{-3} at 0-15 cm to 1.59 g cm^{-3} at 120-135 cm, below which it decreases slightly to 1.43 g cm^{-3}. Observations also show that saturated hydraulic conductivity increases with depth: from 1.42 cm hr^{-1} at 0-15 cm to 5.76 cm hr^{-1} at 75-90 cm (no data deeper than 90 cm).

Prior to the 2014 melt season (in July-August 2013), we characterized the spatial variability of soil depth and surface infiltration capacity. Soil probing with a dynamic cone penetrometer (also known as a knocking pole penetrometer; Shanley *et al.*, 2003) at 17 random locations on the hillslope revealed the mean soil depth to be 265 cm (s = 45.3 cm). In general, resistance to penetration patterns remained relatively uniform in space for the upper 200 cm of the soil profile, but increased with depth below 200 cm until refusal. In most of the 17 profiles, resistance also increased sharply at approximately 15-20 cm depth, for a layer approximately 5-10 cm thick. Another layer of resistance (of varying thickness between 5-20 cm) was observed in most profiles between approximately 60-100 cm below the soil surface. A third, thin (*c.* 5-10 cm thick) layer of resistance exists in some of the profiles between 120 and 200 cm below the soil surface, which likely reflects the clay layer identified in the archived soil profile data. Infiltration capacity measurements have been undertaken with a constant head sprinkler infiltrometer at 62 random locations on Hillslope 2 (Seifert, 2014). They show unfrozen infiltration capacities to range between 0.4 and 63.5 mm hr^{-1}, with a mean of 13.9 mm hr^{-1} and standard deviation of 13.2 mm hr^{-1} (Seifert, 2014). Snowmelt-runoff laboratory experiments with intact soil cores extracted from Hillslope 2 showed frozen surface infiltration capacities at this site are much lower: they range from 0.09 to 2.57 mm hr^{-1}, with a median of 0.33 mm hr^{-1} (Coles *et al.*, 2017; Chapter 2).

Hillslope 2 is under agricultural management with typically an annual rotation of wheat (*Triticum aestivum* L.) and fallow, but with some instances in the last 52 years of grass (*Psathyrostachys juncea* (Fisch.) Nevski), lentils (*Lens culinaris* L.), and peas (*Pisum sativum* L.). Hillslope 2 has undergone both conventional tillage and zero tillage practices. In 2013, the year prior to our field campaign in spring 2014, Hillslope 2 was cropped with wheat and had been under zero tillage management. As a result, from September 2013 to May 2014 (encompassing the snowmelt period studied here) Hillslope 2 had standing wheat stubble residue of variable stubble height of 30-50 cm. Precipitation data (measured using a Belfort weighing gauge) for the period of study were available from a nearby (*c.* 700 m to the south-southeast) Environment and Climate Change Canada standard meteorological station.

4.4 Methods

We used digital topographic analysis, specifically the calculation of two metrics (flow accumulation and downslope index), to develop a theoretical map of fill and spill locations across Hillslope 2. We then conducted high spatial and temporal resolution measurements of key hydrometric variables to explore the changing spatial patterns of runoff production source areas. We combined high-frequency monitoring of runoff rates at the hillslope outlet with stable water isotope analysis of the runoff, snowmelt, and soil water, and with the hydrometric spatial maps to understand the drivers of connectivity and threshold-like water delivery during the snowmelt season. We used the digital topographic analysis' map of fill and spill locations to assess whether our field observations of the controls on connectivity were consistent with the fill and spill mechanism. These steps are outlined in greater detail in the following subsections.

4.4.1 Digital topographic analysis

Following Hopp and McDonnell (2009), we calculated two metrics for each cell of the 2 m DEM cells. The first metric calculated was flow accumulation (FA), which indicated the upslope contributing area of each cell, calculated as the number of cells upslope that drained into each cell. This was determined using the D8 flow algorithm, a common tool to determine the weighting of flow from each cell into the eight adjacent cells (Jenson and Domingue, 1988). The FA also indicates local topographic highs (ridges or sills), which are assigned an FA of 0.

The second metric calculated was the downslope index (DI), which indicated the downslope drainage efficiency of each cell. DI was expressed as DI = V/H, where H is the horizontal distance that must be traversed in the steepest downslope direction to descend to a point at a pre-defined vertical distance (V) from the elevation of the starting cell (Hjerdt *et al.*, 2004). While the DI was initially used to capture near-surface groundwater levels and 'backing-up', it is thought to be a useful tool in different terrains where topographic curvature exerts a control on local drainage regimes (Hjerdt *et al.*, 2004). For calculating DI, we used a V of 15 cm. This value was chosen because it is the maximum elevation change between a ridge and furrow – artefacts of mechanised seeding, where one ridge and one furrow has *c.* 5-15 cm in elevation difference and a cross-section that is approximately 30-50 cm wide. This ensures that any sporadic instances of a ridge or furrow being picked up in the 2 m horizontal resolution DEM are smoothed out from this topographic analysis. A cell with a small DI was caused by a long horizontal distance (H) and indicates that drainage from that cell was slow and inefficient (Tromp-van Meerveld and McDonnell, 2006; Hopp and McDonnell, 2009).

We used the combination of FA and DI as an indicator of potential fill and spill locations across the hillslope (Hopp and McDonnell, 2009). Fill locations – areas where water can be collected and retained – were designated when cells had a large FA (> 10 m^2) and small DI (< 0.015) (typically shallow, long slopes). Spill locations – areas where water can accumulate and then be efficiently drained – were designated when cells had a large FA (> 10 m^2) and large DI (>0.015) (typically steep, short slopes). These threshold of FA (10 m^2) and DI (0.015) are their median values. For DI, a value of 0.015 represents approximately the general hillslope gradient.

4.4.2 Hydrometric field measurements

We measured volumetric soil water content at 0-6 cm depth on a 480-point grid (a 10 x 10 meter spatial resolution; Figure 4.1) during several daily field campaigns (19th July 2013, 2nd August 2013, 9th August 2013, 3rd September 2013, 23rd September 2013, 24th October 2013, 28th March 2014, 7th April 2014, 13th May 2014, and 19th June 2014) using a portable Stevens HydraProbe POGO. Snow cover and frozen ground prevented these measurements being taken over winter. Therefore, the last soil water content mapping prior to freeze-up (24th October 2013) was used to capture the spatial variability in soil water content at the onset of frozen conditions. Mapping resumed on 28th March 2014 once there was no longer snow cover and the soil was thawed sufficiently for the probe to be inserted. For each survey, we made further soil water content measurements at smaller spatial resolution, within random 10 x 10 m grid squares, for geostatistical analysis. Variogram analysis following the first soil water content survey (19th July 2013) showed that the variance of the data stabilized at approximately an 80 m resolution, giving us confidence that the use of a 10 x 10 m spatial resolution was adequate to capture the variability and spatial patterns.

We measured volumetric soil water content and temperature for five depth intervals (0-6, 6-15, 15-30, 30-60, and 60-90 cm) at four locations using Stevens HydraProbes. Each location was representative of a key landscape unit on the hillslope – upland area (Profile 1), two surface depressions (Profiles 2 and 3), and a slope (Profile 4) (Figure 4.1). These measurements were logged continuously at 30 minute intervals for a period of 12 months (October 2013 to September 2014). Only soil water content data for time intervals when the soil temperature was > 0°C is useful for our analysis. This is because the sensors determine soil water content via dielectric permittivity, which is not applicable to the quantification of water content in frozen, freezing, or thawing soils without considerable uncertainty (Williamson, 2016). We used the data to assess the change in soil water content from pre-freeze up to post-melt, with data used for these from 24th October 2013 and 24th April 2014, respectively.

We measured snow depth, and calculated density and snow water equivalent (SWE), on a 225-point grid (a 10 x 20 meter spatial resolution in the lower two-thirds of the hillslope, and a 10 x 40 meter spatial resolution in the upper third of the hillslope) by manual snow surveys. These snow surveys were carried out several times through the winter and just prior to the onset of snowmelt, and then daily, every morning before any significant melt, through the snowmelt period (9th March – 20th March 2014). We calculated daily ablation at the 225 points, and the hillslope-average, using Equation 4.1:

$$ablation_{\,day(x)} = SWE_{\,day(x)} - SWE_{\,day(x+1)} \qquad \text{(Equation 4.1)}$$

We also used eight 2 meter long ablation lines on the hillslope (positioned in each of the four key landscape units) to measure snow depth, and calculate density, SWE and ablation, on a daily basis at 20 cm intervals. At a sub-daily scale, we measured snowmelt rate from the base of the snowpack manually using 18 snowmelt lysimeters at 11 locations (with 1-3 duplicates at some locations) at irregular time intervals (10-120 minutes) depending on melt rate (Figure 4.1).

We measured surface thawed layer depth (depth to the top of the frozen ground) daily on a 60-point grid (a 20 x 40 meter spatial resolution) by manually knocking in a length of 11 mm diameter rebar until frozen ground resistance was detected (this was always undertaken by the same researcher for consistency). This was also carried out at 2 hour intervals at three locations to capture sub-daily changes in frozen ground depths.

Seven time-lapse standard-image cameras (Wingscapes) captured snow cover accumulation and ablation, and were used with personal observations to chronicle the snow covered area, and locations of water sources, flowpaths, and ponded water on the hillslope. Finally, runoff from the hillslope was logged at 15 minute intervals through the snowmelt period using a pressure transducer (HOBO U20 Water Level Data Logger) in the stilling well of an Agriculture and Agri-food Canada H-flume at the outlet of the hillslope.

4.4.3 Isotope sample collection and analysis

Stable isotope analysis of water is one more tool that we employed to understand the mechanism of hillslope-scale runoff generation during the melt season. We used it to determine the ratio of 'new' snowmelt water to displaced 'old' soil water in hillslope-scale runoff. During the 2014 snowmelt season, we collected 1422 water, soil, and snow samples. These samples consisted of:

- 308 runoff samples from the flume at the outlet of Hillslope 2, collected using an ISCO 3700 which automatically sampled water flowing through the flume at 30 minute intervals (15 minute intervals during peak flow).
- 454 snowmelt samples from the base of the snowpack, manually extracted from each of the 18 snowmelt lysimeters, and taken at irregular intervals (10-120 minutes) depending on melt rate (*i.e.* the approximate length of time it took to obtain a full 25 ml sample vial of water).
- 50 soil samples, collected prior to snowmelt on 20th February 2014 from two depths (0-6 cm and 6-15 cm) at 32 locations on Hillslope 2 using a slide-hammer corer. These were taken to obtain pre-event soil water, which, along with the snowmelt water, is an important potential end-member in the runoff signature from the plots.
- 217 snow core and incremental snow samples, collected bi-daily during the snow survey, melted down and bottled.
- 63 ponded water (on the soil surface or snow surface) samples, collected several times per day from any areas of ponded water.

All bottled water samples were sealed and stored in a non-refrigerated, cool and dry location. The soil samples were double-bagged and frozen until it was possible to extract the soil water from them. We extracted the soil water by high pressure mechanical squeezing (Orlowski *et al.*, 2016). The isotopic compositions of the liquid water samples were then determined by analysis on a Liquid Water Isotope Analyzer (Los Gatos Research) and reported in parts per thousand (‰) relative to VSMOW (Vienna Standard Mean Ocean Water), a standard of known composition.

4.4.4 Spatial patterns mapping

For all sets of data for each spatially-measured variable (surface soil water content, depth of thawed layer, and snow cover ablation), we interpolated the data points using kriging (Sarma, 2009) to provide gridded data for each variable at exactly the same points. We used ordinary point kriging with a linear variogram model to weight the surrounding measured values to derive a predicted value for an unmeasured location. We used the cross-validation method using all measured values to determine the quality of the gridded data. Maps were generated from the kriged datasets using the software Surfer® (Golden Software).

4.5 Results

4.5.1 Digital topographic analysis

To understand the potential effects of topographic features (Figure 4.2a) on surface runoff from Hillslope 2, we assessed the flow accumulation (FA) metric as an indicator of flowpath organization. We then combined this metric with the downslope index (DI) as an indicator of potential fill and spill locations across the hillslope. FA exhibits a power law distribution, whereby a histogram of the data extends from 0 to 4000 m^2 with the majority of the cells having a FA < 20 m^2 and a long tail of data from 20-4000 m^2. We truncated the mapping of FA to < 100 m^2. The FA map (Figure 4.2b) shows that individual flowpaths with higher FA are distributed across the whole hillslope, including in the upper reaches of the plot. There are five flowpath systems, all draining in a northwest direction towards the outlet of the hillslope. Two drain the lower third (2.5% slope) of the hillslope, and three drain the upper two-thirds (1% slope) of the hillslope. Flowpath system 1 is connected to the outlet by a thin flowpath on the upper west (left) border of the hillslope. However, flowpath systems 2, 3, 4, and 5 do not appear to be connected directly to the outlet: all four are separated to some extent from the outlet by cells of lower (lighter) flow accumulation, and also by the flowpath systems downslope (flowpath system 5 drains through 3 and then 1; while flowpath system 4 drains through 2 and then 1). Any surface ponding or retention of water would occur at the mouths of these individual systems. We calculated the area and volume of depressions on the hillslopes using the “Fill” tool in ArcGIS. This highlighted five areas of ponding, in line with the above analysis, at the mouths of the individual drainage systems (Table 4.1).

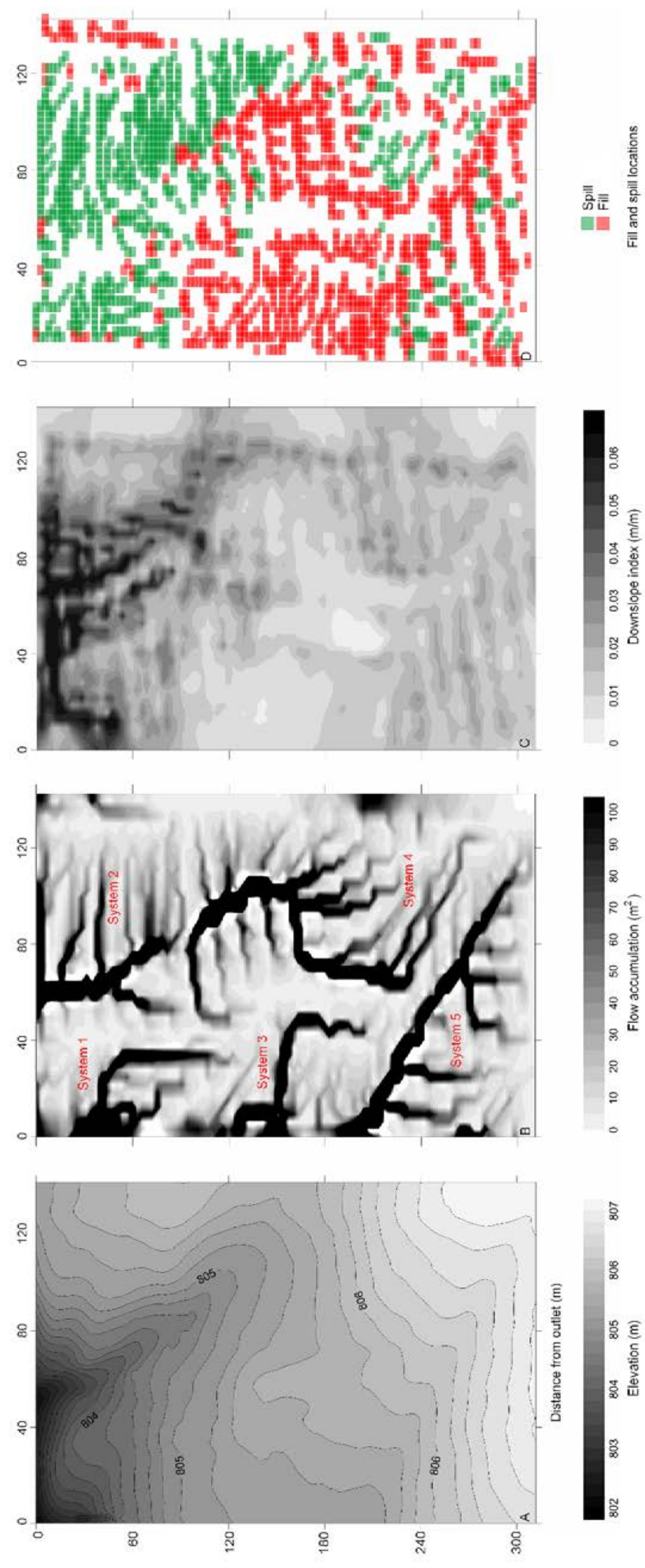
A
B
C
D
System 1
System 2
System 3
System 4
System 5
Distance from outlet (m)
Elevation (m)
Flow accumulation (m2)
Downslope index (m/m)
Spill
Fill
Fill and spill locations

Figure 4.2 (previous page) Maps of (A) the DEM of the surface topography, as in Figure 4.1; (B) flow accumulation (FA), with five sub-hillslope flowpath systems identified (drainage systems 1-5); (C) downslope index (DI); and (D) fill and spill locations, where fill locations are defined as having FA>10 and DI<0.015, and spill locations are defined as having FA>10 and DI>0.015.

The raised grass berms also influenced water ponding at the mouths of flowpath systems 2, 3, and 5. The western raised grass border impedes flow in the north-westerly direction from flowpath systems 3 and 5, causing it to pond between the raised border and a natural topographic sill on the hillslope. Downslope of the mouths of flowpath systems 3 and 5, the western raised grass border acts as a funnel for runoff down the western side. Similarly, the northern raised grass border impedes flow in the northerly direction from flowpath system 2, and also acts as a conduit for flow towards the hillslope outlet along the northern edge of the hillslope.

The DI map (Figure 4.2c) indicates that the majority of the hillslope, most notably the upper two-thirds, has a low DI and therefore a low drainage efficiency. An area of high DIs and therefore high drainage efficiency exists in the lower third of the hillslope and reflects the valley-like surface topography (Figure 4.2a). Combining FA and DI indicates the balance of fill and spill across the hillslope (Figure 4.2d). 29.8% of the hillslope is designated as either a fill location or a spill location, according to the thresholds of FA and DI that we used. The majority (57.4%) of the designated cells are fill locations, with these concentrated in the upper two-thirds of the hillslope. The spill locations (42.6%) are primarily in the lower third of the hillslope, but there are a limited number of small spill locations in the upper third of the plot also. Fill locations appear to be relatively well connected to one another, especially in the central third of the hillslope, and are fed by the small numbers of spill locations further upslope. The fill locations are set back from the

Table 4.1 Area, depth (mean and maximum), and volume of depressions at the mouths of each flowpath system.

Location of depression	Area (m^2)	Depth (m)		Volume (m^3)
		Mean	Maximum	
Mouth of System 1	18	0.0775	0.128	1.40
Mouth of System 2	54	0.0328	0.0564	1.77
Mouth of System 3	27	0.0716	0.113	1.93
Mouth of System 4	45	0.0210	0.0499	0.943
Mouth of System 5	324	0.0593	0.140	19.2

hillslope outlet, with a broad swath of spill locations in the intervening area. This suggests that once the fill locations – in the upper two-thirds – spill, due to water input exceeding the surface detention storage capacity, the released water can be efficiently routed through the spill locations – over the lower third – and to the hillslope outlet. Not captured by the DEM or topographic analysis are the micro-topographic features – ridges and furrows (with one ridge-and-furrow pair being *c.* 5-15 cm in elevation, and 30-50 cm wide) – from seeding in the previous growing season.

4.5.2 Hydrometric analysis

Snowmelt, snow cover, and runoff

The 2014 melt season was characterized by high total winter snowfall (77.5 mm SWE, between 1st October 2013 and 31st March 2014), and large snow cover (78 mm SWE), and a low-medium runoff amount (25 mm). The 78 mm SWE of the snow cover on Hillslope 2 melted over 12 days, between 9th March 2014 and 20th March 2014, with peak snowmelt on 16th March (Figure 4.3a). Runoff from the hillslope began on 12th March and finished on 20th March, with peak runoff occurring, like snowmelt, on 16th March (Figure 4.3a,b). The peak runoff rate on 16th March was 11.6 times greater than the peak runoff rate on the previous day. The instantaneous, threshold increase in runoff after hillslope-wide connectivity was achieved at 15:00 on 16th March was 7.2 times greater than just before connectivity was achieved (13:45 on 16th March). There were four stages (Figure 4.3) in the evolution of meltwater inputs to the soil surface and runoff outputs from the hillslope: Stage 1 (9th – 12th March): initial snowmelt, but no resulting hillslope runoff; Stage 2 (13th – 15th March): continued snowmelt, with hillslope runoff generated; Stage 3 (16th March): high volumes of snowmelt and high runoff; Stage 4 (17th – 19th March): low snowmelt and small amounts of runoff.

The spatial patterns of snow cover ablation (Figure 4.3c) indicate that snowmelt occurred unevenly over the hillslope, with concentrated patches of snowmelt that changed in location over time. Daily snow cover ablation ranged between 0-70 mm over the hillslope. Sub-daily measurements at various snowmelt lysimeters across the hillslope showed that, at its peak on the afternoon of 16th March, snowmelt occurred at 1.17-8.21 mm hr^{-1}. We observed water ponding in the bottoms of

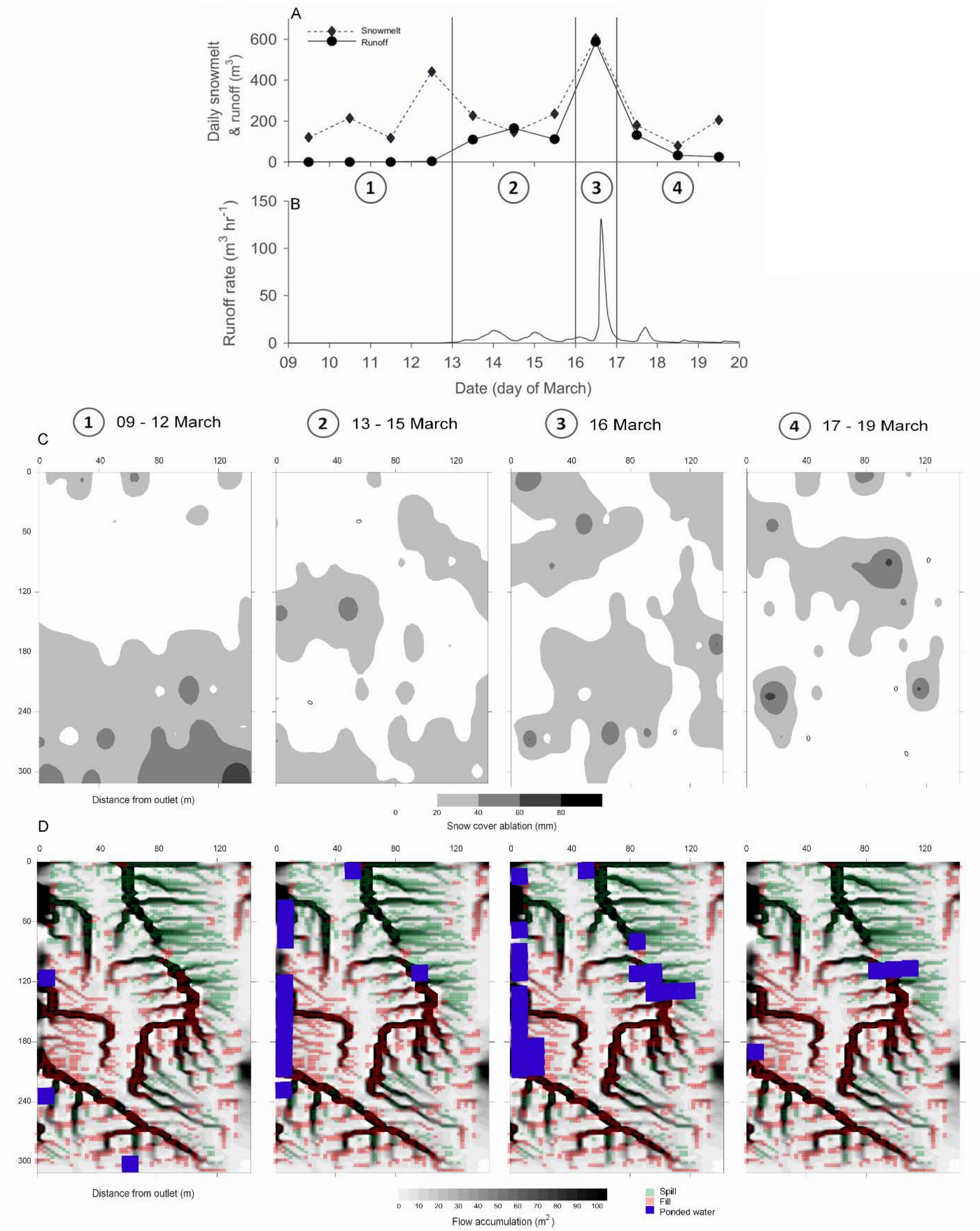

Figure 4.3 For the 2014 spring snowmelt season on Hillslope 2: (A) daily snowmelt and runoff volumes and (B) runoff rates, recorded at 15-minute intervals. Four stages (1-4) are identified in the snowmelt-runoff, referred to in the text. Maps relate to each of these stages: (C) snowmelt over the hillslope for each stage; and (D) ponded water locations during each stage overlaid on fill and spill locations and the flow accumulation map.

the micro-topographic furrows at the soil-snow interface, with flow along these features in the downslope direction. Maximum water movement was in the mid-afternoons, when some snowmelt lysimeters would become overwhelmed (and subsequently abandoned) by water flowing in from upslope. From 11th – 16th March, larger areas of ponded water gradually accumulated along the western edge of the hillslope (Figure 4.3d) (at the mouth of flowpath system 3 and 5), at the top of the valley-like system (at the mouth of flowpath system 4), and at the northern edge of the hillslope (at the mouth of flowpath system 2). Ponded water was at its maximum on 16th March, and decreased in extent in the following days.

Figure 4.4 shows the snow cover ablation and development of ponded water on the western edge of the hillslope. Snow covered area remained high over the hillslopes throughout the majority of the melt season, falling gradually from 100% on 11th March to 90.7% on 15th March, to 75.8% on 16th March (the day of peak melt and runoff), and then rapidly decreasing to 20.5% on 20th March. After the 20th March, a small amount of snow remained on the hillslope at the northern edge of the hillslope, which took another *c.* 6 days to clear.

Figure 4.4 Time-lapse photographs from the west side of Hillslope 2, facing northeast, showing the snow cover ablation and development of ponded water at the mouth of flowpath system 5.

Thawed layer depth

The ground was frozen at the soil surface (*i.e.* no thawed layer at the soil surface) homogenously across the hillslope, in the initial melt days, and during peak melt and peak runoff (16th March) (Figure 4.5a). Following peak runoff, the thawed layer deepened rapidly and unevenly (Figure 4.5b). On 18th March (two days following peak melt and runoff), the ground had thawed to depths of 15-20 cm below the soil surface, but only in isolated patches across the hillslope: most notably, on the upland at the south end of the hillslope, and on a southwest-facing slope in the valley-like feature at the north end of the hillslope. The ground remained frozen to the soil surface where snow cover still remained.

Soil water content

The 2014 melt season was preceded by a dry fall in 2013. On average, the soil water content in the surface 0-6 cm layer of Hillslope 2 was 0.15 (from the 24th October 2013 survey). We found that surface soil water content on 24th October 2013 (Figure 4.6a) showed relatively limited spatial variability, which is typical of all soil water content surveys conducted (Figure 4.6b). We can therefore assume that the soil water content at the onset of snowmelt was relatively spatially homogeneous. Further, because of long soil moisture memory in this frozen, dormant system (Coles *et al.*, 2016; Chapter 3), the pre-freeze up soil water content (Figure 4.6a) is likely representative of soil water content at the onset of the 2014 snowmelt season.

Soil water content generally increased following snowmelt under all four main landscape units (Profiles 1-4) (Table 4.2). The soil water content change was highly spatially variable. The soil profile situated in the shallow sloping, upland region of the hillslope (Profile 1) had the smallest increase in soil water content, with a net increase of 8.67 mm added to the profile (with gains close to the soil surface, but losses at depth). This profile location is representative of the majority of the hillslope. By comparison, soil profiles situated in the depressions (Profile 2 and 3, at the mouths of flowpath systems 5 and 4, respectively) saw net increases of 80.3 mm and 121 mm, respectively. Finally, the soil profile on the slope (Figure 4.4), in the valley-like topography saw a net increase

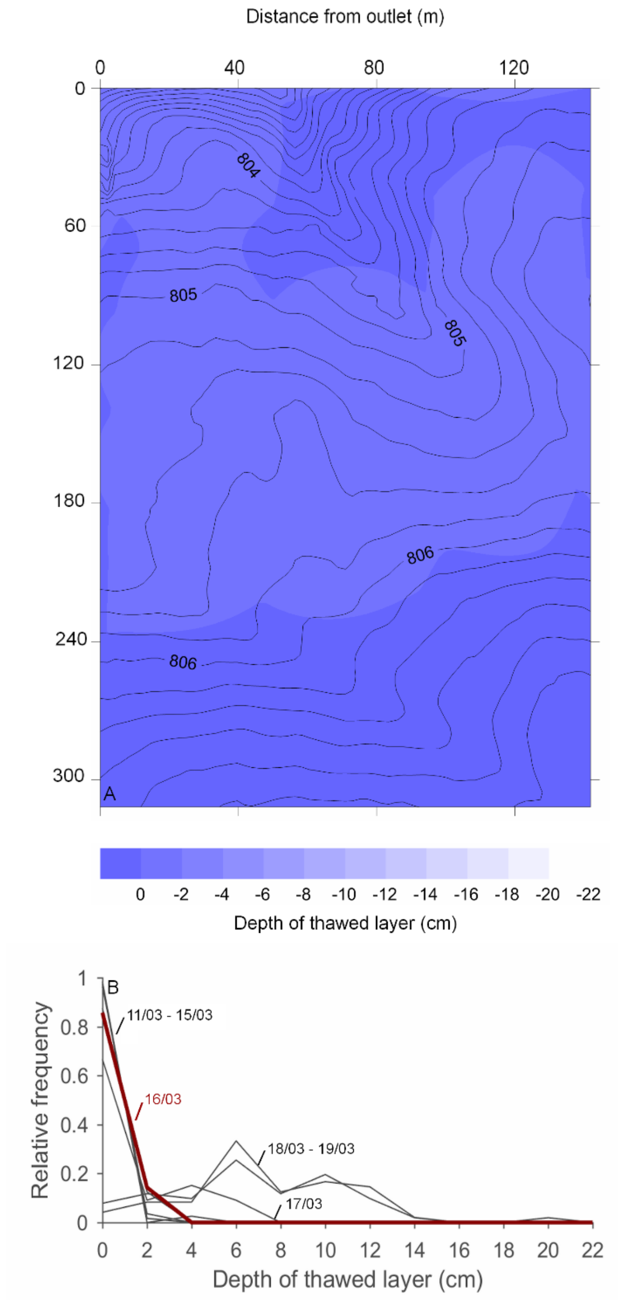

Figure 4.5 (A) Spatial map of thawed layer depth on the day of peak snowmelt and peak runoff (16th March 2014); and (B) frequency distributions of all thawed layer depth surveys conducted, with 16th March 2014 survey highlighted in red. Note that the colour scale of the spatial map (A) is the same extent as the x-axis of the frequency distributions in (B).

of 148 mm of soil water. Weighting the soil profile's water content change by their representative area suggests a hillslope-wide recharge of soil water (over the 0-90 cm depth profile) of 25 mm.

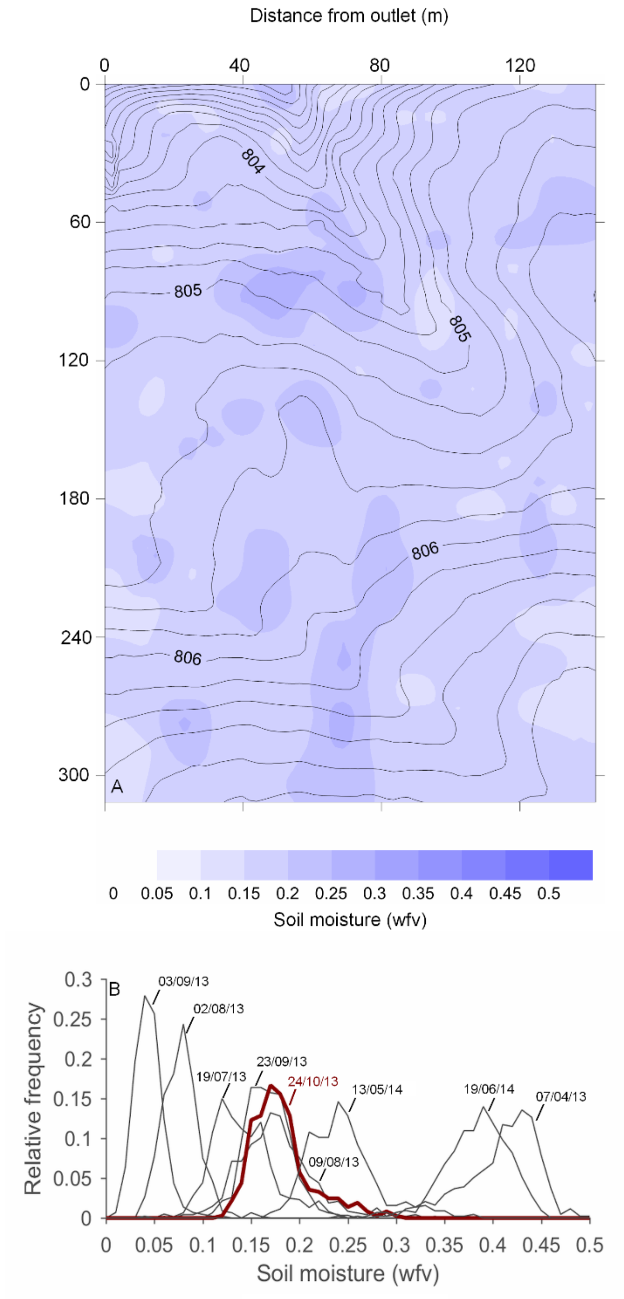

Figure 4.6 (A) Spatial map of pre-freeze up soil surface water content (0-6 cm), measured on 24th October 2013; and (B) frequency distributions of all soil water content surveys conducted, with 24th October 2013 survey highlighted in red. Note that the colour scale of the spatial map (A) is the same extent as the x-axis of the frequency distributions in (B).

4.5.3 Isotope analysis

Stable water isotope analysis shows that the δ^{18}O and δ^{2}H of runoff water was largely temporally-constant (on average, runoff water δ^{18}O was -21.7 ± 0.742‰, and δ^{2}H was -169 ± 4.56‰), but

Table 4.2 Pre-freeze (24th October 2013) and post-melt (24th April 2014) soil water contents at five depths at four key landscape units.

Soil profile	Depth interval (cm)	Pre-freeze soil water content		Post-melt soil water content		Change in water content (mm)
		vwc	mm	vwc	mm	
1 (upland)	0-6	0.134	8.04	0.182	10.9	2.88
	6-15	0.220	19.8	0.261	23.5	3.69
	15-30	0.167	25.1	0.193	29.0	3.90
	30-60	0.110	33.0	0.109	32.7	-0.300
	60-90	0.166	49.8	0.161	48.3	-1.50
2 (depression)	0-6	0.236	14.2	0.163	9.78	-4.38
	6-15	0.242	21.8	0.310	27.9	6.12
	15-30	0.224	33.6	0.316	47.4	13.8
	30-60	0.143	42.9	0.268	80.4	37.5
	60-90	0.144	43.2	0.235	70.5	27.3
3 (depression)	0-6	0.199	11.9	0.209	12.5	0.600
	6-15	0.207	18.6	0.316	28.4	9.81
	15-30	0.177	26.6	0.387	58.1	31.5
	30-60	0.150	45.0	0.319	95.7	50.7
	60-90	0.184	55.2	0.279	83.7	28.5
4 (slope)	0-6	0.196	11.8	0.266	16.0	4.20
	6-15	0.229	20.6	0.316	28.4	7.83
	15-30	0.158	23.7	0.339	50.9	27.2
	30-60	0.063	18.9	0.293	87.9	69.0
	60-90	0.070	21.0	0.202	60.6	39.6

with gradual enrichment through the melt season (Figure 4.7). The isotope signatures of the snowmelt water (on average, snowmelt water δ^{18}O was -22.3 ± 2.10‰, and δ^{2}H was -171 ± 15.4‰) bounded the runoff water, albeit with a high amount of variability. By comparison, the pre-event soil water was much more enriched than the runoff water (on average, soil water δ^{18}O was -14.6 ± 1.97‰, and δ^{2}H was -128 ± 11.0‰). Two-component hydrograph separation using the mean δ^{18}O or mean δ^{2}H soil water values for the pre-event end member, and the mean δ^{18}O or mean δ^{2}H snowmelt values from each lysimeter for the event end member, showed that the runoff water is primarily composed of 'new' snowmelt water, with very little mixing with the pre-event 'old' soil water. Regardless of the isotope (δ^{18}O or δ^{2}H) or which lysimeter we analyzed, the hydrograph separation showed that the runoff water was composed of 100% new snowmelt water in the initial stages of the snowmelt season. On the day of peak runoff (16th March), runoff water was composed of on average 93.9% event snowmelt water. Towards the end of the snowmelt and

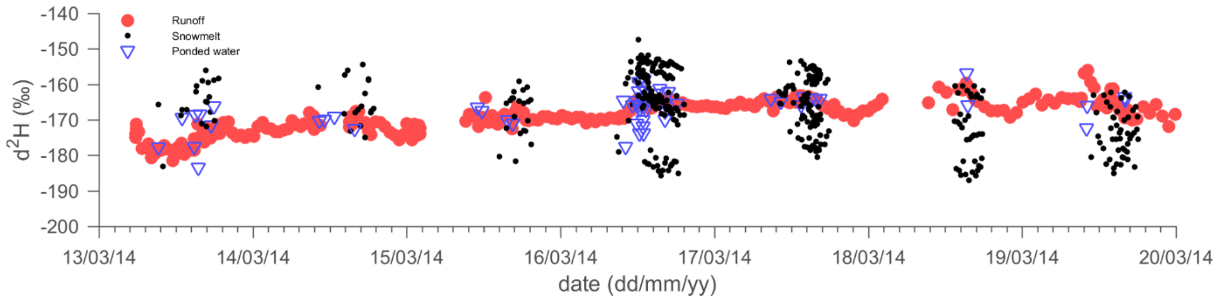

Figure 4.7 Time series of stable isotopes (for δ^2H) of runoff, snowmelt, and ponded water through the snowmelt season. Pre-event δ^2H soil water values measured on 20/02/14 (not shown) range between -112‰ and -140‰.

runoff season, on 19th March, this value had declined to on average 67.6%. Through the entire season, runoff water was approximately 95.1% event snowmelt water and 4.9% pre-event soil water.

4.6 Discussion

Our results suggest that a mechanism analogous to fill and spill explains the generation of snowmelt-runoff over frozen ground at our site. Our stable isotope analysis of meltwater confirmed that runoff water was 'event' snowmelt water with limited mixing with pre-event soil water. Unlike on steep terrain (Eriksson *et al.*, 2013), lateral flow through the flat-lying snowpack at our site was unimportant. The key factor generating fill and spill at our site is the large contrast between the low infiltration rates of the uniformly and fully frozen soil surface and the relatively fast rates of delivery of snowmelt water to the soil surface. This was enough to generate ribbons and ponds of water beneath the snow at the soil surface that accumulated in micro- and meso-topographic depressions, and then spilled downslope. Our observations of ponded water and flowpaths were consistent with mapped predictions of fill and spill activity from high resolution digital topographic analysis. We enunciate these features in the following sections.

4.6.1 Micro-, meso-, and macro-scale topographic controls on fill and spill

Thawed layer depth across the slope showed uniformly frozen soil (to the soil surface) in the days leading up to, and during peak runoff. Frozen ground infiltration capacities have been observed at this site to range between 0.09 – 2.57 mm hr^{-1}, with a median of 0.33 mm hr^{-1} (reported in the long terms analysis of Coles *et al.*, 2017; Chapter 2). Snowmelt rates during peak snowmelt in 2014 were 1.17 – 8.21 mm hr^{-1}. The relatively high rates of delivery of snowmelt water to the soil surface largely exceeded the infiltration capacity of the frozen soil by a magnitude greater than 10^1, which is the hypothetical minimum contrast between bedrock and soil permeabilities required to generate runoff via the fill and spill mechanism at the soil-bedrock interface in Hopp and McDonnell (2009). As such, we had an impeding layer contrast that was apparently sufficient for the retention and accumulation of water on the soil surface.

The concavity in the monitored hillslope is a macro-topographic feature (>10,000 m^2) of the site. The concavity appears to affect the balance of fill and spill: the relatively steeper 2.5% slope section in the lower third of the hillslope was a dominant spill location. The flatter 1% slope section in the upper two-thirds was a dominant fill location. Our terrain analysis indicated a balanced fill-spill regime with 57.4% of the hillslope characterized by 'fill' locations, and 42.6% of the hillslope characterized by spill locations. At finer scales, the meso-topographic features (100 – 10,000 m^2) revealed five flowpath systems in the flow accumulation (FA) mapping. Flowpath systems 1 and 2 drained the lower third of the hillslope; flowpath systems 3, 4, and 5 drained the upper two-thirds. These latter flowpath systems terminated at their downslope edges by slight barriers or 'lips' in the surface topography. These lips were enough to create a backwater effect and to create a fill region. We observed that the lips must be overcome for the upper region of the hillslope to connect to the lower, spill region of the hillslope and thus the hillslope outlet. The FA, DI, and fill and spill maps (Figure 4.2), where these lips are visual, are useful tools to interpret the mechanisms behind ponding and threshold-delivery of water. Important, though, is that each cell of these maps is solely an indicator of the local topographic surface in the near-horizontal distance. Additional research could seek to incorporate a metric for the likelihood of flow pathways being disconnected by a fill location created by a sill. Possible approaches to this could be to experiment with increased values of V (DI=V/H), which would then integrate topography further downslope from the starting point, or to use a metric of flowpath distance to the hillslope outlet and the fill locations it must overcome.

Nested within these macro- and meso-scale topographic systems, the micro-topographic features (< 100 m^2) also exhibited a flow control during meltwater runoff. These small undulations were observable within the 10 x 10 m measurement grid. Most notably, these were ridges and furrows left behind from tractor-based seeding in the previous summer. While these micro-scale features were not picked up by the 2 m DEM, they were an important localized feature in the initial routing and retention of melt water. Melt water pooled in the furrows and were gradually routed downslope in the micro-scale furrows within each of the five flowpath systems. In flowpath systems 1 and 2, fill and spill occurred mainly within the furrows. In flowpath systems 3, 4, and 5, however, the routing of water via these furrows and small undulations was overtopped by ponded water that developed and grew in volume upslope from the lips, after which these barriers were overcome and water could spill over and coalesce at the hillslope outlet. Overall, the hillslope exhibited nested filling and macro-spilling.

Our finding that topography dictates hillslope-scale connectivity and snowmelt-runoff generation over frozen ground is in contrast to Devito *et al*. (2005). They examined a boreal plain site with more surficial geology variation, but importantly with similar low relief and deep glaciated substrate as the Swift Current hillslopes. Devito *et al.* (2005) dismissed the importance of surface topography. The key difference, though, is that their evaluation was for a summer period when the ground was unfrozen and the deep, high-infiltrability, mineral soils promoted vertical flow infiltration. Indeed, an analysis of summer rainfall-runoff events at our site would support the suggestion that topography is unimportant, since all water infiltrates except in exceptional storms (only 28 years of the 52-year record have had summer storms generate runoff, as reported by Coles *et al.*, 2017; Chapter 2). But critically, topography is episodically important during meltwater runoff on frozen ground as shown in our work. During melt onto frozen ground, topographic features acted as both a conduit for meltwater runoff (enabled flowpath formation and connectivity once threshold surface detention levels were exceeded) and a loss mechanism (enabled ponded water to form, and then heightened infiltration and soil water recharge under depressions when the ground started to thaw).

4.6.2 Fill and spill over shallow, frozen hillslopes in relation to other environments

The hydrologic response on Hillslope 2 of the Swift Current hillslopes reflected the fill and spill mechanism already observed in many other environments (Spence and Woo, 2003; Leibowitz and Vining, 2003; Tromp-van Meerveld and McDonnell, 2006; Wright *et al.*, 2009; Graham and McDonnell, 2010; Appels *et al.*, 2011; Du *et al.*, 2016; Jackson *et al.*, 2016; Leibowitz *et al.*, 2016). The fill and spill mechanism was first introduced in a subarctic soil-filled valley with spatially-variable subsurface storage capacities, due to varying soil depths to bedrock, that had to fill up in order to enable surface runoff (Spence and Woo, 2003). The definition of the mechanism was further developed following analogous observations that showed depressions in subsurface or surface topography must fill up to a certain threshold (the downslope sill of the depression) before water can spill downslope (*e.g.* Tromp-van Meerveld and McDonnell, 2006; Leibowitz *et al.*, 2016). These fill and spill observations fall within a storage-excess framework of water delivery (Spence, 2010; Sayama *et al.*, 2011; McDonnell, 2013). The observations presented in this paper are fundamentally the same as those observations of fill and spill of depressions across an impeding layer, and of a storage-excess delivery of runoff.

The particular fill and spill mechanism described here is different to most previous observations, primarily because it is snowmelt over a frozen soil surface. This environment sees months of runoff inactivity with no whole-hillslope connectivity, and then 1-2 weeks where fill and spill over frozen ground delivers the large annual runoff pulse. This short, acute period of runoff occurs with the concurrent conditions of a frozen soil surface and high volumes of liquid water, as also described in Williams *et al.* (2013) for intermittent surface runoff connectivity over frozen peatland. This is unlike the humid, temperate regions where bedrock fill and spill is primed and relatively frequently produces subsurface stormflow (Tromp-van Meerveld and McDonnell, 2006; Graham and McDonnell, 2010; Du *et al.*, 2016; Jackson *et al.*, 2016).

The scale at which we have observed the fill and spill mechanism is different to most previous studies. Observations of this mechanism have typically been at the plot or trench scale (*e.g.* Tromp-van Meerveld and McDonnell, 2006; Wright *et al.*, 2009; Graham and McDonnell, 2010; Du *et al.*, 2016), at the small catchment-scale (Spence and Woo, 2003), and at the landscape scale (*e.g.*

the connected wetlands of Leibowitz *et al.*, 2016). The nested filling and spilling across scales at our hillslope site is essentially the next scale down from the wetland filling and spilling described for the prairie pothole region of the northern Great Plains (Leibowitz and Vining, 2003; Shaw *et al.*, 2012; Leibowitz *et al.*, 2016). The hillslopes of the northern Great Plains deliver water to these wetlands, whose connectivity is in turn also dictated by fill and spill, albeit a fill and spill mechanism that is influenced and mediated by additional factors such as groundwater-surface water interactions (Brannen *et al.*, 2015) and storage memory (Shook and Pomeroy, 2011).

Our study site is a low gradient end member (slope 1-2.5%) in the fill and spill literature. Our analysis indicated a fairly balanced fill-spill regime with 57.4% of the hillslope characterized by 'fill' locations, and 42.6% of the hillslope characterized by 'spill' locations. This is in contrast to previous studies with slightly steeper (yet still relatively shallow in the literature) slopes (a virtual 7.2% slope in Hopp and McDonnell, 2009; and the measured 6-12% slopes in Du *et al.*, 2016) that exhibited fill-dominated regimes, which was attributed to their 'flatness'. As slope angle decreases, hillslopes appear to transition from a spill-dominated (on steep slopes) to a fill-spill balance (on medium slopes) and finally to a fill-dominated system (on shallow slopes) (Hopp and McDonnell, 2009; Reaney *et al.*, 2014). Our fill-spill balance is more typical of medium-angled slopes. We attribute the difference between our fill-spill balance and these other low-angle studies' fill-dominated regimes to the difference in overall hillslope form. While these other studies' hillslopes were largely planar, ours is concave. The downslope barriers at the edge of the concave cross-section created the surface depressions that retained water and dictated upslope fill locations. Downslope of these barriers, any runoff at a point was able to flow unimpeded to the outlet.

Most prior fill and spill observations at the soil-bedrock or soil-argillic interface have reported connectivity as discrete flow networks – almost channel like in their flow architecture (Tromp-van Meerveld and McDonnell, 2006; Hopp and McDonnell, 2009; Graham and McDonnell, 2010; Williams *et al*, 2013). By contrast, our fill and spill connectivity across this frozen hillslope occurred as a set of more amorphous ponds that intermittently and individually connected to the hillslope outlet (analogous to wetland to wetland connectivity on the Prairies; Leibowitz and Vining, 2003). These areas of ponded water exhibited heightened infiltration and soil water recharge beyond that exhibited by the general, gently-sloping hillslope area. This is akin to enhanced groundwater recharge observed under bedrock depressions at the soil-bedrock interface (Appels *et al*., 2015).

Finally, an important difference between our findings and those at most other sites is the spatial variability of the precipitation input. The high spatial and temporal variability in snowmelt at our site is in contrast to rainfall over a similar small area, which would likely be relatively spatially constant and therefore its input distribution not important for rainfall-runoff connectivity modeling. The four distinct stages in the evolution of meltwater inputs to the soil surface and runoff outputs from the hillslope can be explained by the pattern, rates, and interaction of snowmelt on the frozen, reduced-infiltrability soil, and the pattern and layout of the micro-, meso-, and macro-topographic features: Stage 1 (9th – 12th March): the first amounts of snowmelt gradually accumulated in the furrows across the hillslope, with no resulting hillslope runoff. Stage 2 (13th – 15th March): snowmelt was continuing to accumulate in the furrows and be routed through each of the five flowpath systems. In flowpath systems 1 and 2, snowmelt water was then able to flow uninterrupted to the outlet, which generated the first hillslope runoff and low hillslope runoff ratios. Meanwhile, ponded water was accumulating behind the downslope barriers at the mouths of flowpath systems 3, 4, and 5. Stage 3 (16th March): high volumes of snowmelt caused the water ponding at the mouths of flowpath systems 3, 4, and 5 to reach capacity and spill over their downslope barriers. This connected the upper region of the hillslope with the lower region, and created continuous flowpaths connecting all five flowpath systems to the hillslope outlet with a threshold-like increase in runoff and high runoff ratios. Melt rates were highest in the upper region of the hillslope, which ensured the downslope depressions were continually fed, their barriers exceeded, and hillslope-wide connectivity maintained for 3-4 hours. Following this, the ponded water fell below the downslope barriers and disconnected the upper two-thirds of the hillslope from the outlet. Stage 4 (17th March onwards): low runoff was from slower, prolonged snowmelt, routed via micro-topography to the hillslope outlet from the remaining snow cover in the sheltered coulees over the lower third of the hillslope. The ground rapidly began to thaw from 17th March, enabling the remaining ponded water in depressions to readily infiltrate and contribute to soil water recharge. Overall, these four stages exhibited a dynamic contributing area, which is a feature of connectivity and fill and spill (Martin *et al.*, 1983; Shaw *et al.*, 2011). The contributing area was largely restricted to the lower third of the hillslope. It briefly extended to the entire hillslope on 16th March when the ponded water was connected to the hillslope outlet, before contracting back again to the lower region of the hillslope.

4.6.3 Soil moisture based metrics of connectivity perform poorly for frozen hillslopes

Soil water content is critically important for soil infiltrability and hillslope runoff generally (*e.g.* Horton, 1933), and over frozen ground on the Prairies (Granger *et al.*, 1984; Zhao and Gray, 1999) and at this site in particular (Coles *et al.*, 2016; Chapter 3). Despite this, we suggest that, because the soil water content showed very little spatial variability, the spatial patterning of soil water content likely had little effect on the spatial variation in ponded water development and flowpath distribution. The measured mean fall surface soil water content for the hillslope was 0.15. If the soil was on average much drier at the time of freezing, we likely would have seen greater hillslope-wide infiltration, more time for surface depressions to fill and then spill (if at all), and a delayed and damped threshold-delivery of water when connectivity was achieved. The opposite would have been true for a much wetter hillslope. Coles *et al*. (2016) (Chapter 3) saw some evidence of this in the 52-year dataset at the Swift Current hillslopes where runoff ratios were generally higher over wetter soils (likely a result of reduced infiltration). However, this was mediated by the volume of surface depression storage such that runoff ratios were typically lower when there was a high surface depression storage even when the soils were wet (Coles *et al.*, 2016; Chapter 3).

Metrics that use the spatial arrangement of hillslope or catchment soil moisture as indicators of connectivity – because of the way stores of water fill up to generate hydrological connections (Tetzlaff *et al.*, 2011; Bracken *et al.*, 2013) – are likely not helpful for these frozen soils where there is little spatial variation over the hillslope. We also observed that, for what little measured variability there was in soil water content, it was not related to topographic position. This could be attributed to the relatively low relief, and the influence of evapotranspiration in reducing the variability across the hillslope. At a similar prairie site, Peterson (2016) also observed that soil water content was not correlated with topographic relief. They also noted that soil water content variability was much higher under wetter conditions, yet still not related to topographic position (Peterson, 2016). Soil water content might have an effect on ponded water development and flowpath formation if a more undulating site froze very soon after rainfall (where side slopes might freeze dry and swales might freeze wet). We have not observed such effects, however. Analyses that use terrain to infer soil moisture and by extension flowpaths and connectivity (*e.g.* Beven and Kirkby, 1979; Lane *et al.*, 2009) may hold some promise in the determination of frozen ground flowpaths, but likely only due to structural routing of the water, rather than any topographically-induced differences in soil moisture. For example, our testing of the topographic wetness index (TWI; Beven and Kirkby, 1979):

$$\text{TWI} = ln(a/\,tan\,b) \qquad \text{(Equation 4.2)}$$

where a is the flow accumulation area per unit of contour width, and b is the local topographic gradient (Figure 4.8), unsurprisingly produced results very similar to the FA map (Figure 4.2b). The TWI metric does not incorporate any metric for downslope impedance, which we have shown here – with the use of the downslope index – to be an important component in the routing of flow and connectivity via fill and spill.

Previous work at this site has shown that a lumped approach can indeed be fruitful for predicting the seasonal runoff response (*e.g.* the decision tree model of Coles *et al.*, 2016; Chapter 3). We have shown here, though, that in order to understand and predict sub-seasonal time-scale (daily, hourly or weekly) runoff responses then distributed topographic data, distributed snowmelt data, frozen soil infiltration capacity data, and hillslope-average soil water content data are needed. Having determined that fill and spill is the mechanism that dictates hillslope runoff response for snowmelt over frozen ground, and given the underlying phenomenological similarities in fill and spill runoff generation processes at different partitioning surfaces (McDonnell, 2013; Ameli *et al.*, 2015), then we can also look to existing fill and spill modeling approaches, just as Ameli *et al.* (2015) used an overland flow model to predict hillslope-scale subsurface flow. Existing fill and spill-like approaches have the potential to greatly improve predictions of wetland recharge, flooding, and water availability, for the dominant runoff-producing event of the year on the northern Great Plains. Appels *et al.* (2011) and Chu *et al.* (2013) developed numerical models to

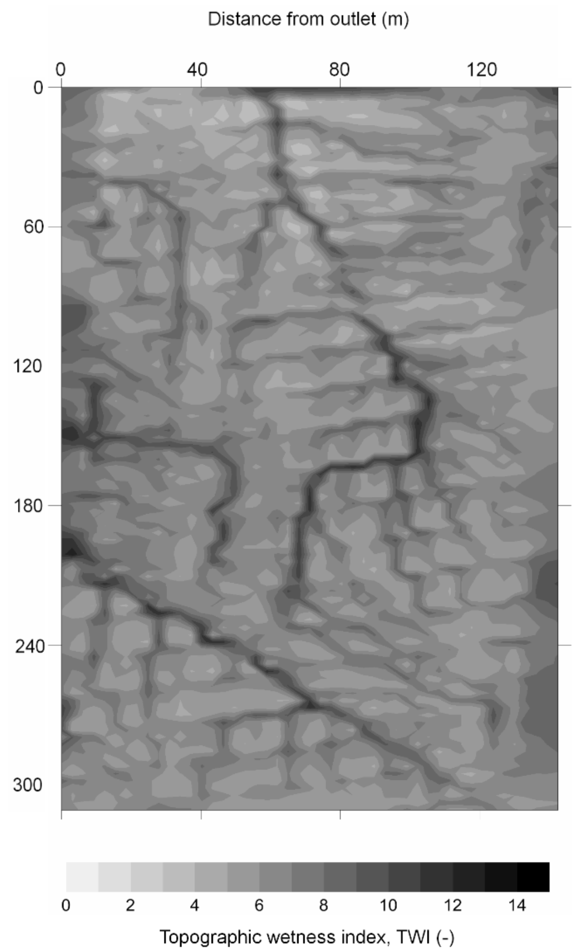

Figure 4.8 Spatial map of the topographic wetness index (TWI).

explore the effects of the spatial organization of meso- and micro-topographic features on flowpath convergence, connectivity, and runoff. Interestingly, these plot- and hillslope-scale ponding and redistribution models are numerically very similar (save for their treatments of infiltration) to a physically-based landscape-scale model devised by Shook *et al.* (2013) to simulate surface storage dynamics in prairie wetlands that have been shown to connect and disconnect via the fill and spill mechanism (Leibowitz and Vining, 2003; Shaw *et al.*, 2012; Leibowitz *et al.*, 2016). Such approaches might therefore be adopted for the modeling of hillslope runoff response for snowmelt over frozen ground.

4.7 Conclusions

We examined snowmelt-runoff processes for the 2014 snowmelt season at a 5 ha research hillslope site on the northern Great Plains. The fill and spill mechanism appears to explain the generation of snowmelt-runoff over frozen ground. Our main evidence for fill and spill is that: 1) the contrast between the slow infiltration rates of the uniformly frozen soil surface and the relatively fast rates of delivery of snowmelt water to the soil surface generated water beneath the snow at the soil surface that accumulated in surface depressions; 2) stable isotope analysis of water showed that runoff water was event snowmelt water with limited mixing with pre-event soil water; and 3) observations of ponded water and flowpaths matched our predictions of fill and spill activity from digital topographic analyses that combined flow accumulation and downslope indices. We observed nested filling at the micro- and meso-scale, followed by macro-scale spilling, where large patches of ponded water coalesced to drive a threshold-like increase in hillslope runoff. The identification of fill and spill as a mechanism to explain meltwater runoff from shallow, frozen hillslopes supports similar findings from peat-dominated permafrost sites in northern Canada where the frost table acts as an impeding layer, and has widespread implications for other areas of the northern Great Plains and similar low-angled, snowmelt-dominated, frozen regions.

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4.9 Author contributions

JJM conceived the idea to determine the controls on connectivity over frozen ground. AEC and JJM discussed the approach to do this. AEC installed the instrumentation, designed the sampling regime, carried out the field and laboratory work, and conducted the spatial mapping and data analyses. AEC wrote the first paper draft. JJM edited and commented on the manuscript and contributed to text in later iterations.

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CHAPTER 5

CONCLUSIONS AND FUTURE WORK

5.1 Conclusions

The basic research questions explored through my PhD have centred on achieving greater understanding of how runoff is generated over gently-sloping, seasonally-frozen hillslopes. The economic prosperity of the Canadian Prairies and northern Great Plains of North America is heavily dependent on water and its partitioning into different components of the landscape (*e.g.* soil water, streamflow, wetlands, and groundwater). In this region, snowmelt on frozen ground is the major runoff-producing event of the year. However, this process is unstable and still poorly understood. Changing climate impacts, multiple interacting controls, and nonlinear responses all challenge predictability. Therefore, there is a need to integrate advances in process and field-based understanding (*e.g.* connectivity and spatial patterns analysis) with the use of long-term datasets to test and quantify change under variable hydro-meteorology.

Prior to my PhD research, there were no long-term climate-runoff analyses at the hillslope scale on the Great Plains, so the effects of observed precipitation trends on hillslope-scale runoff and water availability were unknown. I addressed this (Chapter 2, Coles *et al.*, 2017, in review for *Journal of Hydrology*) by analyzing a 52-year hillslope-scale dataset from three 5 ha agricultural hillslopes to determine whether or not there have been any effects of recent (1962-2013) climatic changes on the hillslope-scale runoff regime. I found that hillslope-scale snowmelt-runoff and spring soil water amounts have indeed decreased in response to winter snowfall decreases. By comparison, interestingly, rainfall-runoff has shown no response to increases in rainfall or shifts to more multi-day rain events.

I hypothesize that this seasonal difference in the runoff response is due to differences in soil infiltrability and soil storage modulation between winter and summer. In the summer, the ground is unfrozen and the soil has a high infiltrability, thereby buffering the runoff response to rainfall.

Although the nature and total amount of rainfall has changed over these 52 years, the frequency of high-intensity rainfall has remained similar. Consequently, the change in rainfall regime has not yet been enough to trigger a related change in the runoff regime. Conversely, during the spring freshet, frozen ground limits infiltration which means that runoff responses via overland flow more closely mirror the trends in snowfall and snowmelt. These findings are counter to climate-runoff relationships observed at the catchment scale on the northern Great Plains (Dumanski *et al.*, 2015). This is likely a result of landscape alteration, most notably drainage, at that scale. This new hillslope-scale information is useful for future planning of water availability for dryland crop production on the northern Great Plains – where hillslope runoff trends are important for on-farm water supplies, and where declining runoff and declining spring soil water content can be related directly to economic costs for agriculture.

Having established the general trends and factors in climate-runoff responses over the 52-year period, my data mining research (Chapter 3, Coles *et al*., 2016, *Hydrology and Earth System Sciences Discussions*) sought to unravel the multiple interacting process controls on snowmelt-runoff, the nonlinearities and feedbacks between them, and their condition-dependent nature. This understanding is needed for model development, spatial extrapolation, and runoff classification schemes (Cammeraat, 2002; Uchida *et al.*, 2005; Barthold and Woods, 2015). This is not possible through the standard short-term experiments or single-season studies where nonlinearities and interactions between various process controls typically are not observable. My data mining research made use of the 52-year dataset and revealed the hierarchical importance of different runoff controls. The nonlinear relationship between total seasonal snowfall and total seasonal runoff was largely controlled by six factors (in descending order of importance): total snowfall, snow cover amount, fall soil surface water content, melt rate, melt season length, and fall soil profile water content. Together these worked to control the fraction of water that infiltrated frozen and thawing ground, and explained overall 70% of the runoff ratio variance over the 52-year record.

Chapter 3 showed that the hierarchy of controls was condition-dependent. When the soil in the previous fall had been dry, runoff ratios were not predictable based upon precipitation amounts or snow cover water equivalent. Further, while soil water content was the most important control on runoff ratios under conditions of high snow cover, it was a relatively unimportant control under conditions of low snow cover. This might be due to mid-winter ablation events driving a change in soil water content. Despite these events, the system generally showed significant memory due to system dormancy over winter. A commonly-used method of predicting infiltration into frozen soil (Granger *et al.*, 1984) explained only 14% of the variance in runoff ratio, which has implications for its inclusions in hydrological modelling. Finally, the hierarchy of controls could be used to guide cost-effective and useful field measurements.

Chapters 2 and 3 used long-term, lumped-hillslope data to show the importance of infiltration in dictating climate-runoff trends and seasonal runoff response. However, high resolution spatial and temporal data that illuminate within-event thresholds and patterns were crucial for driving process understanding. I therefore embarked on a field-based assessment of the factors controlling the patterns and mechanisms of runoff connectivity over frozen ground (Chapter 4, Coles and McDonnell, 2017, for submission to *Hydrological Processes*). I measured the spatial patterns of snow cover, snow water equivalent, soil water content, frozen ground, and topography for the 2014 melt season. I found that filling and spilling of micro- and meso-depressions across a 5 ha hillslope drives water delivery to the hillslope outlet. This was despite the low-angled, gently-sloping nature of the surface topography, as well as previous suggestions (*e.g.* Devito *et al.*, 2005) of the unimportance of topography on a similarly low-angled site with deep glaciated substrate.

The fill and spill mechanism observed here for driving runoff connectivity over frozen ground is fundamentally the same as the mechanism seen in many other environments (*e.g.* Darboux *et al.*, 2002; Spence and Woo, 2003; Tromp-van Meerveld and McDonnell, 2006; Wright *et al.*, 2009; Du *et al*., 2016; Leibowitz *et al.*, 2016). While the decision tree analysis in Chapter 3 showed that a lumped approach could indeed be fruitful for predicting the seasonal runoff response, this field-based spatial analysis showed that, in order to understand and predict sub-seasonal timescale (daily, hourly, or weekly) runoff responses, then distributed topographic data and distributed snowmelt data are needed. Fill and spill-like modeling approaches have the potential to revolutionize snowmelt modeling over frozen hillslopes and improve predictions of wetland recharge, flooding, and water availability, for the dominant runoff-producing event of the year on the northern Great Plains.

The results of this PhD research have advanced our understanding of runoff generation over seasonally-frozen ground. Through a coupled analysis of trends, hierarchies, and patterns, I have demonstrated the seasonality of climate-runoff relationships, the effects of interactions and feedbacks between controls on snowmelt-runoff response, and the processes behind hillslope-connectivity and emergent runoff behaviour.

5.2 Future work

My PhD findings have shown the importance of micro- and meso-topography for dictating runoff connectivity over seasonally-frozen prairie hillslopes. On the predominantly agriculturally-managed northern Great Plains, differences in agricultural and cultivation practices, such as tillage or seeding, are likely to induce a major shift in the importance of micro-topography as a first step in nested filling and spilling and hillslope connectivity in any given snowmelt season. For example, following seeding, the resultant micro-topography is often channelized, in the form of furrows and ridges. Yet following tillage, these micro-depressions become isolated from each other (Moreno *et al.*, 2008; Antoine *et al.*, 2009). Numerical modeling in the form of virtual experiments should address how subtle differences in the micro- or meso-topography might shift the balance of fill and spill and flowpath development on shallow, frozen hillslopes. This could be accomplished with distributed models such as HydroGeoSphere (Brunner and Simmons, 2011) or the Connectivity of Runoff Model (Reaney *et al.*, 2006), where different surface topographical realizations could be generated for the Swift Current (or fully virtual) hillslopes. Modeling could then feature and isolate the effects of different patterns of micro-topography, in conjunction with the patterns of state variables, on connectivity.

There is also considerable scope for lab-based physical models of the processes examined at the Swift Current hillslopes, using the MOST facility at the University of Saskatchewan (mostfacility.usask.ca). Trailer-sized instrumented hillslopes with climate control are able to easily observe and quantify input, output, and storage change and control boundary conditions. With these hillslopes, rainfall experiments under a rainfall simulator with unfrozen conditions could be used to examine the unexpected minor effect of changing rainfall that I observed in the climate-runoff analysis. These would be invaluable in being able to dial in, under different antecedent conditions, on how intense rainfall really needs to be to generate significant runoff. Further snowmelt experiments under imposed freeze-thaw cycles with a tailor-made refrigerated system could be used to explore the effects of variable freezing depth, antecedent soil moisture conditions, and frozen soil infiltration rates on meltwater partitioning, storage-discharge relations, and hillslope connectivity. Finally, these controlled experiments would be useful to test and further develop existing fill and spill algorithms for distributed hillslope modeling of snowmelt over frozen ground.

Finally, all three research chapters here have emphasized the importance of infiltration into frozen ground as the determining factor in hillslope-scale runoff responses. However, there is little known about the effects that future climate change will have on frozen ground depths, duration, and timing of thaw in relation to snowmelt. This is in part because it is not solely affected by air temperature, but also by snow cover and thermal properties of the soil (Ireson *et al*., 2013). Snow cover can have both a seasonal warming and cooling influence on the soil (Ireson *et al*., 2013). If the relative timing of snowpack melt and soil defrosting shifts (*i.e.* if soil defrosting occurs before snowpack melt) and the infiltration regime changes from frozen to thawed, this would have huge knock-on effects for hillslope water balance, water availability, and water resources. Further work is required to evaluate the potential for this and its implications.

5.3 References

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